Landscape Evolution and Holocene Climate Change in Mountain Areas of the Northern Highlands, Scotland

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Numerous Originals in Colour

DECLARATION

This thesis has been submitted in accordance with the Regulations for Higher Degrees by Research: A7/B7. I declare that this thesis has been composed by myself and that it embodies the results of my own research. Where proper I have acknowledged the nature and extent of work carried out in collaboration with others.

ABSTRACT

LANDSCAPE EVOLUTION AND HOLOCENE CLIMATE CHANGE IN MOUNTAIN AREAS OF THE NORTHERN HIGHLANDS, SCOTLAND

Holocene landscape evolution in the tectonically quiet mountain areas of the Northern Highlands of Scotland has been attributed largely to postglacial relaxation, which has left a legacy of stable, relict landscapes, disturbed only by intrinsic local response, and modified to an uncertain extent by human activity. A review of this model was prompted by improved understanding of a) the variability of the Holocene climate in mid latitudes, b) the responsiveness of some geological and geomorphological systems to low amplitude climate fluctuations, and c) a small number of field studies from the region, reporting mid and late Holocene slope mass movement unrelated to anthropogenic impact.

Sixteen catchments were explored using fieldwork and aerial photographic analysis. Slope activity since the end of the last glacial was investigated at five sites which contained evidence of long sequence, shallow slope failure, gully transport of slope debris, and debris fan formation. At two of these, stratigraphic sections, together with sediment and facies analysis, were combined with radiocarbon dating, in order to elucidate slope processes and construct a chronostratigraphy.

Results confirmed widespread Holocene lower slope re-organisation, with mid and late Holocene landscape rejuvenation occurring millennia after apparent adjustment to postglacial conditions at the two dated localities. Mass movement on slopes was found to have parallels in floodplain aggradation and incision. These transformations appear to have operated on several different time scales, and across a strong regional precipitation gradient. Since they are a function of the glacial inheritance of these landscapes, the potential for further transformations exists.

Mid and late Holocene events are only poorly accounted for by paraglacial relaxation. A more robust model of landscape evolution in this setting, incorporates climate change (specifically, precipitation shifts) - interacting with progressive weathering and vegetation cover - as a critical environmental variable. Although no justification was found for the use of dated slope mass movements as palaeoclimate proxies, changes in event frequency on a time scale of 10^3 years may contain a climatic signal.

ACKNOWLEDGEMENTS

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CHAPTER 1

NORTHERN SCOTLAND: THE CONTEXT OF HOLOCENE LANDSCAPE EVOLUTION

1.0 INTRODUCTION

'Little is known about Holocene landform development in Upland Britain,' (Harvey *et al* 1981).

'Our understanding of the timing of initiation and rates of development of Flandrian geomorphic features in the British upland landscape is still quite poor...' (Tipping & Halliday 1994).

'UK work on rapid mass movement as a source of palaeoclimatic evidence ....is largely based on lowland areas.' (Berrisford & Matthews 1997)

As the above quotations demonstrate, the sensitivity of British landscapes, especially that of slope systems, to Holocene climate variability, is not yet well understood. However, there is evidence from, for example, temperate North America (Arbogast & Johnstone 1993, Knox 1993) and continental Europe (e.g. Starkel 1995) that even modest climate change can create critical thresholds in fluvial systems. Comparable conclusions about slope systems in northern Europe are discussed in Chapter 2. Northern Scotland is a climatically sensitive area (section 1.2 below), which constitutes a large proportion of upland Britain. It therefore provides a useful context for continuing research. Figure 1.1 shows the location of Northern Scotland, north of the Highland Boundary Fault. For the purposes of this thesis, it includes the Northern Highlands and Grampian Highlands as defined by the British Geological Survey (British Regional Geology, Regions 2 and 4).

To set the scene for a critical review of work to date on Holocene landscape evolution in upland areas of Northern Scotland, the geology, topography and climate of the region are briefly outlined below. This is followed by an account of the glacial and periglacial inheritances which constitute the context for Holocene processes and events.

1.1   GEOLOGY AND TOPOGRAPHY

Figure 1.2 is a simplified geological map of Northern Scotland. A comprehensive account of the geology, to which reference is made here, can be found in Johnstone & Mykura (1989) and Stephenson & Gould (1995). The most widespread rock type is

Precambrian crystalline basement. Younger sediments of Cambro-Ordovician, Devonian and Jurassic ages fringe the west and east coasts, while Caledonian granites and allied rock suites form large, partly exhumed batholiths such as the Cairngorm massif. Just off the west coast, the Isle of Skye is one of a series of Tertiary volcanic centres associated with aborted rifting of the North Atlantic ocean.

Northern Scotland is structurally complex. It is cut by the largest structural feature in the British Isles - the Great Glen Fault Zone (Fig. 1.2). Associated sets of NE-SW trending major faults and splay faults determine the main grain of the landscape. Other important sets of faults run NW-SE, E-W and N-S. Repeated movement on these faults and their associated splay faults has created numerous planes of weakness excavated into deep valleys by successive ice sheets and the locus for repeated seismicity. A second major feature, the Moine Thrust zone, roughly parallels the west coast from the Pentland Firth to Skye. Within it, slices of crust were carried up to 45km westwards over older basement towards the end of the Caledonian orogeny.

In this old, tectonically stable, mountain area the highest peaks and plateaux stand at little over 1000m. The combined effects of structure, lithology and glacial erosion have been to create a north-south watershed with steep, short, deeply dissected slopes rising directly from the west coast, and gentler longer slopes to the east. No locality is more than about 40km from the coast which, in the west, is highly indented with over-deepened, fjord-like sea lochs.

1.2   PRESENT-DAY CLIMATE

Northern Scotland lies at an atmospheric and ocean current crossroads with a climate characterised by short-term variability and extreme oceanicity. Its latitude places it north of Moscow and of most of Labrador, and mountain ranges and massifs experience severe conditions normally associated with higher latitudes and altitudes. Dominant weather patterns vary from wet, mild weather associated with south-westerlies from the Atlantic, and the warming influence of the Northern extension of the Gulf Stream, to cool dry air masses from the continental land mass to the east.

Much of the following information about the contemporary climate is derived from *The Climate of Scotland, The Meteorological Office, HMSO, London, 1989.* Measurements are largely from the period 1950 - 1980.

1.2.1 Precipitation

The distance from the west coast of Northern Scotland to the Cairngorm massif is only about 130 kilometres (Fig.1.1). But prevailing south-westerly winds from the Atlantic maintain high rainfall in the west and much drier conditions in the east (Fig. 1.3). Precipitation also increases with altitude. For low level sites, the annual range is from c.2000mm in the west to about 600mm in the east. Above about 500m, these figures are at least doubled. Close to the watershed annual precipitation is in excess of 1600mm at lower altitudes. Rainfall in the Cairngorm massif is less than 1000mm at an altitude of 300m - around 50% of what it is in the west at sea level (Birks 1996). However these are averages of figures which vary strongly from year to year and decade to decade. The variability is illustrated by the fact that in a single month in 1938 Kinlochquoich in the west had a rainfall of 1270mm while Braemar in the eastern Cairngorms had only 5mm (Lamb 1982). Weather stations are infrequent, and were not available at most of the sites described in Chapters 4, 5 and 6. Rainfall figures are therefore estimated from published isohyetes which may be subject to annual variation.

1.2.2 Snow and wind

Snow cover like rainfall is very variable from year to year and decade to decade. For instance, at Glasgow airport in the winter of 1946/7 there were 47 days with lying snow but in 1960/61 there were none. Mountain snowlie reflects the high degree of windiness of Scottish winters, with scoured ridges and summits and deeper accumulation in more sheltered areas. Natural hollows in the highest areas may contain semi-permanent drifted snow. On the highest summits (>1000m), snow cover lasts on average for six months of the year, but annual snowfall oscillates strongly around the average.

1.2.3 Temperature

Winter temperatures are strongly influenced by the surface temperature of the surrounding seas. Since the North Atlantic Drift ensures that the Atlantic is warmer than the North Sea, the major temperature gradient, like the precipitation gradient, is from west to east, not from north to south: thus it gets both cooler and drier in winter towards the east. Only around the Cairngorm massif are mean January minimum temperatures slightly below freezing. Superimposed on the influence of sea temperatures are the effects of distance from the coast and elevation. Mean January temperatures are about 1.5° C lower in the east at sea level, and about 2°C lower on the higher ground. Unlike

winter temperatures, average summer temperatures *increase* by 1 - 1.5°C from west to east away from the coast. As with precipitation, this variation is similar to estimates of the range of Holocene palaeotemperatures (Birks H. J. B. 1996). Such estimates take no account of temperature and precipitation gradients within the region.

1.2.4 Discussion: some implications for interpreting palaeoclimate data

The western part of northern Scotland has a moister, more equable climate, while the more continental Cairngorm massif and eastern zone is relatively dry and has a wider winter - summer temperature range. However, large annual and decadal variations in precipitation are the norm as a result of geographical setting. Inherent variability poses problems for palaeoclimate reconstruction when superimposed on the effects of altitudinal and latitudinal variation, short-term (annual and decadal) and long term climate variability. Low magnitude oscillations may be poorly captured by terrestrial data. Interpretation of palaeoprecipitation records is further complicated by suggestions that the precipitation gradient itself has not been constant. When rain-bearing south-westerly winds dominate weather patterns, the west coast mountains rising sharply from sea level create a rain shadow zone to the east, producing the pattern seen in Figure 1.3. But when meridional circulation - i.e. an easterly air stream - is dominant, as has been inferred during 'the Little Ice Age', the precipitation gradient may have been less strongly maintained (Whittington 1985, Birks H. J. B. 1996, Rumsby & Macklin 1996).

1.3 THE LATE-DEVENSIAN/HOLOCENE TRANSITION

Events immediately preceding the Holocene are described below since they constitute the starting point for subsequent landscape evolution.

The Devensian Lateglacial in Northern Scotland is believed to span the period from about 13,500 - 10,500 radiocarbon years BP (Lowe & Walker 1984, Gordon & Sutherland 1993, Ruddiman & MacIntyre 1981, Sissons 1977, 1980). The retreat of polar waters from the seas round the north of Scotland at about 13,500BP, and the rise of temperatures to near present day values, appears to have taken place on a timescale less than the resolution of radiocarbon dating - perhaps no more than a few decades (Atkinson *et al* 1987, Bjørck *et al* 1998, Dansgaard *et al* 1989, Peacock & Harkness 1990, Rahmsdorf 1994, Taylor *et al* 1993).

Radiocarbon dated pollen stratigraphy has been widely used to identify the switch from Dimlington Stadial to Lateglacial Interstadial conditions in lake and kettlehole

sediments (Birks H. H. & Matthews 1978, Kirk & Godwin 1963, MacPherson 1980, Pennington *et al* 1972, Pennington 1977, Vasari 1977). At about 13,500 BP mineral sediments began to be replaced by sediment with a high percentage of organic carbon, accompanied by an influx of new plant taxa. By 13,000 BP much of mainland northern Scotland must have been free of ice, since pioneer, open habitat plant communities had replaced minerogenic sediment, at least on lower ground (Huntley 1994, Pennington 1977).

Arctic conditions, accompanied by partial ice cover centred on the west and south-west Highlands, briefly returned during the Younger Dryas Stadial at around 11,000 radiocarbon years BP (Fig. 1.4). Subsidiary ice-fields and glaciers developed beyond the main ice mass (Ballantyne *et al* 1991, Bennett 1994a, Bennett & Boulton 1993, Sissons 1977). This episode, described in Scotland as the Loch Lomond Readvance, lasted for around 1000 years (Gordon & Sutherland 1993, p36, Sutherland 1980). In the northwest Highlands, break-up of continuous plant communities and frost disturbance of soils was dated between 11,000 and 10,400 radiocarbon years BP (Pennington *et al* 1972, Pennington 1977). Beyond the margins of the Younger Dryas ice sheet periglacial activity, destruction of soil profiles, and accelerated erosion predominated (Sissons 1977); ice wedge casts (some possibly dating from the Dimlington Stade) fringed the ice cover (Rose *et al* 1984, Fig. 18.2); minerogenic soils suddenly replaced organic sequences in lake basin and terrestrial sequences (MacPherson 1980, Pennington *et al* 1972, Vasari 1977); while reworked pollen, interpreted as a signal of renewed periglacial soil disturbance and erosion, appeared in catchment sediment sinks (Huntley 1996). In summary, there is evidence from numerous sites across northern Scotland for a severe, short-lived, return of Arctic conditions.

The precise dates of ice growth and retreat are uncertain, due to a combination of substantial errors on radiocarbon dates and the sometimes poor comparability of dates obtained from different sites (Sutherland 1980, Pilcher 1991). Radiocarbon dating at Stadial/Interstadial boundaries is probably subject to additional error derived from the release of large amounts of old carbon from formerly stratified oceans and areas of permafrost (Becker 1991, Lowe 1991, Pilcher 1991, Sutherland 1980). With these *caveats*, the end of the Younger Dryas Stadial in northern Scotland is estimated as having occurred close to 11,000 cal BP (Gordon & Sutherland 1993, Lowe & Walker 1992).

A number of studies suggest complex patterns of deglaciation. Pollen diagrams and radiocarbon dated basal sediments in enclosed basins within the readvance limits are consistent with ice having persisted in some areas as late as 9800 BP. Elsewhere, it may have disappeared as early as 10,600 BP (Gordon & Sutherland 1993, p37). In the eastern Highlands, radiocarbon-dated plant associations are those characteristic of extensive snowlie, while rapid organic deposition was occurring elsewhere (Huntley 1994). In the northwest Highlands, deposition of purely mineral sediment in lakes ceased by 10,400. Between then and 10,000 BP, pollen data indicate a rapid transition to closed plant communities, while sediment in the basins changed rapidly to muds derived from organic soils (Pennington 1977).

On the Cairngorm massif and adjacent areas, vegetation changes associated with the end of the Younger Dryas have been dated in several localities (Birks H. H. 1975, Birks H. H. & Matthews 1988, Huntley 1981, 1994, MacPherson 1980). The data are reviewed in Gordon & Sutherland (1993, pp 257). Pollen and sediment sequences show changing proportions of snow beds, frost disturbance, and lake water stratification during the Younger Dryas, giving way sharply to Holocene organic muds as the temperature rose.

The Monadliath Plateau, adjacent to the Cairngorm ice centre, was ice-free during the Younger Dryas (Sissons 1974, 1977). A similar pattern was found in the northwest Highlands where deglaciation was probably complete before 13,000BP, and birch woodland was established by about 9500BP (Pennington *et al* 1972, Vasari 1977).

Geomorphic evidence confirms complex patterns of deglaciation during and after the Younger Dryas Stadial. Using present-day ice-fields as analogues, Bennett & Boulton (1993) mapped patterns of retreat moraines and valley connectivity. They concluded that deep incision and isolated remnant ice masses in the northern Highlands created a mosaic of temperatures and environments. On the basis of inferred glacier firn line altitudes, climatic constraints on relict periglacial phenomena, plant community reconstruction, and models of atmospheric circulation, it has been suggested that deglaciation was substantial in the rainshadow area of the Cairngorm Mountains well before temperatures rose to interglacial levels (Ballantyne 1996, Birks & Matthews 1978, Lowe & Walker 1992, Sissons 1979). Ice retreat in a severe glacial climate, combined with restriction of the main ice mass to the west of the watershed (Fig.1.4), must have led to widespread, severe periglaciation.

Compared to July mean temperatures of only 5-6°C during the Stadial, summer temperatures in the early Holocene may have been at least as warm as those of today (Kutzbach & Gallimore 1988, Kutzbach *et al* 1993, Pilcher 1991, Sissons 1977). Radiocarbon dating and pollen analysis of organic sediments behind moraines in high coires (cirques) in the Cairngorm massif has confirmed that they have remained unglaciated throughout the Holocene (Rapson 1985).

1.4   EARLY HOLOCENE SLOPE ADJUSTMENT

Slope adjustment to interglacial conditions was rapidly achieved at the start of the Lateglacial Interstadial, and again after the Younger Dryas Stadial, through faulting, rock fall and debris flows (Ballantyne 1991, 1995a, Benn 1992, Bennett & Boulton 1993, Bennett 1994a, Fenton 1991, Holmes 1984, Peacock 1986, Sissons & Cornish 1982, Sugden 1971).

On some lithologies, periglacial frost weathering generated an abundance of unstable sediment on nunataks as the main Late Devensian ice sheet retreated, and in extensive non-glaciated areas during the Younger Dryas (e.g. Ballantyne & Harris 1994, Blikra & Longva 1995). Drift slopes were extensively modified by solifluction (Gordon & Sutherland 1993, pp 38). In conjunction with intense rock fall (Holmes 1984), the legacy of these processes was large volumes of unconsolidated debris and over-steepened slopes (Ballantyne 1986, 1991). Accumulations of frost shattered debris formed debris cones and talus slopes.

Tripartite slopes of the type rapidly formed after deglaciation in Western Norway, and suggested as an analogue for early Holocene slopes in Scotland (Ballantyne & Benn 1996, Fig. 52.6c), were found to be common in the catchments surveyed for this thesis (Chapter 4). They consist of an upper section of steep exposed bedrock, mid slope gully systems, and convex lower slopes occupied by debris flows coalesced into cones/fans. On valley floors, large changes in the magnitude and patterns of surface and subsurface water flow occurred as permafrost and meltwaters receded (Boulton *et al* 1977, Maizels & Aitken 1991, Richards *et al* 1996). Many of the lower slope debris accumulations in Northern Scotland have been described as 'relict', that is, dating from the very early Holocene. Despite this, there is increasing evidence that slope and valley floor landforms continued to evolve, even after several millennia (Chapter 2).

1.5 THE LEGACY OF GLACIAL EROSION AND DEPOSITION

Slope mass movement is the result of the interaction of surface and subsurface water with bedrock and sediment. This section examines the inherited conditions which influence Holocene sediment transport and water transmission in mountain areas.

1.5.1 Hydrogeology

Water transmission through crystalline bedrock is a function of the frequency and connectedness of discontinuities such as joint and cleavage planes, fractures, faults and lithological and/or rock-drift boundaries, and is strongly channelled (e.g. Lever & Woodwark 1989, McEwen & Lintern 1980, Michie & Hooper 1992). These conditions in turn determine infiltration, residence time and the location of spring lines and rock-cut gullies. The position and flow volumes of springs may change with time, and may be related to slope failure (Fenton 1991).

The saturated hydraulic conductivity of hillslope sediments and bedrock is critical to the development of slopes and landforms (Brunsden 1979, Freeze 1987, Kirkby 1978, 1987). When large storms occur, there is rapid ground saturation and return flow to the surface (Dunne 1978). Gullies channel run-off and provide preferential routes to the surface for shallow groundwater percolating down, and across, slopes. They are the focus for both surface and subsurface water transmission, not only because they concentrate downslope subsurface flow, but because their concave profiles mean the water table is closer to the surface than under a straight slope of similar gradient. High porerewater pressures which trigger rapid mass movement in sediments and unstable bedrock are therefore concentrated in and below gully floors, closely linking water and sediment cycling.

The transmission of water through bedrock influences rates of weathering and denudation as well as sediment transport and accumulation. Weathering occurs through rock:water interactions, both physical and chemical, particularly hydrolysis in Northern Scotland (Bain *et al* 1993). It reduces the strength of rock masses by opening joints and fractures, thus reducing friction. Rates of weathering are therefore related to rates and volumes of groundwater flow. This produces a positive feedback. Since the surface area of a unit of rock doubles if it is split in four, and increases exponentially with further subdivision, the surface area of rock interacting with water increases rapidly with the rate of weathering. Fault and frost shattering in steeply foliated and closely jointed rock

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in which vertical discontinuities enhance infiltration, therefore makes rock masses significantly more susceptible to rapid weathering, whatever the lithology.

Infiltration of groundwater through fractured rock is up to several orders of magnitude slower than through unconsolidated, superficial deposits (Mather 1996, Robins 1990). The length of pathways to stream beds is also one or more orders of magnitude longer through bedrock than through till or glacifluvial deposits a few metres thick (e.g. Chapman & MacKinley 1987, McEwen and Lintern 1980, Michie & Hooper 1992). This means that infiltrating water in a recharge zone might reach a river after a few hours from till-blanketed lower slopes, but only after hundreds, or thousands, of hours from rocky upper slopes.

1.5.2   The legacy of glacial erosion

Holocene cycling of sediment and water through catchments is conditioned by the extent of valley incision, the presence or absence of ice-scoured surfaces related to warm or cold-based ice cover, the formation of subglacial gorges and rock basins, and the excavation of gullies in rock structures.

Warm-based ice reaches the pressure-melting point at its contact with the substrate, generating copious amounts of sediment-laden water at high hydrostatic pressures (Boulton *et al* 1977, Richards *et al* 1996). In bedrock with frequent discontinuities, high pressure subglacial water flows charged with abrasive sediment takes the path of least resistance, often along these discontinuities, creating linear excavations which may persist as numerous gullies after ice retreat. In homogeneous rock with widely spaced structural discontinuities, fewer such linear excavations will form. After ice retreat this exerts a continuing influence on surface and subsurface water transport - hence sediment transport - on slopes.

The extent of ice-scoured landscapes with frequent ice-rounded outcrops is systematically related to the character of the Late Devensian ice sheet (Boulton *et al* 1977, Gordon 1979). The ice divide in northern Scotland was close to, or just east of, the present north-south watershed and precipitation divide (Figs 1.2, 1.3). The elevation of western periglacial trim lines (up to 950m) on nunataks peripheral to the main centres of ice accumulation, and contrasting glacial modification west and east of the divide, indicate warm based, fast-moving ice in the west with deep dissection. To the east of the ice divide, cold-based ice was common (Ballantyne *et al* 1998). That contrast is well

displayed at, for example, the head of Glen Shiel (NH 0012). To the east are areas of ice-scoured bedrock. To the west are thick drift, deep incision and frost shattered, gullied peaks (see Chapter 4). Slow-moving, cold-based ice is much less erosive (Boulton *et al* 1977, Gordon 1979). Extensive areas of thick grus, formed during Tertiary weathering, have survived east of the Great Glen (Fig. 1.2) (Stephenson & Gould 1995).

Glacially overdeepened troughs are characteristic of many valleys excavated beneath warm-based ice in the northern Highlands. As a result, elongate lochs commonly occupy valley floors. Sub-glacial rock gorges may act as built-in powerhouses for generating repeated peaks in the transport of sediment and water by reducing frictional energy loss and increasing flow rates (Baker 1988, Nanson 1986). Rock gorges also isolate river flow from the adjacent landscape and uncouple river and slope processes in their vicinity.

1.5.3 The legacy of glacial deposition

Glacial deposition in the northern Highlands is spatially and sedimentologically heterogeneous (Boulton *et al* 1977, Gordon 1979). Till composition varies from stiff stony clay to sand-dominated facies. Geological Survey 1:50,000 scale Drift maps and Soil Survey 1:125,000 land use capability maps (Futty & Towers 1982) provide general information.

Where the ratio of bare rock to sediment cover is high, run-off is enhanced at the expense of infiltration (Eybergen & Imeson 1989, Freeze 1987). Where upper slopes consist of bare rock, run-off, channelled in gullies, will predominate. The relative proportions, and the spatial distribution, of drift and bedrock in a landscape are therefore probably a significant control on mass wasting and valley floor reworking.

Freely draining sandy tills are widely distributed throughout the Highlands and give rise to leached humus-iron podzols in Northern Scotland (Futty & Towers 1982). Elsewhere, clay-rich and silt-rich tills impede surface drainage, and at the same time locally inhibiting backflow from saturated bedrock. The morphology, as well as the sedimentology, of ice contact and glaciofluvial deposits may determine pathways of landform development. Moundy ice-contact deposits blanketing a valley floor and lower slopes result in complex groundwater flow and energy dispersal. They may also

constrain lateral stream movement and inhibit slope undercutting while buttressing rapid mass movements from upslope.

1.5.4   The legacy of periglaciation

Periglaciation has produced contrasts in weathering above and below trim lines, described by McCarroll *et al* (1995). Measurements on ice-scoured gneiss and on sandstone from NW Scotland indicate that joint dilatation above the trim line, which marks the former ice surface, is double that below, irrespective of altitude, with joint depths of 60-70mm as opposed to 30-40mm beneath the ice surface where the elastic modulus of rock is significantly higher (Ballantyne *et al* 1998). Consequently, Holocene weathering of slope materials operates on juxtaposed environments with contrasting thresholds for infiltration, runoff and debris production. Slopes with sediment sources above the trim line may therefore be more sensitive to erosion than those where ice protected the bedrock from severe periglaciation.

While periglacial processes have continued to modify slopes above 550m at times during the Holocene, their magnitude is small in comparison to processes operating during the Younger Dryas Stadial (Ballantyne & Harris 1994, Mottershead 1978, White & Mottershead 1972). Accelerated solifluction since deglaciation, as determined by radiocarbon dating of organic matter beneath advancing lobes, may have occurred at about 4000BP, again at about 2500BP, and rather more certainly in the last few centuries (Ballantyne 1993).

The extent to which widespread periglaciation during the Younger Dryas affected subsequent slope mass movement in Northern Scotland has been little quantified. Two investigations have touched on the issue. In the Grampian Highlands, Brazier (1987, p307) identified alluvial fans as being larger in periglaciated zones, and debris cones as equally developed within and beyond the ice limits. She concluded that landform adjustment to non-glacial conditions may be slightly less well advanced outside the Younger Dryas ice limits than within them. An alternative hypothesis is that the difference can be accounted for by intense periglaciation beyond the Younger Dryas ice limits, providing larger volumes of transportable material and slopes more susceptible to Holocene processes. In a second study of the Northwest Highlands, contrasts were noted between relatively immature talus slopes within the ice limits, as measured by weathering of slope materials, and those which had been ice-free, and therefore intensely periglaciated, during the Younger Dryas Stadial (Ballantyne 1995b). Slopes

within and outside Younger Dryas ice cover may therefore show a variation in their subsequent evolution.

1.5.5 Paraglacial Processes

The term *paraglaciation* was first applied to the uplands of British Colombia (Ryder 1971) to mean non-glacial processes that are directly conditioned by glaciation. In the Highlands it was applied to the development of debris cones and alluvial fans (Brazier 1987, Ballantyne & Brazier 1989, Ballantyne 1991).

The main source of mobile, landscape-modifying material in the paraglacial model is a finite supply of glacigenic sediment which is redistributed until equilibrium is reached with Holocene system thresholds, or alternatively, until the sediment source is exhausted. Initial Holocene system adjustment in this model is rapid (Ballantyne & Benn 1996), and subsequent development continues to be dominated by postglacial relaxation (Church & Slaymaker 1989) rather than by Holocene rejuvenation. In the South-west Scottish Highlands, sediment exhaustion was inferred as the cause of cessation of fan aggradation before 4000BP, and proposed as a model to explain the formation of other coarse, vegetated fans in the region (Brazier *et al* 1988).

The paraglacial model accounts for features such as vegetated (i.e. inactive) cones and fans as the products of progressive exhaustion of paraglacial sediment from restricted source areas, which, in the absence of further major external system inputs such as human activity, remain stable (Ballantyne 1991, Brazier *et al* 1988). Ballantyne (1993) considered that paraglacial debris flow processes had occurred intensively in the early Holocene and been renewed within the last three centuries, with recent intensification possibly triggered by extreme storms during the Little Ice Age or by overgrazing of vegetation. Holmes (1984) worked on rock slope failures. He concluded that the majority had occurred between 10,000 and 5000 BP due to delayed action response to glacial or deglacial processes which left many slopes in a condition of critical stability.

'Sediment exhaustion' implies a source of debris which is renewed, if at all, at a slower rate than it can be produced. However, the lack of precise information about rates of weathering and mass transport in the uplands of the British Isles hinders confirmation of the scale and impact of any such effect. Long-term weathering rates have been calculated for some upland catchments in Scotland, but these rates are averages over 10,000 years and there is no indication as to how they have changed with time (Bain *et*

al 1993). However, estimates have suggested that rates of mass transport on one Scottish mountain (An Teallach, Wester Ross) are of the same high order of magnitude as those calculated for the more climatically extreme conditions in Northern Sweden (Ballantyne 1991).

1.5.6 Crustal rebound and seismicity

The Northern Highlands consist of tectonically stable, old mountain landscapes formed during the Caledonian Orogeny and underlain by Archaean basement. Faulting and isostatic uplift from depression beneath the Devensian and Younger Dryas ice sheets appears to have been largely accomplished in the period between the end of the Dimlington Stadial and the very early Holocene (Fenton 1991, Firth *et al* 1993, Holmes 1984, Peacock & Cornish 1989, Ringrose 1989, Sissons and Cornish 1982, Stewart *et al* 1998).

There is only a handful of localities where neotectonics have been shown to shape the Holocene landscape of the Highlands through surface faulting and through rebound away from coastlines (Davenport *et al* 1987, Fenton 1991, Firth *et al* 1993). This contrasts with evidence from Finland and Sweden where there is substantial evidence of fault scarps, slope destabilisation and stream diversion as a result of isostatic rebound which continues to be as high as 1-11mm/yr^{-1} (Pan & Sjøberg 1999, RWMAC 1988).

The younger the landform, the less probable is a seismic trigger, since there has been an exponential decrease in rates of isostatic adjustment and associated seismicity (Fenton 1991, Stewart *et al* 1998). The pattern of isostatic recovery from glaciation in Northern Scotland has been complex, because the effects of the Younger Dryas glaciation were superimposed after partial recovery from the retreat of the Late Devensian ice sheet, and landscape response has included both uplift and surface faulting (Fenton 1991, Firth 1986, Firth *et al* 1993, Ringrose 1989, Shennan 1995, Sissons & Cornish 1982). Among the potential consequences of isostatic rebound are lake level changes and associated effects on marginal slopes, reorganisation of groundwater flow, stream diversion, and bedrock shattering in, and adjacent to, fracture zones.

Despite the probable minor direct impact of faulting, the way in which seismic events may have conditioned the landscape, and potential implications for linear feedbacks in

Holocene landscape system evolution, deserve further consideration. Faulted valleys, often with splay faults and sub-parallel shatter zones, are extremely common in the

Highlands (Johnstone & Mykura 1989). Faults commonly exploit pre-existing planes of weakness, (Fenton 1991, Johnstone & Mykura 1989, Mörner 1981, Owen *et al* 1993) and shatter zones are a potential source of debris. In addition, crush zones not related to mappable faults are common enough to have been encountered in several hydroelectric scheme tunnels (Johnstone & Mykura 1989, p 174). Ground deformation also commonly changes groundwater flow paths and spring lines (Fenton 1991). Lateglacial and early Holocene ground shaking might therefore have left a legacy of a range of conditions capable of producing linear landscape change even if more obvious features such as diverted streams and fault scarps are absent. The ubiquity of fault zones and associated sets of splay faults in northern Scotland means these effects are potentially widespread.

1.5.7 Local factors: intrinsic system response

The environmental inheritance of landscapes, sometimes called 'landscape memory', ensures inbuilt system perturbations on a variety of concurrent spatial and time scales, even in the absence of externally generated energy inputs (Brunsden 1996, Dorn 1996, Phillips 1999, Thomas & Thorp 1995). Perturbations may take the form of both negative feedbacks (simple readjustments) and positive feedbacks leading to linear (irreversible) change (White *et al* 1994). Important local controls on lower slope landforms in upland areas of the UK have been suggested as catchment shape and size (Brazier 1987, Wells & Harvey 1987) and lithology (Ballantyne 1991). However, the close relationship between sediment and water cycling, suggests that in the crystalline rock characteristic of the Northern Highlands, the nature of the discontinuities may be at least as important as mineralogy in determining weathering rates and patterns.

1.6 HOLOCENE SLOPE PROCESSES AND MORPHOLOGY

Toppling failures and deep-seated landslides (apart from a single, possibly pre-Holocene failure in upper Gleann Lichd, Chapter 6) and were not identified during field work for this thesis, and the following discussion is limited to processes and landforms associated with shallow rapid mass movement on slopes.

1.6.1 Debris flows

Debris flows are flows of saturated, heterogeneous debris (e.g. Berrisford & Matthews 1997, Blikra & Nesje 1998, Brunsden 1979, Carson & Kirkby 1972, Dunne 1978, Iverson 1997, Kirkby 1987, Rodine & Johnson 1976). In small steep valleys, deposits

may be 2-5m thick (Bovis 1993). Rapid mass movement of slope debris has been described as sediment limited, that is, constrained by either the supply of transportable material (due to weathering rates or availability of paraglacial sediment), or as transport limited, that is, determined by its setting, which may be a steep rock face or gentle slope (Kirkby 1987). Processes of debris movement have been widely described by, for instance, Blikra & Nesje (1998), Brunsden (1979), Carson & Kirkby (1972), Dunne (1978), Iverson (1997), Kirkby (1987), Reneau *et al* (1986), although little research has been carried out on debris flow processes in Britain (Ballantyne & Harris 1994). Separate fluidised phases, and surge fronts of debris with low pore pressures, have been recognised experimentally and in the field. It is the latter which form levées.

Debris flows may be triggered when exceptional precipitation or snowmelt infiltrate at a rate which exceeds the hydraulic conductivity of sediments and soils (Berrisford & Matthews 1997, Freeze 1987, Kirkby 1978). Threshold conditions are generated through a combination of intensity and duration (Caine 1980). Failure occurs when high pore pressures act to reduce intergranular friction below thresholds of stability for a particular slope-sediment system. These conditions may be achieved not only during exceptional meteorological events, but when either persistent moderate precipitation or successive moderate storms produce a high water table (Kirkby 1978). If intense storms which create threshold conditions are frequent, debris flow intensity will be as frequent as the supply of sediment allows (Reneau *et al* 1986). These authors also proposed that increases in storm frequency and magnitude may mean that debris flow frequency is further increased due to an increase in flow from bedrock.

Recent observations demonstrate the sensitivity of unconsolidated sediment on some slopes in the British uplands to intense rainfall (e.g. Ballantyne 1991, Brooks & Richards 1994, Harvey 1986), although they equally show that intense storms do not always provoke debris flows in susceptible areas (Ballantyne 1991, Innes 1983a). Debris flows may also be triggered by precipitation after progressive weathering and soil development (Berrisford & Matthews 1997, Brooks *et al* 1997, Hinchcliffe *et al* 1998).

Grain collisions in moving, saturated debris, are minimised when pore fluid, composed of water with variable amounts of suspended fines, becomes highly viscous and nearly incompressible, causing the debris to flow like a fluid. On coming to rest, the debris resumes its initial rigid nature. Fluidised low pore pressure phases and surge fronts of

debris, which form levées, have been reproduced experimentally (Iverson 1997). Reverse grading occurs when large clasts in low pressure surge fronts 'float' on the surface of the flow, and when rapidly travelling surge fronts over-ride denser material within the body of the flow (Iverson 1997, Rodine & Johnson 1976).

Although the occurrence of debris flows may be closely related to hydrological conditions, in non-glaciated areas of northern Europe they may, in principle, be difficult to relate unambiguously to past climate. Precipitation may vary on annual, decadal, centennial and millennial timescales (Barber *et al* 2000, Rumsby & Macklin 1996). Consequently, where high resolution climatic records are lacking, as in Scotland, their combined effects are problematic to unravel in the stratigraphic record. However, where lengthy periods of slope stability alternate with periods of intensified activity, on a spatial scale which overprints local variables, an external trigger such as climate may be the most plausible explanation (Berrisford & Matthews 1997, Blikra & Nesje 1997, Brunsden & Ibsen 1997). On this large scale, there is scope for distinguishing anthropogenic from climatic signals as long as sites affected by processes unlikely to have been subject to human influence are distinguished, or where human impact itself is patchy and variable but the slope response is not.

1.6.2 Debris cones

Debris cones are fan-shaped accumulations formed by repeated gravity flows at the foot of steep gullies or tributary valleys (e.g. Ballantyne 1991). In the Scottish Highlands small cones have been described as forming from coalescing debris flows on open hillslopes and in gullies (e.g. Brazier & Ballantyne 1989, Innes 1983b).

Sediment sorting in cones is limited, and the proximal-medial-distal division of alluvial fan facies is absent. Reverse grading may be present due to flow mechanics, which allow coarse clasts to float on the surface of the fluidised core of the flow (Iverson 1997). Although similar to alluvial fans in planform, debris cones are normally steeper, with average gradients in northern Scotland typically in the range 12-25°, and smaller source areas relative to fans (Brazier 1987). In the Scottish Highlands, in the absence of contrary evidence, they have generally been considered to be relict features dating from deglaciation or its aftermath (Ballantyne & Harris 1994, Brazier 1987). However, late Holocene cone aggradation in the western Cairngorms has also been described as due to Holocene slope failure triggered by slope undercutting by a river channel (Brazier & Ballantyne 1989), and to late Holocene slope failure (Ballantyne & Whittington 1999).

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During fieldwork for this project, debris cones with up to three abandoned high terraces were identified (Strathfarrar NH 259372; Gleann Fhiodhaig NH 203509; Glen Cannich NH 230313; Chapter 4). This indicates repeated cone extension and surface angle lowering at discrete, and so far undated, intervals.

1.6.3 Alluvial fans

Alluvial fans are fan-shaped accumulations of predominantly fluvially deposited sediment with gradients of, typically, 2-10° in northern Scotland (Brazier 1987). Fans fine from proximal to medial to distal zones (Church 1980, Duff 1993, Rachocki & Church 1990). Hypotheses about fan formation in different climatic and tectonic settings have been summarised by, for instance, Brunsden (1996), Bull (1991), Lecce (1990), Rachocki (1990), Reading (1978), and Wolman & Garson (1978). There is a consensus that fan development can be influenced by climatic and tectonic effects, but the problem has been to identify precisely which processes are involved. Fan dynamics have been approached from three main angles: detailed investigations of deposition and processes and their relative importance in different environments; the effects of catastrophic events; analysis of fan development through sedimentology and dating. Fan incision has variously been associated with dry periods when vegetation dieback facilitates erosion (Rachocki & Church 1990); with anthropogenic vegetation clearance (e.g. Brazier 1987); and with sediment exhaustion (Church 1975, Wasson 1977), depending on the climatic regime and geomorphological setting of investigations. An alternative approach is to describe the interacting processes of aggradation and incision in terms of the threshold of critical power, which shifts with time and system equilibrium (Bull 1991, Harvey 1996, Sugai 1993).

Theories of fan dynamics developed in tectonically active settings, and in arid or semi-arid environments where water-limited vegetation cover is important in stabilising the substrate, may have limited application in the cool, humid, maritime, old-mountain setting of the Scottish Highlands. Here, fans are small, and development may be strongly controlled by the legacy of glacial erosion and deposition as well as the gradient of the source area (Brazier 1987). Sediment availability and source area and gradient have also been emphasised by Brazier (1987) and Harvey (1996).

A combination of fluvial activity and debris flows may contribute to both fan and cone formation resulting from redistribution of glacially deposited sediment (Brazier *et al* 1988, Wells & Harvey 1987). The size and shape of the source area may control

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whether debris flow or stream flow is dominant, and whether transitional forms are found (Brazier 1987). At either end of this series of coarse/steep, to fine-grained /low angle, landforms and processes, the genetic and sedimentological distinctions between cones (built from debris flows) and fans (built by fluvial processes) are clear. However, cone and fan processes and resultant surface dips are an intermixed series: flushes of water may deposit crudely sorted sand between debris flows; some fluvial flows may deposit crudely stratified and sorted debris.

1.6.4   Cones and fans in Northern Scotland

In Northern Scotland, fan and cones are small in comparison to those from arid and tectonically active parts of the world, and distal fining was developed only in two of the sites investigated for this thesis. Typically, their maximum diameter ranges from tens to a few hundred metres (Ballantyne & Whittington 1999, Brazier 1987, personal observation: see Chapters 4 and 6).

It was evident from fieldwork for this project (Chapters 4, 6) that

- a landform defined as a cone on the basis of process may have a surface slope which diminishes towards the distal end in a fan-like way
- a cone may be extended and lowered to form further cones
- a cone may also, or alternatively, be fluvially incised and extended to form a fan
- a cone may be partially fluvially reworked and/or fluvially extended
- water-dominated flow may occur between (or in series with) episodes of debris flow in a cone.

The terminology used in succeeding chapters is intended to avoid potentially ambiguous labels such as composite cone - which could mean nested cones, successively extended cones, or in the terminology of Brazier and Ballantyne (1989), aggradation through a combination of fluvial and debris flow purposes. Instead, definitions based on morphology are qualified in the text by an account of contributing processes. This allows form, process and stratigraphy to be related and interpreted. In general

Cone is used to mean a relatively steep, coarse, bouldery accumulation which does not spread at the base in fan-like lobes.

Fan is used for fan-shaped accumulations (which usually have a lower slope angle than cones), even if the dominant process has been debris flow.

Alluvial fan is used where water-dominated processes demonstrably account for the bulk of extension and aggradation of the landform.

1.7 DISCUSSION

Paraglacial relaxation appears to be a robust hypothesis to account for rapid early Holocene landscape adjustment in northern Scotland. However, it has not been adequately tested as an explanation for mid and late Holocene landscape transformations. Several additional aspects of the environmental inheritance - anthropogenic impact, neotectonic events, intrinsic local feedbacks, and climate - apparently have, in theory, the ability to generate renewed activity. External forcing factors have yet to be shown to be significant in Northern Scotland, but some evidence is beginning to emerge that rates of Holocene bedrock weathering have been non-uniform, and that they are capable of creating threshold conditions. This constitutes a competing hypothesis for the argument that paraglacial relaxation, triggered by randomly occurring intense storms, and overlain by human impact, has continued to be the dominant influence on slope landform transformations through to the late Holocene.
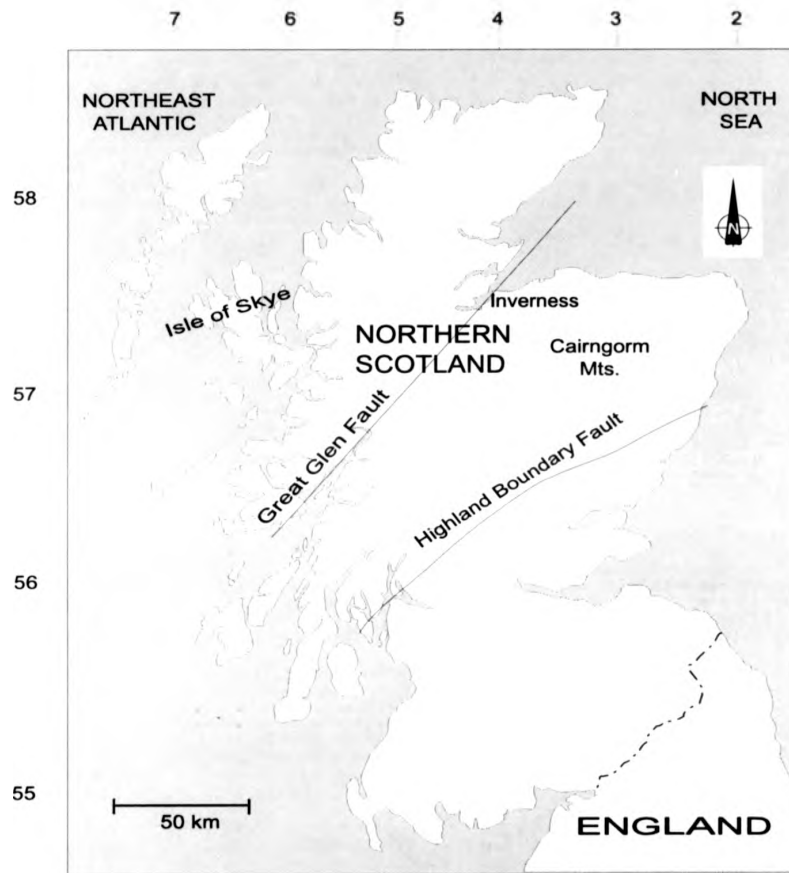
FIGURE 1.1
Northern Scotland
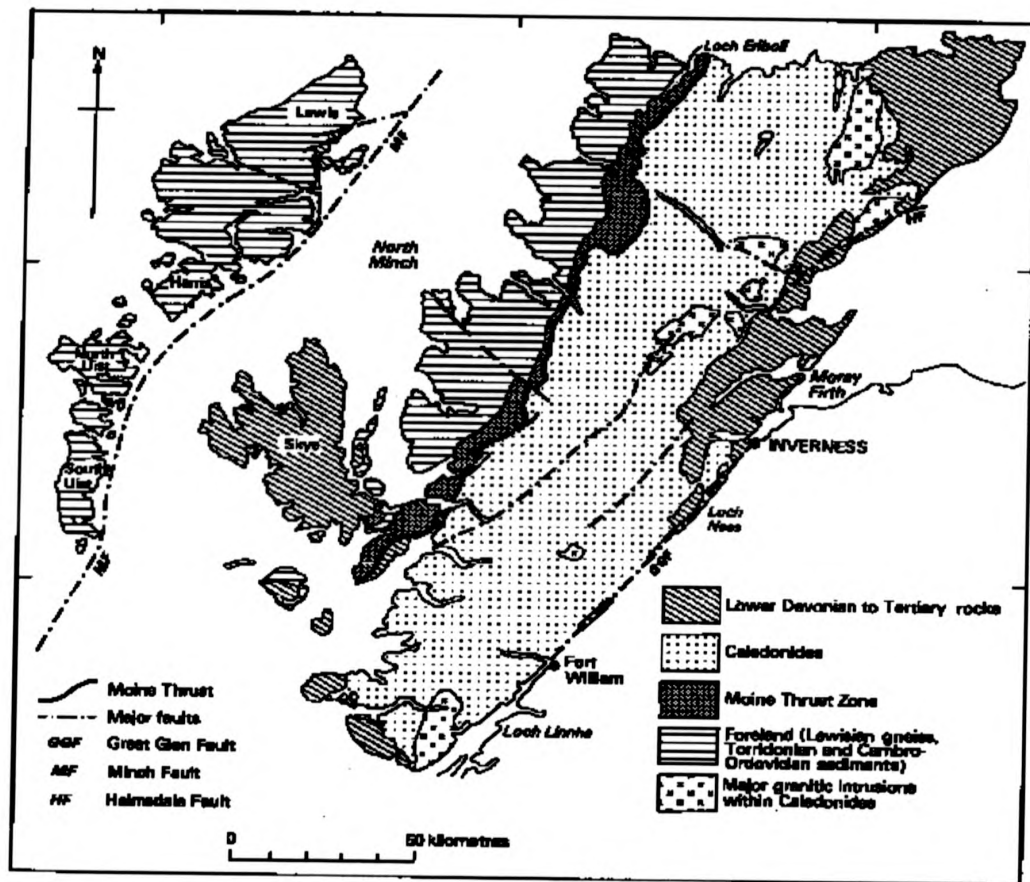
FIGURE 1.2

Geological sketch map of Northern Scotland

(Johnstone & Mykura 1989, Figure 2)
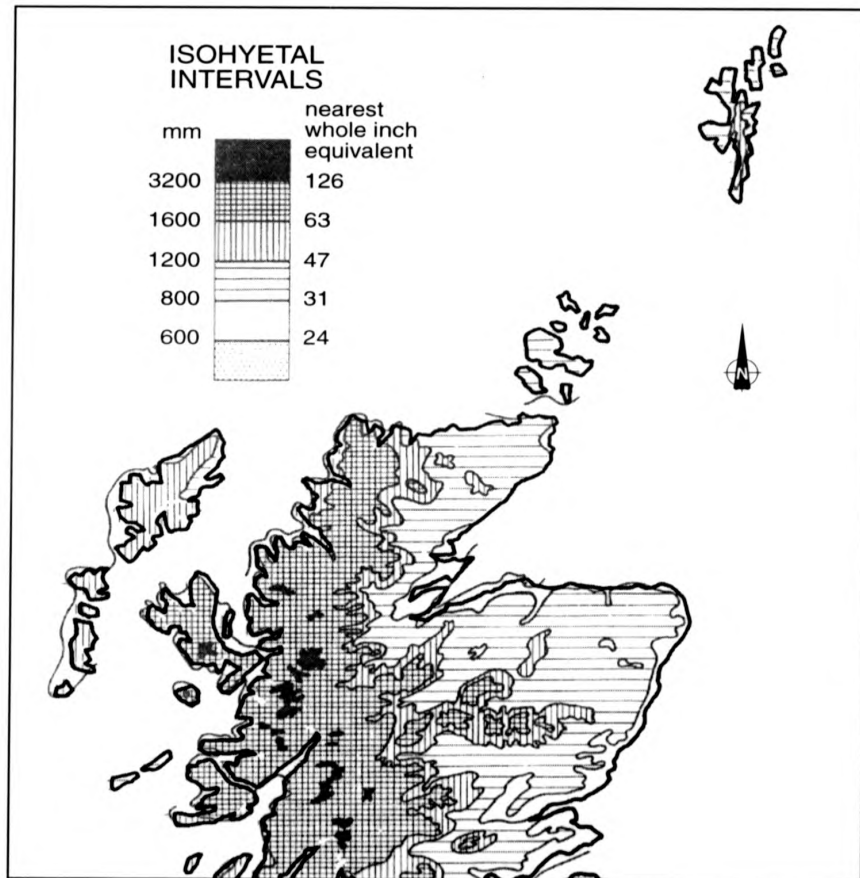
FIGURE 1.3
The East - West precipitation gradient in Northern Scotland

From 'The Climate of Scotland' The Meteorological Office, HMSO 1989
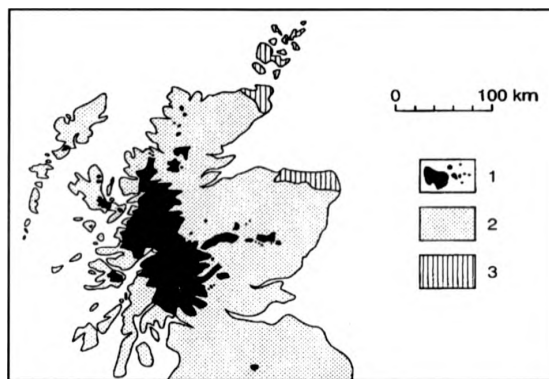
FIGURE 1.4

1   The extent of ice cover during the Younger Dryas (Loch Lomond) Stadial
2 & 3   Areas of severe periglaciation as Devensian ice retreated and during the
       Younger Dryas

(adapted from Ballantyne & Harris 1994, Fig. 2.8)

CHAPTER 2

DATED EVIDENCE FOR HOLOCENE LANDSCAPE EVOLUTION IN MOUNTAIN AREAS OF NORTHERN SCOTLAND, AND ITS ENVIRONMENTAL TRIGGERS

2.0   INTRODUCTION

Holocene environmental change in northern Scotland has been the subject of detailed investigation, using dated pollen stratigraphies and studies of the history of woodland development, particularly in relation to Scots pine (*Pinus sylvestris*).   Overviews of publications on these topics are found in, for example, Bennett (1995), Birks H. J. B. (1996), Gordon & Sutherland (1993) and Huntley (1996). The aim of this chapter is to review the far smaller number of investigations which provide dated evidence for those aspects of environmental change encompassed by the term *landscape evolution,* as well as insights into associated environmental triggers. There is a wider literature on landform-altering events in northern Scotland during the Holocene (e.g. Auton 1990, Ballantyne 1991, 1993, Ballantyne & Whittington 1987, Crofts 1999, Grieve *et al* 1994, Hinchcliffe *et al* 1998, Maizels & Aitken 1991, McEwen & Werritty 1988, Salt & Ballantyne 1997). It is excluded from the following review since, without dated material, the timing of the events is either only broadly bracketed or has been inferred from likely association with other data.

Landscape evolution is defined here as changes in morphology and surfaces, operating at the level of both individual landforms and associated groups of landforms (*sensu* Boardman 1985, Brunsden 1993, Brunsden & Thornes 1979, Bull 1991, Sugden *et al* 1997). The interpretation of such changes requires a chronostratigraphic framework. Event stratigraphies, *sensu* Ager 1973, are a means of defining accurate chronostratigraphic scales based on geographically separate, but widespread horizons produced by events such as marine transgressions, volcanic eruptions or polarity reversals which produce punctuated changes in the stratigraphic record. An extended usage applies the term to a sequence of diverse but related responses, for instance those associated with Holocene environmental changes, or the termination of the last Stadial (Birks 1996, Bjørck *et al* 1998). Birks' (1996) 'event stratigraphy' which summarises the Holocene history of the Cairngorms (Fig. 2.1), illustrates the current predominance of palaeoecological data. It also implicitly extends the definition of the term by including as 'events', active processes such as soil paludification, which led to changes

in vegetation. Event stratigraphies have been little used in U.K. Holocene geomorphology. But since a chronostratigraphic scale is a prerequisite for correlation of data from dispersed sites, for assessing the random or systematic occurrence of geomorphic events and processes, and for understanding their inferred environmental triggers, the concept provides a useful theoretical framework which can be used to assess the geomorphic literature. It can also be applied to understanding and integrating new data.

After a review of the literature, the final section of this chapter sets out the consequent rationale for work undertaken for this thesis. Research aims and objectives, along with methodologies, are defined in Chapter 3.

2.1   DATED EVENTS

Table 2.1 (2 pages overleaf) lists the sixteen currently available radiocarbon (and in one case, cosmogenic isotope) studies providing dated events leading to slope evolution in northern Scotland. Site locations are numbered on Figure 2.2 and cross referenced in Table 2.1 and the text (in square brackets). Figure 2.3 presents the data from Table 2.1 in graphical form, with the dates for each type of mass movement grouped, and shown with error bars.

Four broad categories of event are shown in column B of Table 2.1: high altitude solifluction, slopewash and soil erosion, hillslope debris flows, and cone/fan surface aggradation. One date is also available for floodplain incision, and one for stream incision of a cone/fan complex. The dataset in Figure 2.3 constitutes the current chronostratigraphic record of Holocene landscape evolution in northern Scotland. Rapid slope mass movement has been described in only ten of the sixteen publications. The remainder date high altitude solifluction. As a result, information on each type of mass movement is sparse, and/or limited in time range. Where events are defined only as having occurred after a particular date (in 33% of cases), this is emphasised by a question mark in Figure 2.3.

To allow comparison between studies, some dates in Figure 2.3 have been recalibrated using CALIB Rev 3.0. (Stuiver & Reimer 1993), and expressed as a range of 2σ (c. 95%) probabilities. Some authors have previously rounded dates to define the approximate period of activity within 95% confidence levels. Where these were already calibrated according to Stuiver & Reimer (1993), they are reported in the original form.

	Date/cal BP 2	Processes	Authors	Location [Fig 2.2 Ref. No.]	Age range
1					
2	between 8500 and 6800	enhanced rates of slopewash	Huntley 1981	Coire Fee, SE Cairngorms [1]	8500-7000
3	between 8000 and 7000	enhanced rates of slopewash	Tipping 1995	Carn Dubh, east-central Grampians [2]	
4	between 7000 and 6000	large rotational landslips	Ballantyne et al 1998	Trotternish, Isle of Skye [3]	7000-6000
5	between 6500 and 6200	hillslope debris flows	Curry 1999	Glen Docherty, NW Highland [4]	
6	after 6400-6010	gravel terrace on dissected fan	Bain et al 1993	Glen Feshie, W. Cairngorms [5]	
7	between 6250 and 5800	enhanced rates of slopewash	Huntley 1981	Coire Fee, SE Cairngorms [1]	6000-5000
8	after 5920-5310	high altitude solifluction	Sugden 1971	Cairngorms [6]	
9	between 5900 and 5600	debris flows on talus slope	Hinchcliffe 1999	Trotternish, Isle of Skye [7]	
10	after 5890-4280	debris cone aggradation	Brazier et al 1988	Glen Etive, SW Highlands [8]	
11	after 5600-5300	hillslope debris flows	Curry 1999	Glen Docherty, NW Highlands [4]	
12	between 5550 and 4850	enhanced rates of slopewash	Tipping 1995	Carn Dubh, east-central Grampians [2]	5000-4000
13	between 4900 and 4600	hillslope debris flows	Curry 1999	Glen Docherty, NW Highlands [4]	
14	between 4750 and 4300	enhanced rates of slopewash	Huntley 1981	Coire Fee, SE Cairngorms [1]	
15	between 4700 and 4000	enhanced rates of slopewash	Tipping 1995	Carn Dubh, east-central Grampians [2]	
16	after 4600-4260	high altitude solifluction	Mottershead 1978	Arkle, NW Highlands [9]	
17	after 4080-3720	gravel terrace emplacement	Robertson-Rintoul 1986	Glen Feshie, W. Cairngorms [10]	4000-3000
18	between 3550-2890 & 2120-1730	enhanced catchment erosion	Edwards & Rowntree 1980	Braeroddach Loch, E. Grampians [11]	
19	after 3140-2630	high altitude solifluction	Sugden 1971	Cairngorms [6]	3000-2000
20	between 2700 and 2400	floodplain incision	Ballantyne & Whittington 1999	Edendon Valley, central Grampians [12]	
21	between 2200 and 2100	alluvial fan aggradation	Ballantyne & Whittington 1999	Edendon Valley, central Grampians [12]	
22	between 2300 and 1700	debris flows on talus slope	Hinchcliffe 1999	Trotternish, Isle of Skye [7]	
23	after 2120-1830	debris cone aggradation	Brazier & Ballantyne 1989	Glen Feshie, W. Cairngorms [13]	
24	after 2120-1730	slopewash on talus slope	Innes 1983b	Trotternish, Isle of Skye [14]	
25	between 1900 and 1800	alluvial fan aggradation	Ballantyne & Whittington 1999	Edendon Valley, central Grampians [12]	2000-1000
26	between1050-570 and 660-330	rapid high altitude solifluction	Ballantyne 1986a	Fannich Mts. Northern Highlands [15]	1000-500
27	between 900 and 700	alluvial fan aggradation	Ballantyne & Whittington 1999	Edendon, central Grampians [12]	
28	after 710-560	slope wash on talus slope	Innes 1983b	Trotternish, Isle of Skye [14]	
29	betweeen 700 and 500	debris flows on talus slope	Hinchcliffe 1999	Trotternish, Isle of Skye [7]	
30	after 690	enhanced catchment erosion	Edwards & Rowntree 1980	Braeroddach Loch. E. Grampians [11]	
31	after 650-510	cone incision and reworking as fan	Brazier et al 1988	Glen Etive, SW Highlands [8]	
32	after 510	debris cone aggradation	Brazier & Ballantyne 1989	Glen Feshie, W. Cairngorms [13]	500-modern
33	after 450	hillslope debris flows	Curry 1999	Glen Docherty, NW Highland [4]	
34	after 250	mainly unconfined debris flows	Innes 1983a	SW & NW Highlands & Cairngorms [16]	
35	TABLE 2.1 DATED HOLOCENE GEOMORPHIC EVENTS IN NORTHERN SCOTLAND (page 1 of 2)				

	F	G	H
1	material dated	Fig. 2.2 ref No.	Comment
2	?	1	Calibrated by Tipping 1995, based on Huntley's unpublished PhD thesis, 1976
3	sandy horizons between dated peats	2	Correlated with independent evidence for increased precipitation
4	cosmogenic Cl dates from failure surfaces	3	Result of progressive failure - not post glacial relaxation
5	palaeosols beneath debris flows and slopewash	4	Timegap between soil C age & event?;soil C residence time?pretreatment?
6	top 5mm of organic horizon	5	Timegap between soil C age & event?;soil C residence time?pretreatment?
7	dates calibrated by Tipping 1995	1	
8	buried peat;solifluction probably at a later date	6	Solifluction need not be concident with climatic deterioration (Ballantyne 1991)
9	palaeosols beneath debris flows & slopewash	7	Non-anthropogenic, possibly related to extreme climate-controlled precipitation
10	top 5mm of immature soil	8	End of debris cone accumulation; final phase of local paraglacial relaxation
11	palaeosols beneath debris flows and slopewash	4	
12	sandy horizons between dated peats	2	Increased precipitation?
13	palaeosols beneath debris flows and slopewash	4	
14	dates calibrated by Tipping 1995	1	
15	sandy horizons between dated peats	2	Grazing presssure?
16	peat at 745m OD (above present/palaeo treeline)	9	20mm samples; probably rapid after dated upper surface; v. slow later (incl by implicatn. LIA)
17	charcoal in 30mm A horizon of buried podzol	10	Effect of charcoal reworking & range of ages? Time gap charcoal age/terrace deposition?
18	organic lake sediment	11	Followed by reverse dates interpreteted as denudation of older sediment on slopes
19	peat ; solifluction probably occurred later	6	
20	peat	12	Non-anthropogenic;triggered by very rare, large storms; not climatic
21	peat	12	
22	palaeosols beneath debris flows & slopewash	7	
23	buried soil	13	Possibly caused by local slope undercutting due to river channel migration
24	humified organic matter	14	Non-anthropogenic; probably interaction of local weathering and meteorological events
25	peat	12	
26	buried soil	15	
27	peat	12	
28	humified organic matter	14	
29	palaeosols beneath debris flows & slopewash	7	
30	organic lake sediment	11	Followed by reverse dates (erosion of older sediment);indistinguishable from modern date
31	top 5mm of buried soil	8	Max. age for initiation of fluvial reworking of debris cone; possibly due to vegetation clearance
32	5 buried soils; one root	13	Slope stable during previous 1500 years then bulk of cone formed during this time interval
33	palaeosols beneath debris flows and slopewash	4	Timegap between soil C age & event?;soil C residence time?;LIA extreme storm/treeclearnce?
34	lichens on debris flow surfaces	16	Attributed to heather burning and overgrazing, not rainfall variation or the 'Little Ice Age'
35			TABLE 2.1 contd. EXPLANATION (page 2 of 2)

As a preamble to discussing what conclusions can be drawn from the synthesis in Table 2.1 and Figure 2.3, the data are reviewed below as a sequence through time. Arbitrary, millennial or 500 year divisions are used, to avoid the implication that events are clustered, hence causally linked. However, Figure 2.3 allows the presence, absence and significance of any temporal groupings to be assessed. Large error bars mean that some age ranges cross the boundaries of the time intervals defined below. These are noted in the discussion. The number of records in each millennial/quintennial division conceals the fact that some authors have defined a period of time identified as *between date A and date B* within which several - or in one case, many - discrete events were dated.

2.1.1 8500 - 7000 cal BP

Two dated sequences at sites in the Grampian and Central Highlands are about 25 km apart (Huntley 1981 [1], Tipping 1995 [2]). They contain organic material, or mineral inwashing in organic sediments with ages older than 7000 cal BP. In both cases small peaty upland basins received relatively high volumes of sandy slopewash over a period of a millennium or more.

2.1.2 Between 7000 and 6000 cal BP

Large rotational landslips on the NE coast of the Isle of Skye occurred between 7000 and 6000 cal BP (Ballantyne *et al* 1998 [3]), probably occurring in a single episode, although cosmogenic ^{36}Cl dating (6500±500 cal BP) did not allow finer resolution of the timing. Within the same time interval, a palaeosol on a steep slope in the NW Highlands was buried by a debris flow (Curry 1999 [4]). The emplacement of a thick gravel terrace by the incising stream of a coarse fan (Western margin of the Cairngorm massif, Bain *et al* 1993 [5]) may also have occurred within this time interval. No common trigger for the three events has been identified in the literature: large rotational landslips on the Isle of Skye were interpreted as the product of long term, progressive failure, while both the rapid deposition of a gravel terrace on a formerly vegetated surface in the Cairngorms, and hillslope debris flows in the NW Highlands, may reflect exceptionally high water tables and/or sediment availability. The reasons for both progressive failure and these latter conditions remain to be established.

2.1.3 Between 6000 and 5000 cal BP

A period of accelerated slope wash spans the boundary with the previous millennium (Huntley 1981 [1]). Other processes described as active during this millennium include

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high altitude solifluction in the Cairngorms (Sugden 1971 [6]), debris cone aggradation in the SW Highlands (Brazier *et al* 1988 [8]) (but possibly not until the next millennium; Fig. 2.2), talus slope modification in NE Skye (Hinchcliffe 1999 [7]), accelerated slope wash (Tipping 1995 [2]), and hillslope debris flows which buried soil dated at between 5600 and 5300 cal ka BP (Curry 1999 [4]). The advance of solifluction lobes may be widely separated in time from the ages of the organic horizons they engulf (Ballantyne 1986a [15], White & Mottershead 1972). Moreover, according to (Ballantyne 1991) rapid solifluction in Holocene times cannot necessarily be attributed to episodes of colder, stormier weather since a causal connection between climate deterioration and solifluction has not yet been established.

All the processes in this time interval indicate slope instability, but on the basis of the authors' comments, no conclusions can be drawn about a common cause. Debris cone aggradation in the SW Highlands was interpreted by Brazier *et al* (1988) as the final phase of exhaustion of glacigenic sediment, after which the cone surface became vegetated and remained stable until human intervention at a much later stage. Sediment analysis led Hinchcliffe (1999 [7]) to the conclusion that talus slope modification in NE Skye was due to progressive addition of fines weathered from the cliff face since deglaciation. This altered the composition and stability thresholds of the sediment wedge, and resulted in renewed movement of slope debris. Up to 30% of rockwall retreat since deglaciation above talus slopes in the northwest Highlands was found to be due, not to rockfall, but to Holocene granular weathering (Salt & Ballantyne 1997, Hinchcliffe *et al* 1998).

2.1.4 Between 5000 and 4000 cal BP

One event, or set of events, has been dated to the first half of the 4th millennium BP; (hillslope debris flows, Curry 1999 [4]). Two others - high altitude solifluction in the NE Highlands (Mottershead 1978 [9]), and accelerated slopewash into a small upland basin in Perthshire (Tipping 1995 [2]) - have been placed in the latter half of the millennium. Huntley's (1981 [1]) data for accelerated slope wash spans the interval between them. All four datasets reflect slope instability, but the wide range of the probable calibrated ages produces an overlap that prevents any clustering of events within this millennium being proposed.

Within the same time period palaeoecological evidence, specifically pine woodland retreat, and increases in water-tolerant plant species, have been interpreted as indicating

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a deteriorating climate (e.g. Dubois & Ferguson 1985, Gear & Huntley 1991, Pennington *et al* 1972). But on the basis of so few dated landform changes of diverse types, landform evolution cannot securely be attributed to climate change, since it is not possible to exclude alternative and/or complementary causes of slope instability such as purely local developments or progressive change such as soil leaching and paludification.

2.1.5 Between 4000 and 3000 cal BP

Gravel terrace emplacement was recorded by Robertson-Rintoul (1986 [10]) at an early point in this time interval. The next dated event, in the second half of the millennium, and extending far into the next one, is initiation of accelerated catchment erosion which may then have persisted over a long period (Edwards & Rowntree 1980 [11]). The large uncertainty on the latter dates could place this record in the following millennium. Landform change between 4000 and 3000 cal BP was either very limited or is poorly recorded.

2.1.6 Between 3000 and 2000 cal BP

In addition to the catchment erosion described by Edwards and Rowntree (1980 [11]), three other sites in mountain areas of northern Scotland have been described as active in this interval. Processes described are high altitude solifluction (Sugden 1971 [6]), floodplain incision and alluvial fan aggradation (Ballantyne & Whittington 1999 [12]), and debris flows on a talus slope (Hinchcliffe 1999 [7]). The immediate trigger for the fan aggradation, fluvial incision and slope debris instability was by implication, in each case, high porewater pressures. But the reasons for their presence can be difficult to infer unambiguously from landforms and stratigraphy. High water tables may trigger system transformations during either intense storms or as a result of persistent moderate rainfall which causes the infiltration capacity of sediment and bedrock to be exceeded (Kirkby 1978). Shallow mass movement is therefore not necessarily a response to intense precipitation or sudden snowmelt, and is mediated by the state of the Earth materials - itself a secular variable. Permeability contrasts created by soil profile development, internal changes in slope stability resulting from Holocene weathering, progressive sediment accumulation upslope, and local change in groundwater flow paths, may also be implicated.

Anthropogenic impact on mountain slope vegetation has been invoked to explain soil

and sediment instability (Ballantyne 1986b, Brazier *et al* 1988, Innes 1983a). But there is also some evidence consistent with rapid mass movement having occurred in the absence of any human activity. A talus covered slope in northern Skye was the site of discrete debris flows, widely separated in time (Innes 1983b [14]), and palynological data was found to be inconsistent with human impact on fan aggradation in the central Grampian mountains (Ballantyne & Whittington 1999 [12]). However, data on past sediment yields and weathering rates in northern Scotland are sparse (Ballantyne 1991). And although average weathering rates have been calculated on the timescale of the whole of the Holocene, their variation in time and space is poorly known (Bain *et al* 1990). Large, temporary increases in sediment yields have been measured in relation to planting and clearfell in forestry plantations in recent times (e.g. Carling *et al* 1993, Johnson 1993). However, mechanised ground preparation and monoculture harvesting are a poor analogy for pre-industrial land use. Variable sediment yields in a small basin in the Eastern Grampian mountains were attributed by Edwards & Rowntree (1980 [11]) to human impact most securely within the last few hundred years. But these authors concluded that earlier evidence of enhanced sedimentation was less clearly distinguishable from the effects of progressive soil leaching, vegetation succession, and possibly climate variability.

2.1.7 Between 1000 and 500 cal BP

The first record in this set - high altitude solifluction (Ballantyne 1986a [15]) - may extend into the post-500 cal BP group. It is followed by two studies of talus slope modification (Hinchcliffe 1999 [7], Innes 1983b [14]), renewed deposition on the surface of an alluvial fan after a prolonged period of stability (Ballantyne & Whittington 1999 12]), and extension of a debris cone to form an incised alluvial fan (Brazier *et al* 1988 [8]) again after prolonged stability on the surface of a vegetated cone. Build-up of high porewater pressures due to intense rainfall combined with substrate weathering are cited as the most likely immediate triggers.

Anthropogenic causes are reported as unlikely, except at the site in Glen Etive examined by Brazier *et al* (1988 [8]) where some palynological evidence is consistent with increased water flow and cone sediment reworking coinciding with the onset of cultivation. These authors found that stream incision of the upper part of a cone had led to redeposition of fluvially worked sediments at its lower end within the last 600 years. Charcoal and pollen analysis of a buried soil on the lower cone, placed the timing of

incision and cone extension by alluviation after cone surface clearance by burning. Subsequent agricultural and/or pastoral land use was indicated.

However, the argument that shrub and woodland clearance by burning destabilised the slope and caused alluvial incision and cone extension, remains difficult to substantiate, because vegetation clearance at the lower end of the cone is not easily linked to sediment movement at the upper end. The cone, whose surface gradient was about 14°, was composed of coarse bedrock detritus, and a mature podzol profile on its surface indicated a freely draining landform. Burning off shrubs in the growing season (approximately 6 months of each year) would have raised the water table by removing the pumping effect of plant transpiration (Greenaway 1987). But burning outside the growing season in early spring or autumn when the vegetation was dormant would have had little effect on groundwater and soil moisture conditions. Since the cone had previously been stable for several millennia, it must have had a strong capacity to resist system transformations by dispersing shallow groundwater in a diffuse way regardless of variations in seasonal precipitation and snowmelt. Shrub and tree removal in the growing season seems therefore an uncertain trigger for changing cone hydrology sufficiently to initiate channelled flow.

An alternative or perhaps complementary trigger for fluvial incision, is unusually intense rain or snow melt. The source area - a long, steep, rock slope with several feeder gullies rising to over 750m, is compatible with this hypothesis, and the flow energy for incision seems likely to have been generated at the cone apex rather than on its lower slopes. Vegetation clearance could not have triggered incision in this scenario, since the source area for runoff was unvegetated. Stream incision of cones and fans is further discussed in Chapters 6 and 7.

2.1.8 Within the last 500 years

The three studies from this time interval listed in Figure 2.2 include several hundred lichenometric dates from debris flow surfaces in three widely geographically separated areas of northern Scotland, the majority within the last 250 years (Innes 1983a [16]). Innes' dating method was not capable of resolving surfaces older than 700 years (with increasing uncertainty attendant on the oldest dates) and skews the balance of evidence towards recent events, since later flows often overlie earlier ones (Ballantyne 1991). On the other hand, in the absence of a mechanism for removing material from lower slopes, deposits are far too thin in total for the high rates of activity in the last few hundred

years to have been characteristic of the whole of the Holocene. Innes' conclusion that grazing and burning were the most likely causes of accelerated slope instability remains to be confirmed. Some source areas, for instance, the slopes of the Lairig Ghru, a glacial breach in the Cairngorm massif, are currently so poorly vegetated as to make this implausible. Moreover, as discussed above, tree and shrub removal by humans or grazing animals, cannot be assumed to affect slope hydrology so as to destabilise slope sediments and landforms.

A second study indicating enhanced activity in the last few hundred years was based on several radiocarbon dated buried soils and debris cone growth in upper Glen Feshie on the western margin of the Cairngorm massif (Brazier & Ballantyne 1989 [13]). Slope instability was triggered after a period of some 1500 years devoid of evidence for instability. Slope undercutting by the river channel at the exit from a narrow, rock-walled, gorge section was suggested as a trigger. Other possible explanations include progressive changes, with *in situ* weathering of slope debris altering sediment composition and permeability, hence critical thresholds. Local bedrock is psammitic to semipelitic schist intruded by granitic veins (personal observation). The most important Holocene chemical weathering process in the Highlands has been identified as hydrolysis (Bain *et al* 1993), and its predominant solid products in mafic minerals in schist, are clays and micas. Both have a platey structure, which, in conjunction with a finer grain size than their parent minerals, would tend to decrease frictional forces in wedges of slope sediment. The absence of organic material older than about 2000 cal BP at this site could be the result of a very long maturation period for talus destabilisation triggered by progressive weathering. Alternatively, earlier episodes of slope undercutting adjacent to the gorge, could have removed debris with developed soil horizons.

In Glen Docherty in the northwest Highlands, Curry (2000) suggested intense rainfall and/or woodland clearance as the trigger for a recurrence of slope instability after 450 cal BP. Landslips and debris flows have been common in parts of the northern Highlands within the last few decades (e.g. Ballantyne 1986, Innes 1983a). Although the timing of debris flows suggests a causal link with the tree removal which has occurred in historical times, it has not been demonstrated, and in the context of the whole Holocene record, the association of anthropogenic woodland clearance and slope instability appears more speculative. On active slopes, recent deposits obscure older

ones, and good exposure of the stratigraphy is infrequent. Preservation and exposure are therefore likely to favour detection of the most recent events. Where the stratigraphy is dissected, typically coarse deposits may have erosional bases, and thin organic layers which provide datable evidence of time gaps between deposits may be susceptible to erosion. So arguably, the more frequent the events, and the more sensitive the site to change, the less likely are the intermediate organic horizons essential for dating to a) develop and b) be preserved. Events which are distant in time may therefore be most easily detected at sites which favour preservation of organic horizons that have had ample time to develop. Such conditions pertain where geomorphic activity has punctuated extended periods of stability.

If this scenario is realistic, records are likely to be biased towards resolving frequent events in the recent past on the one hand, and rare, episodic events in the more distant past on the other. The relative abundance of debris flows in the last 200 years or so may thus reflect (to an unknown degree) the difficulty of detecting older unconformities in piles of typically very coarse, poorly dissected sediment, and of dating many landforms within more than a wide age bracket. In conclusion, there is some uncertainty about the extent to which the apparent acceleration of rates of activity in the last few hundred years, as codified in Figure 2.3, is an artefact.

2.2   EXTERNAL FORCING FACTORS PROPOSED IN NORTHERN SCOTLAND

The literature focuses on seismic, climatic and anthropogenic forcing factors, and to a lesser extent, progressive change. The evidence for this is discussed in the following sections. The question of whether conclusions can be drawn about temporal clusters of geomorphic events is revisited in Section 2.3.

2.2.1   Seismicity as a trigger for landscape evolution

The extent and timing of seismicity has been closely related to rapid isostatic recovery during and immediately after ice retreat. This is thought to have produced numerous slope failures adjacent to fault traces and ice margins (Fenton 1991, Holmes 1984, Peacock & Cornish 1989). Surface fault scarps associated with small stream diversions (Fenton 1991, Ringrose 1989) indicate some later, Holocene, activity, but the number and distribution of such features remains to be confirmed (Stewart 1998). It is unlikely that after the immediate aftermath of deglaciation, surface deformation has been more than locally important (Stewart 1998). Although no radiocarbon dates were available,

Fenton (1991) took the view that faulting which visibly disturbed blanket peat must have occurred after about 6000BP, since widespread development of blanket peat had been reported from this date onwards. His assumption is incompatible with several dates for basal blanket peat older than 9000BP (Birks H. H. 1975, Charman 1992, Haggart & Bridge 1992, Tipping 1995). Early and mid-Holocene seismicity cannot, therefore, be differentiated on the basis of deformed peat in fault zones.

The direct effects of Holocene faulting have not therefore been demonstrated to have substantially affected landscape evolution. However indirect effects (Chapter 1, Section 1.5.7) may have conditioned subsequent system behaviour.

2.2.2 Climate and slope mass movement

Three sources of Holocene palaeoclimate information for NW Europe are

- General Circulation Models (GCMs)

- The marine sedimentary record

- The terrestrial record (stratigraphy, pollen stratigraphy, lake level changes, distribution of fauna and flora, dendroclimatology).

2.2.2.1   General Circulation Models (GCMs)

In general, the Holocene climate history of Europe is still poorly understood (Harrison *et al* 1997). However, there is consensus on warmer and drier conditions in NW Europe as a whole from about 6-5 ka BP (radiocarbon years), and wetter, cooler conditions thereafter, especially after c. 3kaBP (Cheddadi *et al* 1996, Harrison & Digerfeldt 1993, Houghton *et al* 1995, Huntley & Prentice 1988, Kutzbach *et al* 1993, Lamb 1995, Prentice *et al* 1998).

2.2.2.2   The Marine Record

Marine stratigraphy records well-defined millennial-scale oscillations of warm and cold water in the North Atlantic and North Sea during the Holocene (Alley *et al* 1997, Bond *et al* 1997, Jiang *et al* 1997, Bianchi & McCave 1999). The 'Little Ice Age' is increasing recognised as simply the most recent of these events rather than as an exceptional climatic episode (McManus & Oppo 1999, Oppo 1997). Iceland-Scotland overflow water (ISOW) - a cold mass of surface water - has been identified as an important component of the thermohaline circulation which modulates European climate (Bianchi & McCave 1999). Peak fluxes of cold surface water are believed to have occurred in the

North Atlantic and Nordic Seas at about 5900, 4200, 2800 and 1400 cal BP, bringing cool, ice-bearing water characteristic of the Iceland-Faroes Front as far south as Ireland at the expense of the warm North Atlantic Current (Bond *et al* 1997). Estimated surface water cooling during these southward advances of polar water was about 2°C. Off the coast of Denmark a warm period from 7.7-5.1 cal ka BP followed by cooling, was identified from faunal changes (Jiang *et al* 1997). If these results are representative, they suggest different patterns of warming and cooling off the Atlantic and North Sea coasts of Scotland.

2.2.2.3   *The Terrestrial Record in Northwest Europe*

Heat flux at the Earth's surface is controlled by coupled atmosphere-ocean systems, and is reflected in terrestrial as well as oceanic palaeoclimate proxies. But the effect of ocean cooling on terrestrial systems in northern Scotland is likely to have been mediated in complicated ways by the influence on the heat budget of relative sea level rise and fall around the North and Baltic seas, and by geographic and climatic zoning (Goudie 1992).

Holocene landscape evolution has been described as a consequence of the changing frequency and/or intensity of storms and/or floods associated with low amplitude Holocene climate variation (e.g. Matthews *et al* 1996, Rumsby & Macklin 1996, Sandersen 1997). In southern and western Norway, colluvial systems have been described as highly sensitive recorders of regional climatic changes, with enhanced activity linked to dated periods of glacier advance and increased solifluction (Blikra & Nesje 1997, Matthews *et al* 1997). The latter authors describe slope stability and relatively low water tables between 8400-8100 and 4440-3980 cal BP (2σ confidence levels). Between the latter date and 2710-2190 cal BP, coarse debris flows and thin, poorly humified peats separated by 'distal flow deposits' (See Section 2.2.2.5 below) were interpreted as the outcome of high water tables and frequent slope instability. Thereafter, continuous peat accumulation until 1560-1320 showed that the slope subsequently stabilised for a period of 600 (min.) - 1200 (max.) years. A final debris flow post-dated 670-520 cal BP and, in conjunction with evidence for frost heaving of the dated peat underlying it, was interpreted as co-incident with the 'Little Ice Age'. Long dendroclimatological records from northern Fennoscandia (Eronen *et al* 1999) now suggest that climate deterioration at 2500 cal BP was probably more severe than during the so-called 'Little Ice Age'.

Results from Matthews *et al* (1997) conform with the model of punctuated slope evolution. However, punctuated slope evolution could arguably be a multi-proxy signal - the product of interacting processes which predominate at different times, rather than a series of responses to a single external trigger (in this case climate). For instance, it may be difficult to distinguish periods of slope stability arising from an absence of triggering mechanisms such as wetter and/or stormier climate, from stability due to sediment exhaustion, which is the result of a previous period of frequent slope mass movement. Renewal of mass movement after an extended period of frequent debris flows in some settings might, therefore, critically depend on progressive processes such as weathering and small scale mass wasting in the source area restoring the supply of material on which an external forcing factor could act.

An alternative approach to high resolution local studies is to make region-wide compilations of evidence for rapid slope mass movement. A synthesis of this kind was produced for Norway (Blikra & Nemec 1998). In alpine areas of Norway, unlike Scotland which has remained unglaciated since the end of the Younger Dryas Stadial, Holocene slope mass movements have been inferred to be closely associated with variations in snowfall, and have been interpreted, in conjunction with the timing of periods of glacier advance, as the outcome of colder climate. At lower altitudes, time lags between climate episodes and slope response have been inferred from dated debris flow sequences, and attributed to the time needed for debris production and time in interim storage after climate deterioration. The immediate trigger for such debris flows has been suggested as snowmelt associated with increased annual precipitation and/or anthropogenic activity, together with progressive weathering. By implication, sub-alpine debris flows are considered by these authors as less direct and reliable indicators of 'climatic deterioration' [cooling and/or increased precipitation] than the changing incidence of snow avalanches which entrain variable amounts of debris.

Blikra & Nemec (1998) detected a major hiatus in slope activity in Norway. (snow/debris avalanches, debris flows, rockfalls, and solifluction) between the end of the Younger Dryas Stadial and a resumption in the mid and late Holocene. They identified the most marked slope activity (quoted as uncalibrated, radiocarbon dates) as occurring between 4700 and 4200 BP, 3800 and 3000 BP, 2500 and 1800 BP, around 1400BP, and in the Little Ice Age as defined by Grove (1972). These active intervals include dated palaeosols which indicate fluctuating conditions with short spells of

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stability.

The value and difficulties of using shallow slope failures of various kinds as climate indicators in Europe (European Community EPOCH programme) were summarised by Brunsden & Ibsen (1997), and on a shorter timescale, Dikau & Schrott (1999). The former authors concurred with Berrisford & Matthews (1997), Blikra and Nemec (1998) and Matthews *et al* (1997) that mass movements on slopes carry a broad climatic signal judging by their contemporaneity with climate episodes inferred from independent proxy data such as glacier advance and pollen stratigraphy. But patterns of activity in northern Europe are not synchronous with those further south and east (Berrisford & Matthews 1997). Brunsden & Ibsen (1997) further concluded that current databases on a European scale suffer from deficiencies: they are very patchy in quality and in temporal coverage; they are supported by very limited radiocarbon dating; and over-generalisation from one or two dates has been used to indicate that a particular period was characterised by landsliding. Like Blikra and Nemec (1998) they found that landslide-rainfall relationships are subject to significant time lags in response patterns. Their data show that the main periods of landslide activity, in northern Europe based on upland, lowland and coastal sites, are (radiocarbon years BP) 7500 - 6000; 5500-3000; 2500-2000; 1400-1100; 1550-1850AD (Little Ice Age); and the modern period. However, they concluded that the availability of radiocarbon dates is 'almost random' so that it is difficult to know how representative the data are of total landslide occurrence frequencies.

Some site specific dates for slope evolution from alpine Norway and the uplands of northern England are shown in Table 2.2 overleaf. At 95% confidence levels, large error bars and overlaps are evident. So although the radiocarbon dates are consonant with mid and late Holocene renewal of slope instability within the broad time spans noted above, the existence of discrete episodes of enhanced activity is unclear. A complicating factor in comparing evidence for northern Scotland (Fig. 2.2) is that most data for rapid mass movements in Europe before historical times comes from alpine zones, or in the UK, from lowland and coastal sites where the record is affected by anthropogenic effects and changes in sea level respectively (Berrisford & Matthews 1997, Brunsden & Ibsen 1997).

Date/2σ cal BP	Processes	Authors	Location
4440-3980 to 3830-3470 [3840±80 - 3410±70BP]	Large debris flows	Matthews *et al* 1997	Alpine S. Norway
3690-3370 to 2730-2300 [3310±70 - 2410±95BP]	Frequent debris flows	Matthews *et al* 1997	Alpine S. Norway
1560-1320 to 980-720 [1570±60 - 960±70BP]	Debris flow	Matthews *et al* 1997	Alpine S. Norway
1290-950 to 1050-730 [1200±70 - 980±90BP]	Cone aggradation	Harvey & Renwick 1997	N. England
1050-670 [940±95BP]	Cone aggradation	Harvey *et al* 1981	N. England
After 670-520 [610±60BP]	Debris flow	Matthews *et al* 1997	Alpine S. Norway

Table 2.2
Dated geomorphic events in upland areas of Norway and England

2.2.2.4 *Evidence for geomorphic response to climate fluctuations and exceptional meteorological events in Northern Scotland*

This section focuses on the literature which invokes climate change in northern Scotland in seeking to explain the occurrence of high altitude solifluction, soil erosion from slopes, hillslope debris flows, and fan and cone aggradation during the Holocene.

High altitude solifluction

Solifluction has been dated to the fifth, fourth and second millennia BP, and to within the last thousand years. However, the advance of solifluction lobes may occur hundreds of years later than the formation of the dated organic horizons they over-run (Ballantyne 1986a, Mottershead 1978, White & Mottershead 1972). Although solifluction is the product of freeze-thaw processes, comparison of the timing of enhanced rates of solifluction in Norway with the incidence of snow avalanches and glacier advance, suggests that this phenomenon may be more closely related to increased annual precipitation than to climatic cooling (Blikra & Nemec 1998). In the Scottish Highlands, the timing of solifluction has not been securely matched to independent climate indicators.

Soil erosion

References to climate and/or extreme precipitation as triggers of soil erosion in northern Scotland have been noted in Section 2.1 and Fig. 2.2 above. Both increased precipitation and cooling (enhanced cryoturbation during the 'Little Ice Age') have been invoked. Establishing a positive correlation between enhanced soil erosion and palaeoclimate on the basis of currently available evidence poses some difficulties which are discussed below.

Edwards & Rowntree (1980) reported reversed dates in two sections of a sediment column in a small loch in the Grampian Highlands. They attributed this to rapid stripping of surface soils, accompanied by erosion and redeposition of the older substrate in the catchment sediment sink. On a timescale of thousands of years, rates of erosion were found to be uneven, varying by up to 300%. In Neolithic times, accelerated erosion appeared in parallel with some anthropogenic changes in vegetation communities. But relatively low, though still variable, sedimentation rates pre-dating archaeological evidence for human impact, were interpreted as a possible climatic response in a landscape undergoing climate-related vegetation changes. Nevertheless, these authors concluded that there was no means of establishing unequivocally whether climate change and/or human impact and/or a third possible factor - progressive, long-term leaching of mineral soils - was the principal, or sole trigger of slope soil erosion.

In a second study, Tipping (1995) matched episodes of intensified sandy inwashing derived from eroding slopes and deposited in a peaty basin with independent evidence for increased precipitation in Scotland. These data were: a stable isotope study of sub-fossil pine wood, (Dubois & Ferguson 1985), raised bog studies (Dupont 1986), and lake-level analyses (Digerfeldt 1988, Magny 1992, Mannion 1982). Tipping concluded that elevated levels of precipitation were co-incident with, and a credible explanation for, some phases of enhanced slope activity. Where the most intense sandy inwashings were time-discordant with independent evidence for climate, in particular with the inferred climatic amelioration which allowed a brief northward and westward expansion of pine woodland in the 5th millennium BP (Gear & Huntley 1991), but were associated with local pollen evidence for human settlement, Tipping proposed grazing pressure as an alternative explanation. Calibration of Gear and Huntley's (1991) dates to match those in Figure 2.2 shows the pattern summarised in Table 2.3.

Process	Date /calBP 2σ	Author
Intense slopewash	Between 4700 and 3850	Tipping 1995
Retreat of pine woodland to its former limits	Before 4410 to 3990 (3815±50 ^{14}C BP)	Gear & Huntley 1991

Table 2.3

Comparison of timing of reported response to mid Holocene climate change in N. Scotland (Gear & Huntley 1991) and human activity at Carn Dubh (Tipping 1995)

Gear & Huntley based their conclusions on the ages of sub-fossil pine stumps, correlated with pollen stratigraphic evidence for local presence of pine woodland. According to their Figure 2, by about 4200 ^{14}C BP (c. 4800 cal BP) only a small number of trees survived, and no new trees were able to grow. By about 3800 ^{14}C BP (c. 4300 cal BP) pine had completed an 80km retreat south and east to its former limits. But the retreat had been initiated some 400 radiocarbon years earlier. If that was, as inferred, a response to climatic deterioration, the deterioration must have begun well within the date range for Tipping's period of intense slopewash. There therefore seems to be little discrepancy between the timing of Tipping's slope erosion and the climatic deterioration invoked by Gear and Huntley to explain pine retreat. A climatic explanation for Tipping's (1995) slope instability is therefore a possibility, and the argument for an anthropogenic trigger less clear cut. The pollen record at this site indicates human activity, but the effect of vegetation change in accelerating rates of slopewash in mountain areas of northern Scotland, is, as discussed above, still poorly understood. On the basis of Gear & Huntley's inference about the cause of pine retreat, a climate trigger, or at least climatic sensitisation, are alternative explanations.

Debris Flows

Brazier (1987) estimated an average postglacial surface lowering of 1m or more in gullies in the Grampian Highlands as a result of debris transport. But the limited information about debris production and transport in gullies in the Scottish Highlands has been noted by Young (1975), Brazier (1987), Brazier *et al* (1988) and Ballantyne & Harris (1994).

In northern Scotland, Innes (1983a) found 780 unconfined hillslope debris flows concentrated in the last 250 of 700 years, and occurring in most decades, based on lichenometric dating. He concluded that the small numbers of debris flows before 250

deterioration in the Cairngorms, with the absence of activity during the 'Little Ice Age' particularly notable.

The paucity of data on debris flows and their causes in northern Scotland is problematic. Single site studies are insufficient to resolve the relationship between individual meteorological events, local groundwater conditions, and the prevailing climate. Equally, they can not, individually, be used to exclude a climatic trigger. On the other hand compilations such as those reviewed by Matthews *et al* (1997) and Brunsden & Ibsen (1997) are, as discussed above, prone to spurious patterns. The lack of high resolution, long-term records such as those provided elsewhere by tree-rings, currently makes the link between debris flow occurrence and climate in northern Scotland difficult either to establish or refute.

2.2.2.5   *Climate-driven landscape evolution: discussion*

Two main problems have surrounded the interpretation of climatic triggers for landscape evolution in northern Scotland. The first relates to the nature of the geomorphic data itself; the second to the existence of independent evidence for climate change against which such inferences can be calibrated.

In North America, river terrace development over wide areas has been shown to be sensitive to switches between moister and more arid climate phases, characterised by changes of temperature and precipitation of modest magnitude (e.g. Arbogast & Johnson 1993, Knox 1993). Comparable responses have been described in northern England (Macklin 1999, Macklin *et al* 1992, Macklin & Lewin 1993, Rumsby & Macklin 1996) and continental Europe (Starkel 1995a). Only one dated study of a floodplain in northern Scotland (Ballantyne & Whittington 1999) unambiguously provides evidence for enhanced fluxes of groundwater and/or sediment, which caused a switch from net aggradation to net incision at about 2700 cal BP, in the absence of any evidence for human impact.

Scotland lacks alpine zones, and has a very variable winter snow cover (Chapter 1, Section 1.2.2), but in alpine zones of northern Europe, slope evolution has been strongly linked to changes in palaeoprecipitation and palaeotemperature during the Holocene (e.g. Blikra & Nesje 1997, Blikra & Selvik 1998, Brunsden & Ibsen 1997, Grove 1972, 1981, Matthews *et al* 1997). The relationship of different types of rapid mass movement on slopes to meteorological events and well-defined ground conditions, has been

Process	Date /calBP 2σ	Author
Intense slopewash	Between 4700 and 3850	Tipping 1995
Retreat of pine woodland to its former limits	Before 4410 to 3990 (3815 ± 50^{14}C BP)	Gear & Huntley 1991

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In northern Scotland, Innes (1983a) found 780 unconfined hillslope debris flows concentrated in the last 250 of 700 years, and occurring in most decades, based on lichenometric dating. He concluded that the small numbers of debris flows before 250

years ago, demonstrated that grazing and burning which exposed regolith to erosion, was more significant than rainfall as the primary control on slope stability. But it is possible that his results are based on a multi-causal signal, which does not distinguish the relative contributions of climate, human impact and progressive weathering. Undated palaeoflows have also been attributed to the effect of intense rainstorms on sensitive sites by, for example, Ballantyne (1986b, 1997), Hinchcliffe *et al* (1997), Innes (1983b), and, in northern England, Macklin *et al* (1992). More recent evidence for debris flows in the mid and late Holocene in northern Scotland has given rise to cautious speculation about a climatic trigger (Curry 1999, Hinchcliffe 1999), with the *caveat* that extreme, possibly random, events remain the most straightforward explanation.

From Figure 2.3, fan and cone development, through new debris flows and debris redistribution, have been dated to the sixth, fifth, second, and first millennia BP. No common trigger has, however, been distinguished. Explanations given are: sediment starvation after the fifth millennium BP and the final episode of paraglacial sediment adjustment (Brazier *et al* 1988); very rare large storms (Ballantyne & Whittington 1999); and river channel undercutting of a debris-laden slope (Brazier & Ballantyne 1989). From contemporary evidence in the U.K. there is confirmation of cones forming during large, rare storms. Harvey (1986) described cone formation in response to a 'hundred year' intense, convectional storm in northwest England. In an earlier study, also in northern England, Harvey *et al* (1981) suggested intensified agricultural use in the 10th century AD as a trigger for fan formation. A climatic explanation has not been proposed, and the argument that climate change could be characterised by more frequent, and/or widespread, cone or fan formation in parallel with more frequent 'hundred year' storms, remains difficult to substantiate or refute on the basis of the handful of records currently available.

Where the authors were to discount human impact, intense precipitation and a high water table were invoked as the triggers for three gully erosion events leading to renewed late Holocene fan aggradation in the eastern Highlands (Ballantyne & Whittington 1999). In this case, sedimentological and stratigraphic evidence indicated emplacement of fan deposits in single (or very closely spaced) large events probably associated with extreme rainstorms. A climate trigger was rejected because of the lack of coincidence in timing of the debris flows with known episodes of climate

deterioration in the Cairngorms, with the absence of activity during the 'Little Ice Age' particularly notable.

The paucity of data on debris flows and their causes in northern Scotland is problematic. Single site studies are insufficient to resolve the relationship between individual meteorological events, local groundwater conditions, and the prevailing climate. Equally, they can not, individually, be used to exclude a climatic trigger. On the other hand compilations such as those reviewed by Matthews *et al* (1997) and Brunsden & Ibsen (1997) are, as discussed above, prone to spurious patterns. The lack of high resolution, long-term records such as those provided elsewhere by tree-rings, currently makes the link between debris flow occurrence and climate in northern Scotland difficult either to establish or refute.

2.2.2.5   *Climate-driven landscape evolution: discussion*

Two main problems have surrounded the interpretation of climatic triggers for landscape evolution in northern Scotland. The first relates to the nature of the geomorphic data itself; the second to the existence of independent evidence for climate change against which such inferences can be calibrated.

In North America, river terrace development over wide areas has been shown to be sensitive to switches between moister and more arid climate phases, characterised by changes of temperature and precipitation of modest magnitude (e.g. Arbogast & Johnson 1993, Knox 1993). Comparable responses have been described in northern England (Macklin 1999, Macklin *et al* 1992, Macklin & Lewin 1993, Rumsby & Macklin 1996) and continental Europe (Starkel 1995a). Only one dated study of a floodplain in northern Scotland (Ballantyne & Whittington 1999) unambiguously provides evidence for enhanced fluxes of groundwater and/or sediment, which caused a switch from net aggradation to net incision at about 2700 cal BP, in the absence of any evidence for human impact.

Scotland lacks alpine zones, and has a very variable winter snow cover (Chapter 1, Section 1.2.2), but in alpine zones of northern Europe, slope evolution has been strongly linked to changes in palaeoprecipitation and palaeotemperature during the Holocene (e.g. Blikra & Nesje 1997, Blikra & Selvik 1998, Brunsden & Ibsen 1997, Grove 1972, 1981, Matthews *et al* 1997). The relationship of different types of rapid mass movement on slopes to meteorological events and well-defined ground conditions, has been

investigated through short-term process studies designed to improve understanding of mechanisms and timescales of specific types of failure, and hence the meteorological conditions in which they occurred (Berrisford & Matthews 1997, Brunsden & Ibsen 1997, Jonasson *et al* 1997, Sandersen 1997). But attempts to match debris flows, snow avalanches and landslides to Holocene climate change on a wider European scale are error prone, due to the diversity and patchiness of the evidence (Matthews *et al* 1997; Brunsden & Ibsen 1997). Moreover, extrapolating from short term process studies and historical records (which form the bulk of evidence) to the long term geomorphological record, introduces scaling effects and large uncertainties (Brunsden 1996, Knox 1997, Thomas & Thorp 1995). These are further discussed in Chapter 3.

Large datasets show clusters of slope activity, with contrasts in timing between Southern and Northern Europe (Berrisford & Matthews 1997). They also introduce problems of interpretation. Records of rapid mass movements before historical times come from alpine zones, or in the UK, predominantly from lowland and coastal sites where they are affected by anthropogenic effects and changes in sea level respectively (Brunsden & Ibsen 1997, Berrisford & Matthews 1997). Depending on the setting, a period of increased debris flow frequency detected from a large compilation could encompass variations in snow cover or snow melt - hence colder climate, enhanced precipitation relative sea level change, or human impact. As with short term process studies, there are therefore some inherent problems in using large data sets as climate indicators.

Where climatic interpretations of slope instability are difficult to discount, the second main challenge in northern Scotland, where high resolution records such as those obtained from dendroclimatology are lacking, is to tie the evidence to independent proxy climate data. The main source of climatic evidence is palaeoecological, sub-divided into: pollen stratigraphy (Birks 1972, Birks & Matthews 1978, Pennington *et al* 1972, 1977); peat humification studies showing past changes in wetness on the surface of raised bogs (Anderson 1995, Anderson *et al* 1998, Binney 1997, Blackford 1993, Charman 1993); investigations of changes in treeline elevation (Binney 1997, Dubois & Ferguson 1985, Pears 1968); changes in stable isotope ratios in fossil wood (Dubois & Ferguson 1985); and changes in the geographical spread of pine woodland, whose survival is strongly affected by both wind exposure and ground wetness (Bennett 1995, Birks H. H. 1975, Birks H. J. B. 1977, 1989, 1996, Bridge *et al* 1990, Charman 1993,

Gear & Huntley 1991). Within these disciplines, a number of workers have emphasised the ecological insensitivities of plant communities to climate change as a further source of uncertainty (e.g. Birks 1996, Haggart & Bridge 1992, Lowe 1993).

Combining data from different sources and locations, Anderson *et al* (1998) inferred increasing wetness after 3900 cal BP, on the basis of peat humification, treeline data and changing water level in a small lake. However, the correlation of climatic data of various types from northern Scotland with wider information from the north Atlantic region may be problematic, not only because terrestrial responses in, for instance Scandinavia and Scotland are not necessarily easily compared, but because marine and terrestrial records preserve different types of data.

Long sediment cores from the eastern North Atlantic contain proxies for both sea surface temperatures and deep thermohaline ocean circulation. They indicate a pattern of rapid (centennial scale) temperature oscillations of about 2°C at one to two thousand year intervals (Section 2.2.2.2). From this evidence the 'Little Ice Age', for which historical records of climate deterioration exist, was not a unique event, but one of a series of Holocene climate shifts driven by coupled ocean-atmospheric circulation. How the landscapes of Northern Scotland responded at these times remains to be clarified. But in the 17th and 18th centuries (Little Ice Age maximum in the UK from Rumsby & Macklin 1996) there were repeated crop failures and famine due to wet cold weather, with hundreds of thousands of people reduced to destitution and begging (Lamb 1977). Severe floods, extreme temperatures, heavy snowfall, and frequent storms, became the norm (Whittington 1985). Arctic waters spread well south of their present latitude displacing fish stocks round the north coast. A village at an altitude of about 300m in the Southern Uplands was abandoned (Lamb 1995), and permanent snow was reported on mountains in Northern Scotland (Lamb 1977). Violently destructive storms affected coastal communities from the mid 16th to the mid 18th centuries (Ross 1992). Lamb (1995), quotes from the Philosophical Transactions of the Royal Society of Edinburgh 1675 reporting that *a little lake* [presumably a coire lochan] *in ...Glencannich* [Northern Highlands]... *in a bottom between the tops of a very high hill .... never wants ice on it in the middle, even in the hottest summer.*

Independent terrestrial evidence complements dendroclimatological and marine sediment core data, adding to the uncertainty of the validity of considering the Little Ice Age as 'the' Holocene 'neoglacial'. Eronen *et al* (1999) identified a period of harsh

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climatic conditions in Finland at 2500-2000 BP. It was an unfavourable time for the growth of forest-limit pine, and in that sense may have been a more severe climatic disturbance than the "Little Ice Age". Glacier fluctuation records in Iceland (Gudmundsson 1997) are not now believed to contain a simple climate signal, and are better understood as a complex response to several interacting factors that operate at different timescales. As a result, it has been suggested that the use of the term 'Little Ice Age' should be restricted to describing a period of extended glacier cover, rather than being used to define a period with specific climate conditions.

Even this usage may be problematic. The 'Little Ice Age' has been described as a complex global event, which varied in importance geographically, encompassing both warm and cold climatic anomalies (e.g. Bradley & Jones 1993, Briffa *et al* 1990, Grove 1981, Lamb 1977, Oppo 1997). Fluvial systems in northern England have been interpreted as responsive to shifts in hydroclimate on timescales of 10-30 years (Rumsby & Macklin 1996), at a time when fluctuations of temperature and precipitation in northwest Europe, within the pattern of generally colder weather, were high (McCarroll & Matthews 1997). A high degree of short-term climate variability of this kind has been recorded in ice core, tree-ring and North Atlantic sediment records at the boundaries between episodes of major abrupt, climate change (Alley *et al* 1997, Barnett *et al* 1996, Bond *et al* 1997, Briffa *et al* 1990, Dansgaard *et al* 1993, Ely *et al* 1993, Taylor et al 1993).

Evidence for repeated Holocene climate fluctuations in NW Europe raises the question of whether climate has influenced landscape evolution to an extent unsuspected when the Holocene climate of northern Europe was thought of as extremely stable until 'the Little Ice Age' (Dansgaard *et al* 1993). But evidence for a response to Holocene climate variability in northern Scotland is currently weak. Clear peaks of climate deterioration which match rapid cold water incursions with peak fluxes in the North Atlantic and Nordic Seas at about 5900, 4200, 2800, and 1400 cal BP, have not been described. Instead, an increase in precipitation after about 6000 BP with intensification after about 3000 BP has been inferred, mainly from the palaeoecological record. There is evidence for blanket peat expansion coinciding with wetter conditions, as signalled by plant assemblages, in parts of the Northern and Grampian Highlands after about 6000 BP (Bridge *et al* 1990, Bennett K.D. 1995, Birks 1996 H. J. B., Pennington *et al* 1972). But soil paludification could be due to factors other than increased precipitation (Birks H. J.

B. 1996, Huntley 1996). Radiocarbon dated stumps older than 5000 years have been found at higher altitudes than trees living today (Dubois and Ferguson 1986, Haggart and Bridge 1992, Pears 1968). A higher treeline suggests less wind stress and waterlogging, as does a fluctuating areal distribution of pine woodland, although human impact and soil evolution have also played a part (Bennett 1995 K. D., Birks 1996, 1997, Gear and Huntley 1991, Lowe 1991, 1993, Pennington *et al* 1972). A further humid event at about 3900 cal BP was inferred by Anderson *et al* (1998).

One Scottish study which suggested four Holocene 'pluvials' (Dubois & Ferguson (1985), is much quoted, possibly because it appears to have a quantitative rigour. These authors inferred elevated precipitation from the oxygen and deuterium isotope ratios in radiocarbon dated tree stumps buried in peat in the Cairngorms. On that basis, four 'pluvials' at about 7300BP, 6200-5800BP, 4200-3940BP and 3300BP were identified. However, Buhay *et al* (1996), Loader & Switzur (1995), Pears (1988), Saurer *et al* (1998) have questioned the reliability of $^{18}O/^{16}O$ ratios in trees as palaeoprecipitation proxies and emphasised the very complex environmental variables affecting oxygen fractionation in wood. Studies of contemporary trees have shown that major variables include tree age and growth rate as well as temperature and moisture fluctuations in air and soil. Although Dubois & Ferguson's 'pluvials' have been used as a standard against which to correlate other palaeoclimate proxies (e.g. Anderson *et al* 1998, Binney 1997, Bridge *et al* 1992), because of complex isotope fractionation in growing wood, palaeo $^{18}O/^{16}O$ ratios in modern tree trunks are now considered to be multi-proxy environmental indicators in which the moisture signal, as opposed to temperature and biochemical controls on isotope fractionation, is neither easily identified, nor easily correlated with patterns of precipitation. Applying the technique to sub-fossil wood, introduces very large uncertainties and appears to be unreliable.

In Scotland there is no strong evidence for high levels of precipitation associated with a southern incursion of the Faroes-Iceland Front and cold ISOW at about 4700 cal BP (Bianchi & McCave 1999) which was contemporary with general glacier advance in Iceland (Gudmundsson 1997) and an increased frequency of snow-transported debris avalanches in western Norway (Blikra & Nemec 1998). On the contrary the timing of the marked climatic amelioration inferred from Gear & Huntley's (1991) short-lived but well-documented pine woodland expansion some 80 km north and west of its normal range, which took place between about 5310 - 4850 and 4410 3990 cal BP (4450+65 -

3815 +50 ^{14}C BP) coincides with this ocean cooling event. However, pine is sensitive to water logging and strong winds rather than temperature (within the range applicable in Northern Scotland) (Bennett 1995), illustrating the need for careful use of the terms 'deterioration' and 'amelioration' in respect of Holocene climate.

Many factors potentially complicate identification and interpretation of terrestrial response to marine cooling, including regional variation due to the influence of the North and Baltic seas, different types of response to the same event as measured in glaciated and non-glaciated areas of NW Europe, lag times between marine conditions and terrestrial impacts, and co-variables such as climate and weathering rates. Different patterns of climate change in regions dominated by summer or winter rainfall may also contribute to variability (Yu and Harrison 1995, Rumsby & Macklin 1996). If present day summer/winter contrasts in rainfall pattern have applied throughout the Holocene, evidence of high levels of run-off resulting in shallow slope failures and perhaps river channel avulsion, may best reflect times when atmospheric circulation was dominated by easterly, continental air masses which produce intense, convectional storms. Continuous rainfall records from Aviemore on the western margin of the Cairngorm show that convective storms generated under continental conditions are of shorter duration and higher intensity than frontal storms generated under maritime conditions. Convective storms also release about 75% of their rainfall in the first half of the storm, while frontal storms have a more even distribution of rain throughout (Brooks 1997). Convective storms are therefore more likely to generate high levels of run-off and shallow slope failure.

In a palaeohydrological study of several catchments in the eastern Cairngorms, Maizels & Aitken (1991) found strong evidence of adjustment to postglacial conditions, but no buried organics which could provide evidence of sensitivity to Holocene climate. But Werritty & Ferguson (1980), working on braided sections of the River Feshie on the margin of the western Cairngorm massif concluded that major channel changes over the last 200 years were in accordance with Wolman and Miller's (1960) dictum that most geomorphic work is done by events of modest magnitude and fairly high frequency. Wider application of Werritty and Ferguson's findings is uncertain because braided rivers are exceptional in Britain today, although braided reaches can develop temporarily in response to intense storms, before reverting to the meandering systems which are in equilibrium with today's conditions (Harvey 1986). But in the light of more

recent understanding of the high frequency, modest-scale, variability of the Holocene climate of northern Europe discussed above, it is worth noting that Werritty & Ferguson found fluvial reorganisation linked to low magnitude climate change.

Additional sources of evidence used to infer periods of elevated precipitation in northern Scotland are sediment chemistry (Pennington *et al* 1972, Edwards & Rowntree 1980) and uneven sediment accumulation rates (Bain *et al* 1990, Ballantyne & Whittington 1987, Edwards & Rowntree 1980, Innes 1983a, b, Tipping 1995). But because no long dendrochronological or dendroclimatological sequence for northern Scotland has been compiled, comparisons have sometimes been based on co-incidence of events in Scotland with those in Scandinavia.

The status of lake level changes in Scotland as palaeoclimate indicators, and the validity of extrapolating better-studied changes in Scandinavia to northern Scotland, are uncertain (Harrison *et al* 1991, Harrison & Digerfeldt 1993, Mannion 1982, Smith 1996, Yu & Harrison 1995). Dated Scottish studies are few in number. Cessation of sedimentation between 5650±80 and 3250±60 radiocarbon years BP at the margin of a small (100m diameter) loch in northern Scotland with no inflow point, was interpreted as the result of a sustained drop in water level (Smith 1996). The 2σ calibrated age range for this event (6670-6300 to 3630-3370 cal BP), places the start of this drier period in the previous millennium. That is in contradiction to the palaeoecological data from northern Scotland which suggest increasing soil wetness at this time (Bennett K. D. 1995, Birks H. H. 1972, Birks H. J. B. 1977, 1996, Bridge *et al* 1990, Gear & Huntley 1991, Lowe 1993, Pennington 1995, Pennington *et al* 1972). That soil wetness can however, be interpreted as the product of leaching and progressive paludification (Birks 1996, Huntley 1996) rather than the effect of increased rainfall.

In respect of northern Europe as a whole, current lake status data has been considered an insufficient and regionally variable basis for palaeoclimate reconstruction (Harrison & Digerfeldt 1993). Asynchroneity of records between southern Sweden and Scotland has been noted by these authors and by Yu and Harrison (1995).

Conditions in four Scottish lakes, three in the Grampian Highlands, Loch Piltyoulish, Loch a' Chnuic and Loch Garten, together with Linton Loch in south-east Scotland are reported as indicating conditions drier than at present in Scotland at about 6000 radiocarbon years BP (Harrison *et al* 1991, Fig. 1). Unlike most of northwest Europe, these lakes show no divergence from today's levels at c. 3000BP. At least one,

additional lake in the northwest Highlands was incorporated in a follow up study (Yu and Harrison 1995) although no site list is provided. No change in lake level is indicated in their summary diagrams (Fig. 3a and 3b) except for drier conditions at 7000BP in the Grampian Highlands. Their data for Scotland, plus Smith's (1996) record, are summarised in Table 2.4 below. One of the lakes in question in southeast Scotland was investigated by Mannion (1982), who did not directly date the timing of the hiatus in sedimentation ($6528\pm100 - 5341\pm70$BP; 7540 - 5920 cal BP), but inferred it from correlation of pollen stratigraphy with a dated peat column from a nearby peat bog.

Table 2.4, column 4, shows that, based on the number of records of each type, there is little divergence in lake status from that current today, except possibly at about 7000 and 5000 BP when there is some indication of conditions drier than those of today. It is unclear whether this could be due to complexities of response in settings with different variables, or a real lack of change.

^{14}C date BP	Wetter than today	Drier than today	Similar to today
1000	0	0	5
2000	0	1	3
3000	1	0	3
4000	0	1[+:Smith 1996]	5
5000	1	2[+:Smith 1996]	3
6000	1	1	3
7000	1	3	2

TABLE 2.4
Holocene lake level status in Scotland. Derived from Yu & Harrison (1995) Figure 3. Numbers refer to the number of lakes in each category.

Interpretation is complicated by the fact that both wetter shifts and evidence for no change of status occur during the same time intervals. But the records are few in number, subject to unconstrained variables and sources of error. Interpretation is therefore difficult. The varied nature of the lakes, and the limited number of sites, means that climatic triggers are not clearly dissociated from other environmental triggers (Harrison & Digerfeldt 1993, Vassiljev 1998, Prentice *et al* 1998). Lake levels

are therefore an ambiguous climate proxy in respect of northern Scotland, and as with isotope data in subfossil trees, an uncertain database against which to assess the timing of landscape evolution.

In conclusion, a clear chronology of Holocene climate change in northern Scotland has not yet been established, although annual average temperature changes of up to 2°C have been estimated to have affected northwest Europe on a number of occasions since the last Stadial; nor has the impact of climate on landscape evolution been securely demonstrated.

2.2.3   Human impact on landscape evolution

The literature for northern Scotland contains little unambiguous evidence for impacts of vegetation clearance on groundwater and soil moisture conditions, hence on sediment transport on slopes before recent historical times. Dated investigations where there is good evidence for human activity contemporary with slope sediment erosion include Brazier *et al* (1988), Edwards & Rowntree 1980, Huntley (1994), and Tipping (1995). However the mechanisms which might causally relate slope destabilisation and shrub/woodland clearance by burning and/or grazing, are difficult to confirm. Tree and shrub removal can reduce the pumping effect of evapotranspiration and allow the water table to rise (e.g. Greenway 1987). But in the cool, temperate, oceanic climate of upland Scotland at about 57°N, the stabilising effects of trees or shrubby vegetation on slope soils and sediments will be least effective precisely when erosion is most likely - that is, during intense winter storms or rapid winter/spring snow melt. At these times, evaporative pumping and water table control by ground-covering plants will be negligible relative to the transient water budget, due to the long autumn/winter/spring dormant season which lengthens with altitude. In these circumstances, the distribution of intense rainfall between winter and summer - which may itself have been subject to secular variation during the Holocene (Rumsby & Macklin 1996) - may be critical in determining whether tree clearance, such as that described by Brazier *et al* (1988; Section 2.1.7), could have affected slope erosion.

Anthropogenic impact has been invoked to explain other types of landform change. In the uplands of northern England, on the basis of an association between the timing of instability and the timing of invasions of Viking pastoralists, overgrazing was suggested as the explanation for gullying and alluvial fan formation (Harvey 1996, Harvey *et al* 1981, Harvey & Renwick 1987). Grazing pressures were also suggested as the cause of

the recent enhanced frequency of debris flows on slopes at sites in Highland Scotland (Innes 1983a), and of erosion of high altitude aeolian sediment (Ballantyne & Whittington 1987). Co-incidence in the timing of human activity and slope instability is suggestive, but alternative explanations have not yet been excluded.

Other palaeoenvironmental studies have highlighted the difficulty of establishing a causal link (e.g. Tipping (1995), Section 2.2.2.4). Edwards and Rowntree (1980) argued that forest clearance by early inhabitants had a significant effect on rates of erosion on the slopes adjacent to a small sediment sink (loch). However, they acknowledged that this conclusion masked difficulties of interpreting and integrating multi-proxy geochemical, sedimentological and pollen data. For instance, during the late Bronze to Iron Age, deep erosion as indicated by reversed radiocarbon dates and the carbon content of loch sediment, was not matched by an expected significant increase in deposition rates. However, in sampling a sediment sink such as a loch, which receives its input from a whole catchment, it is difficult to distinguish the effects of restricted erosion in one or more gullies, perhaps related to drainage activities, from a more general increase in slope surface sediment instability triggered by climate or agricultural practice. In the same study, evidence for greatly accelerated erosion in the last few hundred years (from an indeterminate radiocarbon date) was less ambiguous, and was attributed to agricultural activity, because of sudden deposition of a 5cm thick sand layer perhaps due to cutting of drainage ditches. On the other hand, in two studies from northern Scotland, investigators have specifically excluded anthropogenic influence. At one of these sites, sediment and run-off source areas were unvegetated (Innes 1983b). At the other, pollen stratigraphy indicated no correlation between slope instability and vegetation change (Ballantyne & Whittington 1999).

While there is great deal of evidence in the Scottish Highlands of ecological damage to tree and shrub cover through overgrazing associated with large increases in numbers of deer and sheep since the mid 18th century, process studies demonstrating consequent enhanced slopewash and/or slope mass movements are restricted to artificial woodland monocultures (Section 2.1.6 above). 'Hagged' blanket peats are vertically incised to the substrate in irregular patterns. But from field observations, it is unclear whether flow in subsurface depressions, or incision initiated at the surface by wind, water or herbivore impacts, is responsible. In a review of the extent of upland peat erosion throughout Scotland, Grieve *et al* (1994) found that half the areas mapped as blanket peat are

currently subject to erosion. They drew no conclusions about the causes. From aerial survey, the percentage of peat affected by erosion within each mapped 10km^2 area ranged from 20% to less than 0.5%. It follows that more than half the blanket peat cover is currently unaffected, or little affected, by erosion, despite universal anthropogenic impact, and severe ecological degradation. The extent of erosion was also found to increase with altitude. Given the rapid increase of exposure and precipitation with altitude (Chapter 1, Section 1.2.1), this trend is more obviously explained as the interaction of topography with weather and climate, than with grazing pressures, although the impact of overgrazing may be enhanced in climatically sensitive zones. Blanket peat, removed by whatever agency, exposes till which is often unconsolidated and prone to erosion. That in turn produces sand and silt-laden run-off which, because such run-off is many times more effective at removing peat than clear water (Grieve *et al* 1994), creates a positive system feedback, resulting in further peat degradation, and more exposure of the potentially unstable mineral substrate. No comparable study has yet been done on mineral soils.

Deforestation was proposed by Maizels & Aitken (1991) as the cause of '4m of post-5000BP alluviation the eastern Cairngorms, which interrupted well-established postglacial stability'. Process studies which could confirm this are restricted to sediment budgets in commercial plantations. Sediment yields attributable to planting and clear-felling range from 120 - 600% of pre-existing yields, but have only a short-term effect, with a rapid reduction in transport of bedload and suspended sediment within ten years once trees are well established (Johnson 1993). Run-off and soil loss on a range of soil types associated with ground clearance and ploughing have been found to be directly proportional to furrow length and rainfall intensity (Carling *et al* 1993). Above the gentle angles at which deposition predominated, slope steepness was not found to be a major control on erosion. Extrapolation from these observations to the behaviour of natural vegetation and slopes would be dubious (Section 2.1.6), since vegetation loss in linear channels at high angles to slope contours in forestry plantations is dissimilar to the effects of intensive grazing, animal tracks, burning, and tree clearance, in which vegetation forms a mosaic rather than dense monocultures. Data on continuing, or long term, sediment budgets if trees were replaced by other vegetation, are not available.

Proof that anthropogenic impact in the mid and late Holocene, before the advent of mechanisation, has been the primary trigger of slope instability and landform change in

northern Scotland, therefore remains elusive. The relationship between anthropogenic impact and erosion on natural slopes is suggested by a small number of studies where an association has been shown. It has been excluded as a trigger in a small number of additional studies of late Holocene landforms; and remains one of a number of possible explanations in others. The situation in northern Scotland exemplifies a wider northern European problem commented on by Matthews *et al* (1997) who concluded that information about human impact on postglacial landforms derives mainly from a body of research based on lowland sites which has been little tested in mountain environments.

2.3   DISCUSSION: THE CURRENT EVENT STRATIGRAPHY

The status of the information in Figure 2.3 as an event stratigraphy (defined in Section 2.0) depends on the size and scope of the database, the precision of the data, and whether external triggers (random and systematic) can be differentiated. These are the topics of the first three sub-sections below. The fourth sub-section addresses the robustness of current evidence for non-random landscape evolution.

2.3.1 The size and scope of the database

Figure 2.3 and Table 2.1 list sixteen published papers with 'absolute' dates for landform changes with a wide geographical spread. Only one study (Innes 1983a) sampled and compared the timing of comparable events at geographically dispersed sites. In that instance, the dating methodology (lichenometry) restricted results to the last few hundred years.

Three papers date slow mass movement (solifluction); three date periods of enhanced rates of erosion (slopewash) on slopes; eleven date rapid mass movement of four distinct types; and two of the eleven also assign dates to fluvial incision. No one class of process or landform is well represented. Only in the last few hundred years does one type of rapid mass movement - hillslope debris flows - dominate the results, but the methodology skews the overall picture of debris flow frequency in favour of recent times. On the whole, the database is a collection of unrelated, single-site studies which allow few direct comparisons. It suggests, but provides no conclusive evidence for, both some clustering of events, and for lengthy periods of stability interspersed with episodic activity.

2.3.2 The precision of the database

Not all the dates in Figure 2.3 are equally robust, and some have large error bars. Imprecision is exacerbated by the fact that some events and processes (identified by a question mark) are only dated as occurring after a given date. The reasons for lack of precision noted in Sections 2.1 and 2.2 include: time gaps (uncertain or undeterminable) between the age of the dated organic material and deposition of the overlying mineral horizon; the precision of dates obtained from buried organic soils as a result of carbon translocation, contamination and uncertain residence times; limitations of the precision of dating methods themselves (radiocarbon, cosmogenic chlorine, lichenometric); and difficulties of interpreting and integrating multi-proxy signals for erosion rates. Uncertainty about timing is increased by the fact that peat dates in the studies quoted have been obtained from homogenised slices ranging from 0.5 - 4.0 cm in thickness.

2.3.3 Evidence for cause and effect: environmental triggers

The causes of landform evolution which have been described in detail in northern Scotland remain open to interpretation (Section 2.2). Complex responses, lags and other environmental controls may complicate system reactions to climate change so that they may or may not reflect the rate, magnitude and variability of climate (Phillips 1999). The interaction of anthropogenic and climate impacts further complicates arguments which propose either as the principal trigger for landscape evolution. The lack of quantitative or semi-quantitative estimates of the effect of vegetation clearance on sediment and water budgets on natural slopes in mountain areas of the Highlands compounds that difficulty. 'Coeval' cannot be extended to mean 'causal'. Moreover, it is not yet possible to define 'coeval' accurately in terms of climate in Northern Scotland since the magnitude, scaling, duration, modes and timing of climate changes are only broadly known. Inferred positive correlation between climate change and integrated groups of indicators of palaeoenvironmental change (e.g. Anderson *et al* 1998, Haggart & Bridge 1992), are therefore open to review because of uncertainty about the timing and duration of climate change, as well as the unsatisfactory or ambiguous nature of some climate proxies.

Climate indicators such as those from Northeast Atlantic sediment cores which record southern influxes of Arctic surface water, and climate models for earlier stages of the Holocene which predict changes in temperature and precipitation, are not unambiguously reflected in the terrestrial data from northern Scotland. Unfortunately,

no dendroclimatological sequence for the region exists. So where hydrological conditions free from human impact can reasonably be inferred as the immediate trigger for landform evolution, random, rare meteorological events, interacting with intrinsic factors, and/or human impact, have been proposed as the best explanation (e.g. Ballantyne 1997, Innes 1997). This does not, however, exclude a direct relationship between the climate and the frequency of erosive precipitation.

2.3.4 Non-random landform evolution

The first requirement of a synthesis is that data be comparable. Comparison of calibrated dates from studies which predate CALIB 3.O (Stuiver & Reimer 1993) with those calibrated afterwards, can produce variations amounting to hundreds of years, and therefore spurious correlations. Chronologies based on uncalibrated dates pose a similar problem. Attempts to match chronologies and events in northern Scotland with those elsewhere in northern Europe are hampered by the extreme oceanicity of Scotland's climate. Contrasts of climate within the wider setting of the British Isles and NW Europe further complicate interpretation.

The term *event stratigraphy* was defined in Section 2.0 as an accurate chronostratigraphic scale for the occurrence of particular events or sets of events. Since the nature and extent of the database in Figure 2.3 is partial, any conclusions can only be tentative and are open to revision. However, some tests can be applied to separate random and systematic variation.

- If random meteorological events, together with intrinsic, progressive change specific to individual sites, have been the dominant controls on Holocene landscape evolution, no systematic patterns or trends should be evident in an event stratigraphy.

- If paraglacial relaxation has been the dominant context for landscape evolution, an absolute (exponential?) decrease in activity from the early mid-Holocene onwards should be evident at sites where anthropogenic factors can be considered absent or minimal.

- If climate change has been a significant external forcing factor, clusters of events which match the timing of known climate episodes should be discernible.

- If human impact has significantly affected landscape stability, trends should emerge which show a positive correlation with evidence for human settlement, negative

correlation with sites and processes little affected by human impact, and a marked increase in landscape instability in parallel with the impacts of pastoralism and agriculture.

According to the model which argues for paraglacial relaxation as the continuing driver of mid and late Holocene landscape evolution, the timing of events depends on intrinsic local response and should therefore show a random pattern through time. If non-random patterns were to be confirmed, this could be interpreted as weakening the case for the continuing importance of paraglacial relaxation into the mid and late Holocene. Currently, none of the above hypotheses can be adequately tested. While the distribution of dates in Figure 2.3 generally has a random appearance, with no millennium except the earliest after deglaciation unrepresented, the size of the error bars on dates means that several records could be shifted to other millennial groupings. For instance, there may be only one record for the earliest part of the third millennium BP, leaving that period of time largely 'event-free'. Similarly, two out of three records in the 2000-1000 cal BP time interval may belong to the subsequent group.

Figure 2.3 can be interpreted in a number of ways if the data are considered to be sufficiently precise and complete to justify this. For instance, the most active periods were shortly after 6000 cal BP (5 types of process); shortly before 2000 cal BP (5 types of process); at about 600 cal BP (5 types of process); and at about 4500 cal BP (4 types of process). While these active periods are widely separated in time, they cannot firmly be interpreted as evidence of punctuated geomorphic activity, since the scale of the evidence for widespread slope instability with a variety of different expressions could conceal both random and purely locally significant events. Nor is it clear what relative significance, in terms of punctuated landscape evolution, should be given to the simultaneous occurrence of different types of slope instability, or indeed more general landform evolution, including floodplain transformations, as opposed to large numbers of a single type of event, for instance, hillslope debris flows in the last few hundred years.

An event stratigraphy based on 1σ rather than 2σ probabilities would show 'clusters' of events and periods of 'quiescence' - that is, episodicity - not visible in Figure 2.3. A further complicating factor is the argument for the dataset being skewed towards the most recent events because of better preservation and exposure. While clustering is not clearly established, further investigation could, with equal probability, validate clusters

and randomness.

The best evidence of episodicity in Figure 2.3 as it stands, is that more records appear in the fifth and first millennia BP than at other times. Against this, the partial nature of the data makes it difficult to exclude other reasons for the apparent cluster between 6000 and 5000 cal BP whose first entry could be included in the previous millennium. But the signal contained in apparent clusters in these two millennia remains ambiguous. It could reflect three things: random variation in the distribution of events through time, and/or an imbalance in sampling, and/or the influence of an external trigger (or combination of triggers) whose impact has been larger than the combined effect of random meteorological events and progressive intrinsic processes. Possible external triggers are climate change and human impact, the latter expressed mainly through the effects of grazing and burning on land surface stability. Both the unquantified nature of human impact on landform stability, and the co-variability of these forcing factors if measured *via* pollen stratigraphy or other indicators of vegetation change, have been discussed above. Spurious identification of intervals when the landscape was quiescent is equally difficult to exclude.

The stratigraphy of single sites listed in Figure 2.3 provides strong evidence for punctuated geomorphic activity, and irregular rates of erosion and deposition. But single site studies can only provide information about local conditions. They cannot by themselves discriminate between local, intrinsic triggers and external forcing factors even if the timing of events coincides with independent data. This does not change when a number of events at a range of individually investigated site is collated, as in Figure 2.3. The synthesis cannot provide unambiguous information about the influence or absence of external forcing factors even where there are apparent clusters and gaps, since it represents an unknown combination of signals. However, there could come a point when the weight of evidence made such a conclusion tenable or highly likely.

2.4   CONCLUSIONS

The literature demonstrates that the problems of investigating landform evolution and its triggers in northern Scotland are substantial. But landscape evolution has been convincingly linked to Holocene environmental change in many parts of the world, including parts of Northwest Europe. If the mountain landforms of Northern Scotland have been exempt from such impact, the reasons for it, and future consequences arising

from that fact, are surely as interesting as those underpinning system changes.

There is evidence from the Northern Highlands that

- Slopes have been active in the mid and late Holocene after lengthy periods of stability.

- Intense rainstorms (?or snowmelt) can trigger slope instability.

- Granular weathering of rock faces during the Holocene has been of sufficient magnitude to cause abrupt change in slope sediment systems.

- Human activity has coincided with periods of enhanced slope erosion.

On the other hand, it is clear that existing data are partial, and may be unrepresentative in respect of both the frequency of different types of mass movement and its timing.

At the same time, there is evidence that

- The Holocene climate of the North Atlantic region has been variable on timescales of millennia, centuries and decades, with sometimes rapid oscillations.

- Landscape evolution in parts of Northern Europe and North America has been triggered by low amplitude Holocene climate change.

The question then arises of whether mountain landscapes in northern Scotland may also have been responsive to climate change and/or human impact. However, no clear chronology of climate change has been verified, the precise nature and extent of human impact both today and in the past is poorly understood, and the current event stratigraphy is inadequate in scope and detail. Possible lag times between climate change and landform response add to the problem of defining any coincidence of timing, and inferring a causal relationship. Although the Highlands constitute a large proportion of upland Britain, limited progress has been made on understanding whether Holocene landscape evolution has been widespread, and the ages and processes of development of many landforms and surfaces are enigmatic.

The aims and specific objectives of this thesis arise directly from the literature review. and are set out in Chapter 3.
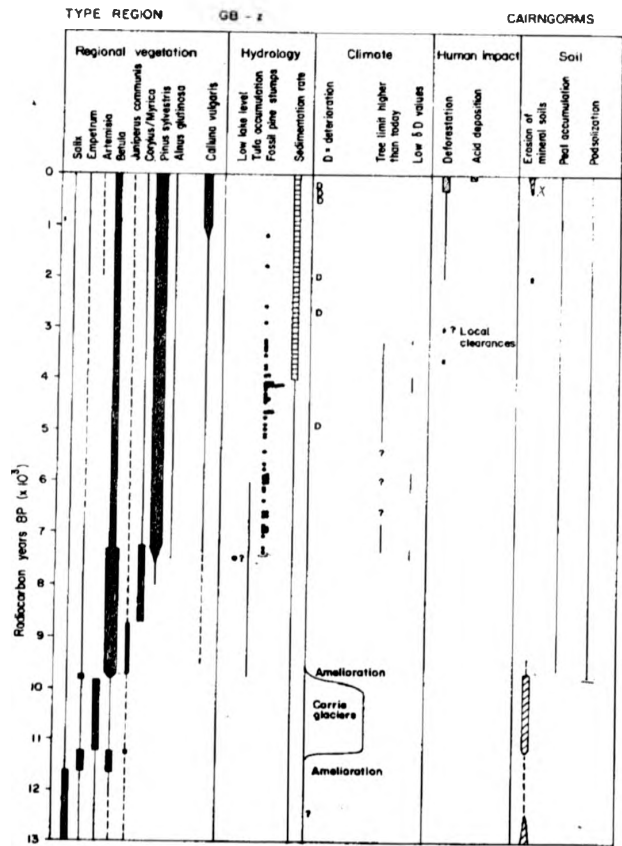
FIGURE 2.1
An Event Stratigraphy for the Cairngorms
(Birks 1996, Fig. 4.12)

60
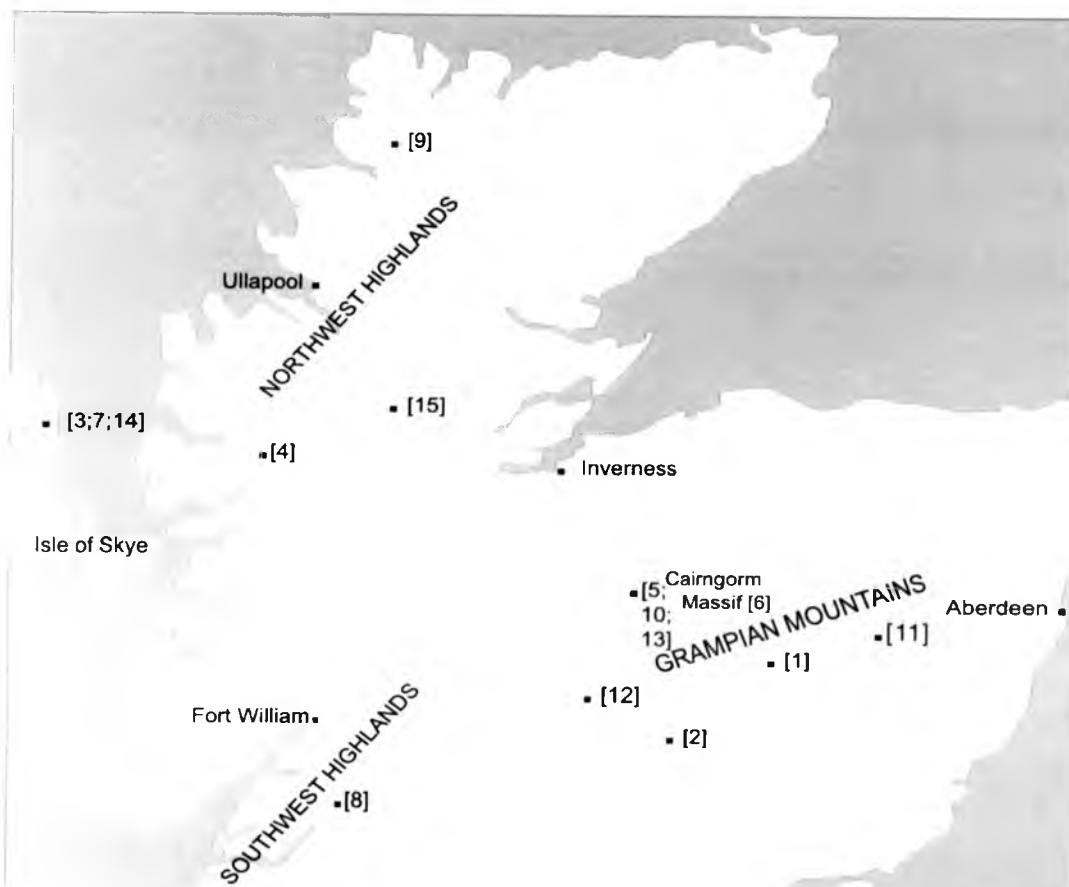
FIGURE 2.2
Locations of dated sites recorded in the literature.
Numbers in brackets correspond to those in Table 2.1 and Figure 2.3

1	Coire Fee	(Huntley 1981)
2	Carn Dubh	(Tipping 1995)
3	Trotternish	(Ballantyne *et al* 1988)
4	Glen Docherty	(Curry 1999)
5	Glen Feshie	(Bain *et al* 1993)
6	Cairngorms	(Sugden 1971)
7	Trotternish	(Hinchcliffe 1999)
8	Glen Etive	(Brazier *et al* 1988)
9	Arkle	(Mottershead 1978)
10	Glen Feshie	(Robertson-Rintoul 1986)
11	Braeraddoch Loch	(Edwards & Rowntree 1980)
12	Edendon Valley	(Ballantyne & Whittington 1999)
13	Glen Feshie	(Brazier & Ballantyne 1989)
14	Trotternish	(Innes 1983b)
15	Fannich Mountains	(Ballantyne 1986a)
16	SW Highlands, NW Highlands, Cairngorms	(Innes 1983a)

FIGURE 2.3
Dated Holocene geomorphic events in Northern Scotland
Numbers in brackets correspond to locations inTable 2.1 and Figure 2.2
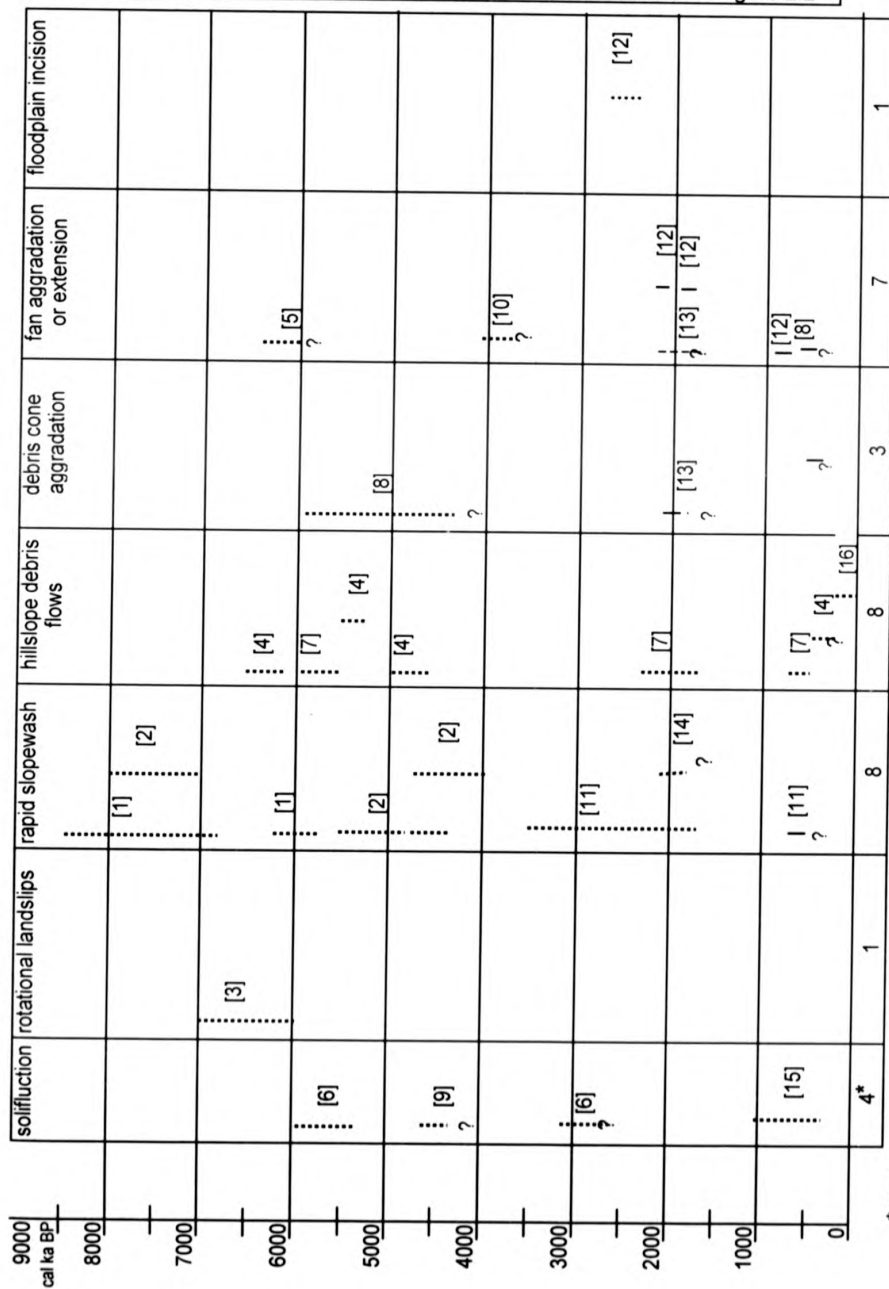
* Number of reported dates. The number of *events or processes* ranges from one to (exceptionally) hundreds. See text for comment.

? indicates that the event or process occurred at an unspecified time *after* the range of $2\_$ calibrated ages shown by the error bar

CHAPTER 3: PROJECT DESIGN AND METHODOLOGIES

3.0 AIMS AND OBJECTIVES

The aims of this thesis arise from the literature review. They are: to identify sites in Northern Scotland where landforms have been transformed during the Holocene; to generate new data about the nature and timing of the processes responsible for change; and to investigate, if possible, if and why these landscapes have been sensitive to change.

Specific objectives are to

1. Identify sites where exposures of Holocene stratigraphy allow the evolution of typical Northern Scottish landforms to be investigated (Chapter 4)

2. Describe and date these changes (Chapters 4, 5 & 6)

3. Identify any changes in the frequency and nature of the events which reshaped landforms (Chapters 5 & 6)

4. Seek to include and/or exclude triggers of system change (Chapter 7)

5. Add to the current event stratigraphy described in Chapter 2

6. Critically evaluate existing models of landscape evolution in mountain areas of Northern Scotland in the light of these new data (Chapter 8)

7. Propose an alternative model if justified by new data (Chapter 8).

3.1 PROJECT DESIGN

The combination of sparse reports in the literature of buried Holocene organic deposits, and the prevalence of coarse sediment and steep slopes in Highland catchments, meant that finding accessible exposures of Holocene stratigraphy with datable stratigraphies was not a foregone conclusion. An initial plan to match sites closely to previously dated pollen stratigraphic sites, in order to compare results with an independent dataset, proved impracticable for two reasons. Pollen sites are located where there has been long term stability, while preservation of organic horizons on slopes requires periodic instability. Secondly, exposure of buried slope organics was too infrequent for systematic matching. The first task was therefore to identify suitable locations. The second was to establish whether the locations identified embodied landforms which

were typical or unique, and widespread or rare, since unusual or rare sites would invalidate generalisation.

Major variables considered in site identification included catchment size, slope length and aspect, present-day precipitation, lithology, and geological and glacial history. However, since the investigation was at the scale of landforms, not whole catchments, the nature of slope sediment transport and storage was the key issue. These processes reflect geological structures as well as the local history of glacial erosion and deposition - features which could be investigated using aerial photographs and field reconnaissance.

Sites suitable for detailed investigation were, in principle, defined as common and characteristic lower slope sediment traps, linked to sediment sources, and without large scale interim storage. Complicating factors which project design sought to avoid, were slopes with complex sediment traps, and sources of sediment and groundwater susceptible to anthropogenic impacts. This suggested that productive sites for investigation would be landforms such as cones and fans and their feeder gullies, sourced in upper slope bedrock. Fans incised by their own streams and/or river erosion might produce suitable exposures, and could be identified by using aerial photographs and field survey.

The results of site survey and selection are described in Chapter 4. Detailed investigations at representative sites are reported in Chapters 5 and 6. The data generated are assessed in Chapter 7 in the light of a) the current event stratigraphy and understanding of causal processes in Northern Scotland, and b) wider data from north-west Europe and the North Atlantic area.

3.1.1 Scale Effects

The timescale on which stratigraphy is investigated influences conclusions about the frequency, duration and significance of events. In most catchments, long term rates of sedimentation are several orders of magnitude lower than short to medium term rates (Kirkby 1987, Knox 1997, Thomas & Thorp 1995, Thomas 2000). Moreover, it is not appropriate to use data from short term process studies in models of complex open systems, with response times orders of magnitude greater than the processes quantified (Brunsden 1996, Sugden *et al* 1997). Conversely, geomorphic events which cause

irreversible change are much more episodic and/or variable than would appear from short-term records in which their inclusion may be a matter of chance.

The search for precision and predictability can lead to a tension between methodology and aims. The goal of short-term process-response models is to define causal relationships between geomorphological events and their immediate triggers. But on the one hand, slope failures respond unpredictably in the short term, while on the other hand, the long term geomorphological record is a record only of persistent events, some of which may be rare. This project does not attempt a full specification of boundary conditions and thresholds of system change. These factors, combined with the limitations of radiocarbon dating (Section 3.2.2 below), and the interacting short, medium and long timescales on which processes may operate, therefore dictated that viable sites were those where sensitivity to changes in Holocene system threshold conditions could be studied on timescales of hundreds to thousands of years.

A regional (that is, Northern Scotland) transect of sites whose time resolution could capture long term trends and episodicity, would have demanded a large number of radiocarbon dates. NERC Radiocarbon Dating Allocation No. 660/0896 allowed for thirty-six dates at two representative localities (3.2.2 below). To supplement this, the prevalence and stratigraphy of Holocene landscape evolution was confirmed through description of sites in four additional catchments (Chapter 4).

3.2 METHODS

Processes which had contributed to palaeoenvironmental change at the sites identified were investigated by aerial survey, field mapping, stratigraphic logs, sediment analysis and micromorphology. The timing and frequency of events was established through radiocarbon dating.

3.2.1 Aerial photography and field survey

Aerial photographs, in stereo pairs at a scale of approximately 1:24,000 (Scottish Office Central Research Unit Survey, 1988/1989), were made available by Scottish Natural Heritage, Edinburgh. They were obtained between 14/5/88 and 19/10/89, and viewed using a Topcon Tokyo Kogaku Kikai K.K. stereoscope, and were used to:

- complement and suggest locations for ground surveys
- identify landforms which post-dated ice-contact landforms

- identify sites where fluvial or slope incision suggested that Holocene stratigraphy might be exposed

- identify linked sediment sources and traps - specifically, gullies originating in upper slopes with fans/cones at their bases

- differentiate gully types

- relate gullies and sediment transport to geological structures

- ensure that sites chosen for detailed investigation were neither unusual nor unique in form or setting

- map a gully whose formation post-dated OS maps

- produce location maps.

The identification numbers of the photographs, from Lines 70 -74, and 76 are listed below. Locations are identified in Chapter 4, Figure 4.1.

Glen Affric/Gleann a' Choillich	12 89 189-199
Glen Cannich	14 89 035-8
Glen Carron	26 88 171-172 and 208-210
Gleann Chaorainn/Glen Orrin	26 88 162-165 and 219-26 and 266-268
Glen Elchaig/Srath Duilleach	12 89 241-242
	14 89 021-026
Gleann Fhiodhaig	26 88 166-172
Gleann Lichd/Glen Shiel	41 88 151-153 and 207--210
Glen Ling	14 89 105-110 and 241-243
	30 89 046-048
Glen Markie	41 88 233-24
	22 89 080-082
Strath Conon/Glen Meig	26 88 215-217
Strathdearn/ Findhorn River catchment	10 88 073-075
	11 88 041-043
	16 88 105-108
	41 88 238-244 and 288-290
Strathfarrar	14 82 148-159

3.2.2 Radiocarbon Dating

Conventional radiocarbon dates for material described in Chapters 5 and 6 were determined at the NERC Radiocarbon Laboratory, East Kilbride (Radiocarbon Dating Allocation 660/0896). Two samples were dated using Accelerator Mass Spectrometry (AMS) at the University of Arizona from targets prepared at East Kilbride. Pre-treatments are reported in summary tables below. Radiocarbon dates are quoted to 1σ. They have been calibrated as 2σ ranges according to Stuiver & Reimer (1993), CALIB. Rev. 3.0 using the intercept method. Only calibrated dates were used in interpretation, since the uneven variance with time of radiocarbon and calendar years means that chronostratigraphies based on uncalibrated radiocarbon dates suffer from spurious synchrony and diachrony (Bartlein *et al* 1995).

Calibrated dates are quoted as cal BP. Approximate ages were rounded to the nearest 10 years, since the precision of the dating method does not justify the common practice of quoting calibrated ages accurate to ±0.5 years, and radiocarbon ages have already been rounded in this way (B. Miller, NERC Radiocarbon Laboratory, East Kilbride). In comparative tables using data from the literature as well as from this project, fifteen dates were recalibrated according to Stuiver & Reimer (1993) to enhance comparability.

Radiocarbon dating of peat and soil is subject to additional uncertainties. Peat organic matter is composed of a number of chemically distinct fractions which have widely variable mobility and persistence in natural environments, and often therefore widely different ages (Pilcher 1991, Shore *et al* 1995, Bol *et al* 1996). Humin consists of acid and alkali insoluble organic detritus. The other major components - fulvic and humic acids - are more mobile than humin as a result of both chemical complexing and translocation in groundwater. The majority of peat radiocarbon dates in the literature of the last 30 years have been obtained from bulk samples from which only the fulvic acids (which could cause carbonate contamination) have been removed (Pilcher 1991). The pre-treatment of samples should therefore be a known quantity, and is shown below for all the samples in this study. The fraction of organic matter dated is recorded along with the derived dates in Table 5.1 (Chapter 5) and Table 6.1 (Chapter 6).

The thickness of the slice of organic material dated ranges from 0.5 to 5.0 cm in the literature and is a source of error. It controls the range of ages represented by a single radiocarbon date, as does the accumulation rate of the OM which is rarely well-established (Pilcher 1991). Peat dates for this study were derived from 1.0cm thick slices. Podzol samples were more variable and are described below.

The problem of mixing of carbon of different ages is particularly acute in podzols (Chapter 5) since their existence depends on persistent upward and downward translocation of OM over long periods of time (Matthews 1984, 1993, Schaetzl & Isard

1996, Shore *et al* 1995). The presence of carbon from both younger and older sources is therefore not so much 'contamination' as endemic. Nevertheless, work on alpine podzols (Caseldine & Matthews 1985, Matthews 1984, 1993) has indicated a consistent age-depth correlation. In temperate regions, at lower altitude, where soil development and translocation of chemical species is more rapid, it is uncertain how precisely this relationship is maintained. A similar relationship was inferred in river terrace podzols in Glen Feshie in the western Cairngorms (Bain *et al* 1993, Robertson-Rintoul 1986), although the ages of the soils in Glen Feshie were derived from the depth and colour of soil profiles, and have not been confirmed. If the relationships described for by the above authors are valid, it is still uncertain to what extent they apply to slope podzols in the Scottish Highlands.

In this study, illuviated OM in palaeopodzols was taken from the base of A horizons to sample the oldest carbon preserved, since that constituted the best available approach to the average date when profile differentiation began. Carbon with the longest residence time is expected at the base of the horizon and represents the earliest stages of profile differentiation (Caseldine & Matthews 1985, Matthews 1984). Because of the coarseness of the sediment (sandy till) in which the podzols were developed, samples were taken from horizons up to 2 cm thick.

Podzols are by definition a product of long term leaching, and lateral flow in the infilled gully system with which they were associated, may also in this case have contributed to carbon mixing. Because of this, contamination by younger carbon was suspected, and was confirmed by separate dating of the humin and humic fractions of a single sample (GChS 7a/3), reported as SRR-6061 and SRR-6062 respectively (Chapter 5, Table 5.1). The age discrepancy of 2125 radiocarbon years from different fractions of the same sample reflects previous trials of this type (Shore *et al* 1995). The younger age was therefore discarded in favour of the environmentally persistent humin date (SRR-6061) which was taken as the best available representation of the age of the start of podzol profile differentiation. The variable residence time of carbon in leached podzol horizons may lead to homogenised dates for material of widely separated ages (Caseldine & Matthews 1985, Matthews 1884, 1993). Unless the dates form part of an internally consistent chronology, or can be independently verified, large errors may be difficult to recognise. Where the precise clement of the soil which has been dated is not specified, the uncertainty is increased. Bain *et al* (1993) refer to dating the 'organic' (O?) horizon.

Curry (1999) does not specify the soil layer, but O horizons may be implicit in a reference to the absence of evidence for local burning of vegetation.

Near-surface peat and soil samples were frequently contaminated by rootlet penetration, and hence rejected for analysis. This limited the ability of the project to resolve the most recent events. Sampling of OM also took into account the fact that the lateral hydraulic conductivity of peat is up to two orders of magnitude greater than its vertical hydraulic conductivity (Clymo 1991). Peat pinchouts against permeable sands and gravels were therefore avoided, and vertical rootlet penetration was also considered to potentially compromise low peat vertical hydraulic conductivity. All organic samples were taken with a clean steel trowel, wrapped immediately aluminium foil, then polythene, labelled, marked for orientation, and frozen within 12 hours.

Disconformities created by the burial of organic horizons by mineral horizons potentially introduce large uncertainties into an event stratigraphy based on radiocarbon dating. Erosional and passive contacts between an incompetent, easily deformed, but poorly stratified, organic horizon such as peat, and the superimposed mineral layer, can be difficult to distinguish in the field. If surface material has been removed, the date obtained will be anomalously old. There may be a time gap (on a scale of $0 - 10^3$) years between the date obtained for the organic material and deposition of the overlying mineral horizon. Because the time gap is generally unquantifiable it may be ignored in interpretation (e.g. Curry 1999, Matthews *et al* 1997). The unknown error may be smaller for dates obtained from thin slices of buried peat than for those obtained from palaeosols. In an effort to clarify the nature of peat/mineral boundaries, their micromorphology was examined (Section 3.4.4 below.)

Charcoal may vary from the age of its stratigraphic horizon, and may be derived from wood whose life span is up to hundreds of years. *Pinus sylvestris* in northern Scotland is known to survive for as long as 500 years); and it has been shown to survive, and mix in the environment, over more than one period of sedimentation (Pilcher 1991, pp21-22). The latter evidence however generally comes from lowland sites, and is poorly substantiated on mountain slopes in the British Isles.

These issues combine to materially affect the precision of dates in ways which vary from study to study.

3.2.3 Sediment analysis

Samples of up to 2kg were collected and oven dried at a temperature of $100°$ C until the fine particles flowed freely. They were then sieved and weighed using an Endicotts Certified Laboratory Test Sieve set and an OHAUS TS400D balance. Five features of each fraction and facies were taken into consideration: clast size, mineralogy, weathering, degree of rounding and macroscopic organic content. Mineralogy was identified according to standard texts (e.g. Deer *et al* 1966). Standard grain size analysis based on the Udden-Wentworth scale (e.g. Deer *et al* 1966, Pettijohn 1975 Table 3.3), and definitions of grain sphericity (e.g. Tucker 1981, Fig. 2.5), were used. Clasts of granule size or less, were examined at a magnification of x 20 with a Swift Stereo 80 microscope, and silt-sized particles at up to a magnification of 400 with a Meiji-Labax PTM-1A/05 petrological microscope. Weathering was determined using criteria standard in geological fieldwork and microscopy (e.g. Duff 1993, Pettijohn 1975) to identify the presence of break-down products of the original rock-forming minerals, and physical evidence of weathering, including changes of colour and angularity. The characteristic features of till, debris flows and fluvial sediment were then determined from these data.

Fe/Al concentrations in illuviated horizons of palaeopodzols were measured. The significance of iron and aluminium in podzol pedogenesis lies in the fact that they are released from many minerals in the course of weathering. They may then be retained *in situ*, forming crystalline oxides or hydroxides, or be leached under acid and complexing conditions. Potassium pyrophosphate extracts iron and aluminium, whose values indicate the extent of translocation, from amorphous and organic-bound forms. Given a suitable substrate, favourable conditions for soil profile development are linked to both vegetation and climate. Chelated iron and aluminium complexes become concentrated in B horizons of humus-iron podzols (Avery 1990, Fanning & Fanning 1989, Fitzpatrick 1956, Matthews 1993). The quantity of free iron and aluminium is related to the intensity of weathering, leaching and illuviation during soil development. A high concentration of iron and aluminium indicates either a prolonged period of development, or conditions conducive to particularly rapid podzol formation.

Only limited testing of the relationship of Fe and Al concentrations to soil age and to profile depth was possible. Analysis required samples with around 200g of <2 mm clasts from a thin slice at the base of a B horizon to allow repeated assays. This posed

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problems where the B horizons were as thin as 10 cm, with sometimes gradational boundaries, and clasts of up to 10 cm in diameter embedded in the profile. As a result, only three dated, and one undated B horizon, proved suitable for sampling.

Samples were sieved to separate the <2 mm fraction (W). Fractions weighing from 0.48g - 0.54g were shaken for 16 hours with 50 ml 0.1M $K_4P_2O_{17}$ (potassium pyrophosphate) before centrifuging at 2000 rev./min. for 15 minutes along with a reagent blank. Percentage Fe concentration in the supernatant liquid (C) was then calculated using conventional flame Atomic Absorbtion Spectrometry (AAS; ppm) using the formula (C-Blank)/ (200 x W) (University of Stirling laboratory procedures handbook). Each result is the mean of three assays.

Al was extracted in a similar way but its concentration was measured using Graphite Furnace AAS (ppb).

3.2.4 Micromorphology

Microscopic analysis was designed to detect possible hiatuses in the stratigraphy which could have been caused by shearing and/or erosion at peat-debris flow boundaries. Erosional boundaries of this type which could invalidate dates assigned to debris flows lying on dated peat surfaces. A second aim was to establish whether microscopic features of peat and debris flows were consistent with interpretation based on morphology and stratigraphy. In one case, a dated boundary between well humified and poorly humified peat was investigated. Seven Kubiena tin samples were taken from the two radiocarbon dated localities. Kubiena tins were tapped in a vertical orientation into cleaned, smoothed, sediment sections, until the tin was tightly filled with sample material. Orientation was marked before the filled tins were double wrapped in cling film to prevent desiccation. More extensive sampling was planned, but poor sorting of even relatively fine sediment meant there were few sediment boundaries where grain size was consistently fine enough to be sampled in this way. Each of the seven samples was matched to a sample taken for radiocarbon dating.

Processing was done in the soil micromorphology laboratory of the University of Stirling. Most of the samples combined peat and unconsolidated mineral sediment, and the acutely contrasting properties of the two materials made them technically challenging to process. Due to the low permeability of peat, samples were dried over acetone vapour before being impregnated with resin thinned with acetone. Very gentle

outgassing was needed in one or two cases, in order to replace air bubbles with resin before large 70 x 60mm thin sections could be produced.

Sections were examined using an Olympus BX50 petrological photomicrography system, with an Olympus OM 10 camera. Both plane polarised light and crossed polars were used to obtain maximum information. Photomicrographs, with this information and magnifications specified, are reproduced in Chapters 5 and 6.

Since the emphasis in this thesis is on depositional processes, not pedogenesis, the terminology used to describe the slides is derived from standard geological descriptions of sedimentary particles (e.g. Tucker 1981), supplemented by soil micromorphological descriptions based principally on Kemp (1985) and Elliott (1996). Thin sections descriptions are therefore hybrids, tailored to meet the unconventional demands of the data.

CHAPTER 4: SITE SELECTION AND CHARACTERISATION
4.0 INTRODUCTION

In accordance with project design (Chapter 3, Section 3.1, objectives 1-7), aerial photograph and field surveys of sixteen catchments were undertaken from an east-west transect across the precipitation divide (Fig. 1.3). Locations are shown in Figure 4.1. A combination of aerial photograph and field survey was used to identify Holocene landforms, settings likely to provide long stratigraphic sequences and exposures, and identify sources of slope sediment, and to describe typical sources and sinks of slope sediment and their links, especially gullies.

Five catchment segments where preservation, exposure, and accessibility combined to allow productive investigation, were selected for detailed fieldwork. Additional criteria were the richness of the information available for long time-span study, and wide geographical distribution. They were: Strathfarrar (east of Loch Monar); Glen Cannich at the eastern end of Loch Mullardoch; the middle Findhorn valley (Strathdearn); Gleann Chorainn; and upper Gleann Lichd (Fig. 4.1). The first three sites are described in detail below in order to demonstrate a) that Holocene landscape evolution had taken place as inferred from the preliminary survey, b) that it had occurred across wide areas of northern Scotland and c) that the evolution and geomorphological settings of the two sites selected for dating (Chapters 5 and 6) were characteristic, not rare or unique. Localities described in this chapter also provide a testbed for conclusions about dated sequences.

4.1   PRELIMINARY SURVEY

Aerial photographs had suggested fan-shaped lower slope accumulations together with linear erosion sites on middle slopes, as common landforms likely to preserve and expose Holocene stratigraphy (Chapter 3, section 3.1). Sections of valley floor with a trapezoidal cross-section further suggested onlap of alluvium and interdigitation with lower slope deposits. Little altered, relict land surfaces could sometimes be distinguished from those substantially reworked since deglaciation, by the partial removal of swarms of elongate moraine mounds, reworking of inverted ice-contact topography, truncation of eskers, fluvial terracing or erosion of fan toes, cone and fan extension, or incision of glaciofluvial terraces by gullies terminating in fans.

Field work verified identified datable sequences in exposures reworked areas contained. Many slope sections consist of eroded stacks of coarse mineral sediment, with rare

clean faces, and apparently single stratigraphic units. Vegetated fans and cones were common, but frequently too coarse for coring. Linear erosion of hillslopes at the margins of reservoirs predominantly exposed sections consisting of a single layer of thin peat, sometimes with buried pine stumps, over sandy, bouldery till. Organic matter beneath the surface of dissected fans was not always *in situ,* and therefore inappropriate for dating. Some hillslope debris flows overlay thin soils which were possibly both transported as thin slabs, and susceptible to carbon contamination. In other cases, single horizons offered little scope for long sequence dating. In one locality (Glen Markie, Fig. 4.1), the stream of an active fan, found at the end of one field season, had obscured and/or eroded a sequence of stacked mineral/peat couplets by the start of the next, when sampling was planned.

Stepped suites of fluvial terraces with two or more low Holocene risers at the base were found in several areas. But fills were generally devoid of exposures of buried peat or soil, and single, peaty terraces offer little scope for meaningful dating if due to lateral accretion. Lower slope undercutting revealed no exposures to match that dated by Brazier (1987) in Glen Feshie.

As a result, five localities emerged as suitable for detailed work. Sites where evidence of episodic slope failure is preserved must have been active (erosional), but at the same time must have sediment traps which preserve long records. Three types of trap produced viable information: debris fans fed by rock-cut gullies; a lower slope where debris had accumulated over a concave bedrock surface above a river gorge; and abandoned river channels. All are common features of mountain landscapes in the Northern Highlands, but the first two provided only rare exposures suitable for high resolution dating. Since a strong association was found between gullies and slope failures which had occurred between intervals of organic accumulation, the nature of gullies, and of processes associated with them, were further investigated.

4.2   GULLIES

The impact of slope hydraulic conductivity on landform evolution was discussed in Chapter 1, Section 1.5.1. The prevalence of hillslope gullies in the areas surveyed, therefore suggested they might play an important role in Holocene slope sensitivity. Field observations were related to the literature for Northern Scotland (Sissons 1961, Young 1975, 1980), investigations of recently deglaciated gullied slopes in Norway (Ballantyne & Benn 1996, Curry 1999), and studies of the hydrology of currently

glacierised catchments (Richards *et al* 1996, Sharp & Richards 1996). This section describes typical gully systems seen in aerial photographs.

Three types of gully are distinguished in the literature: headward eroding gullies in till and colluvium, probable postglacial gullies (see below) and structurally controlled gullies. The first of these has been described by e.g. Brunsden (1979), Schumm (1979), Harvey *et al* (1981), and in the Scottish Highlands during exceptional floods and storms by Fairbairn (1967). Although sometimes terminating in small vegetated fans, the latter were not found here to contain buried organic material, and, appeared, as far as could be determined in the field, to have been deposited in single events. They are not further discussed here.

Typically, the first group of gullies identified from aerial photographs and in the field

- form in till and/or bedrock
- have an upper limit coinciding with kame terrace fragments and/or a linear mid or upper slope bedrock/drift boundary
- have a lower limit which coincides with the upper limit of a blanket of valley floor ice contact landforms (i.e. do not occupy the whole vertical slope space)
- lie sub-parallel to the maximum dip of the slope (i.e. are vertical and linear)
- form in closely spaced clusters of irregular lateral extent
- vary in abundance independently of major joint and foliation patterns
- have a strike which diverges from that of rock discontinuities
- are absent where ice-scoured rock predominates
- occupy the same vertical space as oblique retreat moraines
- are unequally developed on opposing valley walls of all orientations
- are largely restricted to the area of Loch Lomond Stade ice cover.

A pre-Holocene, probably sub-glacial origin for these gullies was proposed (Sissons 1961, Young 1975, Young 1980), in part because they have a source area of only a few square metres, yet may be 6-10m deep and 3-125m wide. Young (1975) described them as 'more or less perpendicular to the slope' in a section of Glen Feshie (Western Cairngorms). Like Sissons (1961), he noted that they may start and end abruptly without traversing the full slope length, a characteristic confirmed here. He concluded that they formed during deglaciation when water draining from ice-free upper slopes joined meltwaters beneath the ice, with their upper limits marking the upper surface of former valley ice. Kame terrace fragments in Gleann Chorainn (Fig. 4.2) and Gleann

Fhiodhaig (Fig. 4.3) provide examples of this relationship. A section of slope on the south side of Loch a' Chroisg in Mid Ross, (NH 120580, Fig. 4.1) provides ambiguous information about the timing of their formation. Here crescentic retreat moraines (also mapped by Bennett & Boulton 1993, Fig. 7) appear to either divert the course of near-vertical gullies, or overlie their lower ends (Fig. 4.4).

Three lines of argument are consistent with a postglacial origin of such gullies (Ballantyne 2001, pers. comm). i) lower parts of subglacial gullies are likely to have been closed by ice movement, ii) subglacial meltwater channels are generally aligned sub-parallel to the ice margin (Sissons 1961), and iii) lower parts of gullies are diverted by nested recessional moraines, suggesting that gullies post-date moraine deposition.

A second group of gullies seen in the aerial survey exploit excavated joints and other geological structures. Typically, they

- are developed in excavated rock structures such as joints, foliation, minor intrusions
- have a distribution and frequency which matches these features
- extend the whole length of slopes and are traceable above valley walls into periglaciated terrain and/or onto valley floors
- are oblique to the slope and to deglacial gullies unless rock structures are coincidentally vertical
- zig-zag in shape where they follow the path of intersecting structures such as joint sets.

Their formation has been attributed to subglacial hydrology combined with periglacial weathering. High pressure in the subglacial water system enhances decoupling between the ice and the substrate, reducing the strength and viscosity of subglacial sediments in the absence of permafrost (Richards *et al* 1996, Sharp & Richards 1996). Sediment-laden water under high pressure thus gains access to bedrock, widening and excavating pre-existing planes of weakness such as joints. At the same time, periglacial frost action creates open joints above the ice surface. After ice retreat, long gullies echo the trend of single, or linked, rock structures.

Since investigation of pre-Holocene processes is beyond the scope of this thesis, for convenience, the first group of gullies is referred to as *deglacial* and the second *structural*, although neither term fully describes their origins. Terminal fans with buried organic horizons due to episodic aggradation were found only below structural gullies.

4.2.1 Sensitivity of gullied slopes

This section explores the distribution of deglacial and structural gullies by examining examples seen in aerial photographs. The purpose is to provide a basis for evaluating relationships between gully distribution and the sensitivity of the locations investigated in this, and subsequent chapters.

4.2.1.1 Results

Upper Gleann Lichd

The north-east and south-west walls of the valley present strongly contrasting views (Fig. 4.5). To the north, a valley-parallel joint set is well displayed. Long structural gullies with basal fans are restricted to the opposing flank. This variation is due to the intersection of dipping rock structures with the valley walls (Fig. 4.6). Where joint faces, cleavage planes, or bedding are sub-parallel to a valley wall, run-off is rapid, sediment cannot accumulate easily or persist if it is deposited, and mass wasting is facilitated by the lack of storage sites. At the same time, long open downslope fractures can develop. On the opposing side, where the same structures dip into the hillside, they provide a series of impediments to rapid mass movement and the development of long structurally controlled gullies. The presence of fans, each at the base of a long gully on the south-west flank of upper Gleann Lichd, compared to none on the opposite wall in the upper valley bear out this interpretation.

Gleann a' Choilich

Gleann a' Choilich (Fig. 4.7) runs north-east towards Glen Cannich from the divide with west Glen Affric. Intersecting joint sets are well exposed on high level areas of ice-smoothed rock which have a cover of only thin, patchy till. Some structural gullies have the same trend as one or more of these sets, demonstrating their origin in excavated joints. On sections of both valley walls there are more closely spaced, sometimes discontinuous, gully sets with a distinct vertical limit. These are deglacial gullies. A cluster of small cones well above the valley floor suggests deposition when it was still filled with ice.

Gleann Fhiodhaig

In Figure 4.3 the cross-cutting trend of structurally controlled gullies and deglacial gullies is clear. Only the former terminate in fans. On high ground, ice-polished, sometimes fluted, bedrock has few open joints.

Strathfarrar

Structural gullies and fans are very widely spaced (Fig. 4.8), and no deglacial gullies are visible. Ice-smoothed rock above drift covered slopes shows cross cutting joint sets and fans at the base of long gullies are relatively small. In the field, it is clear that, where major rock discontinuities are sub-parallel to the valley floor and these intersect with high angle, valley-parallel joint sets, sections of the valley walls have a stepped profile and zones of low average gradient. As a result, high level micro-valleys run parallel to the main valley, and are important stores for sediment and water.

Strathdearn (middle Findhorn)

Gullies are absent in much of Strathdearn (Fig. 4.9), and only sparse and fragmentary jointing is visible on higher ground. Given that bedrock consists of schists and gneisses with sheared, syn-deformational granites (Section 4.6 below), a lack of joints is unlikely, and they must be obscured by a combination of grus and peat. Well-developed dendritic drainage is a reflection of poorly channelled flow through peat and grus.

Slow-moving, cold based ice, which melted rapidly during the Lateglacial, accounts for grus preservation in Strathdearn (Stephenson & Gould 1995). The large number of meltwater channels which made broad, shallow indents in the substrate as the ice thinned, dominated Lateglacial drainage. Some of the meltwater channels, which have broad, shallow cross-sections, run in swarms sub-parallel to the valley on the margins of the plateau, providing sediment stores on the valley walls. Others are roughly normal to the main valley, and have been exploited for Holocene drainage by relatively small, high sinuosity, misfit streams (Fig. 4.9). Holocene slope hydrology in Strathdearn thus has different starting points, and different evolutionary pathways, from catchments described above. Other meltwater routes may have been the sites of massive till failures, leaving small V-shaped side valleys with basal cones, sometimes with crude, fluvially incised terraces. There is no sign of repeated cone or fan aggradation burying buried organic horizons, although some cones have been fluvially extended.

4.2.1.2 Discussion

There is no clear relationship between deglacial gullies and the presence of lower slope cones and fans. A likely explanation is that the small source area of subglacial gullies noted by Sissons (1961) and Young (1975) (Section 4.2.1 above) is insufficient to create fans by a process of repeated debris flows.

There is however, an association between long structural gullies and large terminal fans

as in Gleann Fhiodhaig and Gleann Lichd (Figs. 4.3 & 4.5). Fans formed by repeated debris flow are common in the Highlands, for instance in Gleann Lichd NG 976207, Glen Shiel NG 958171, Srath Duilleach NH 033288, Gleann Fhiodhaig NH 108479. Where bedrock is homogeneous and ice-scoured, as in Strathfarrar (Fig. 4.8) there are few such fans, and they are small relative to gully length. Upper slopes of homogenous bare rock restrict infiltration and enhance run-off. But where intense percolation combines with frequent structural gullies, as in Gleann Lichd, repeated slope failures channelled by gullies are more common.

4.3   BURIED ORGANIC MATERIAL

The field survey identified sufficient buried organic matter to suggest that in some areas, landforms had been active during the Holocene to an extent not described in the literature. Organic material beneath Holocene landforms was logged from Glen Markie in the east, to Glen Lichd on the west coast, at altitudes from 50 to c. 300m OD (Fig. 4.1). It included

- buried organic sediments in infilled kettle holes
- *in situ* tree stumps below peat and debris flows
- alternating peat/mineral strata
- truncated podzols beneath hillslope and gully debris flows
- charcoal slope washes and lenses
- palaeochannel fills of peat, fragmented wood and organic muds
- peaty soils within massive debris cones
- brecciated peat and *Calluna* roots in homogenised debris
- disrupted, laminated sands with sedimented twigs
- eroded peat incorporated in debris flows.

Where the stratigraphy suggested a long history, more detailed investigation was undertaken. Two areas, upper Gleann Lichd and Creagan a' Chaorainn, best met the criteria for dating (Chapter 3, Section 3.1). Locations in each of the other three valleys are described below.

4.4   STRATHFARRAR: EVIDENCE FOR HOLOCENE LANDSCAPE EVOLUTION

Strathfarrar (Figs. 4.1, 4.8, 4.10) drains east across the centre of the Northern Highlands. From the headwaters to the localities investigated, it encompasses 215 km^2 of mountainous terrain rising to over 1000m. Bedrock is Moine Series granulites and

psammitic, occasionally pelitic, schists with lenses of Lewisian gneiss. Ice-moulded outcrop rises from 120m to over 960m with a cover of bouldery till with an abundant sandy matrix. Deglacial gullies are absent (Fig. 4.8), and structural gullies are widely spaced, only occasionally terminating in fans. Slope drainage is dominated by two intersecting sets of widely spaced joints. Major joints intersect to form valley-parallel benches which, in places, trap large volumes of sediment and water.

An 'old lake terrace consisting of a considerable thickness of sand and gravel', was noted in a technical report by the IGS for the Hydro Board (Lawrie & Wright 1955), and implies a previous high stand and subsequent fall. The site at Loch Beannacharan (Fig. 4.10) is now occupied by a hydro-electric dam. However, just upstream, a shallow, peat-filled channel terminating in an area now above water level (a palaeo-inflow?) and underlain by thick gravels, found during field work, suggests that a former lake margin lay at least 2-3m higher than today.

Two localities which yielded information about the nature and timing of the events which have shaped the landscape are described below. They lie within the limits of the Younger Dryas ice sheet as mapped by Bennett & Boulton (1993).

4.4.1 Allt Innis na Làrach Fan

The fan (NH 262383) runs north at right angles to the modern River Farrar (Fig. 4.8). It is elongate (Fig. 4.11), and, unusually, narrows towards its terminus because the distal fan and part of the mid fan developed between rock ridges fringed by two 3m terraces (Figs. 4.12, 4.13a & b). A stream originating in the gully (Allt Innis na Làrach) dissects the fan.

The proximal fan

The upper fan is a steep bouldery cone whose apex is at nearly 200m OD (Fig. 4.11). Relict surfaces show that the original landform was twice (?thrice) excavated and extended at a slightly lower angle (Fig. 4.14).

The mid fan

The gently convex mid fan is separated from the upper fan by a slope break (Fig. 4.15) Sheets and lobes of mineral sediment beneath peat give it a convex profile. A slight change of surface angle at the mid/distal fan boundary marks the limit of transport for sediment larger than coarse sand. Down-fan fining indicates water-dominated flow as

the principal agent of lower fan aggradation. A broad, shallow, boggy channel below the 140m contour provides an exit downstream to Loch a' Mhuillidh from the lower part of the mid fan (Fig. 4.11) This feature cuts across processes of fan extension, trending *up* fan. Rock-cut benches fringe the gap in the cross-valley rock ridge through which it passes (Fig 4.16).

The distal fan

A very low angle spread of peat with sand terminates abruptly 3m above, and 60m from, the River Farrar (Fig. 4.13). Fan toe peats and thin sands terminate against a coarse, matrix-supported gravel deposit. Imbrication in this deposit is poorly exposed in the pile of river cobbles, but where visible, clearly indicates *up-fan* flow at this point, matching the mid fan exit to Loch à Mhuillidh.

Fan architecture is known from five bank sections and two gouge augur holes on the adjacent fan surface, all in the distal fan (Figs. 4.15, 4.17 - 4.19), where spreads of sand sourced in the slope above the proximal cone have extended episodically to the most distal zone, interrupting peat growth. Peat forms a high proportion of the sequence in some sections, but five marker horizons, which are reviewed below in stratigraphic order, have been resolved. The key to symbols in all graphic logs is on a fold-out page at the back of this thesis.

4.4.1.1 The fan substrate

Near the fan terminus, the substrate is water-worn bedrock (Sections SF1, 2) or rounded, armoured gravels which persist for at least 20m upstream (Sections SF3, 4, 5). Both are typical of high energy river beds, and inconsistent with small fan stream or flood sheet deposits derived from angular till. Fan gravels in the gently convex mid-fan structure are subangular rather than subrounded, and not imbricated. The basal gravels, like the water rounded bedrock, therefore predate the accumulation of 2-3m of distal fan sediment.

4.4.1.2 Sandy peat

Apart from Section SF5, basal deposits in all the logs show peat with sandy flushes, indicating minor slope erosion, concurrent with humid conditions and periodic sheet floods.

4.4.1.3 A grain flow deposit

The next correlatable deposit above the sandy peat is unstratified and of variable thickness. It appears in Sections SF2, 3, 4 and 5a and in several intervening bank exposures, and represents a grain flow when 'floating' cobbles were transported from above the fan apex for over 600m to Section SF3, in a dense silt/sand/water mix. Pale grey, semi-plastic silt, containing a deformed, sheared, disrupted charcoal layer from the same event, is preserved in Section SF3 and up to ten metres upstream. Silt and sand have a thickness of up to 95cm. Iron staining in the sand, plant fragments, charcoal and silt confirm that the flow included soil.

4.4.1.4 Overbank sands

Sandy, silty overbank deposits with centimetre thick peaty stringers, and sharp lower and upper boundaries, appear in Sections SF2, 3, 4 and 5. They indicate that suspensions of sand and silt repeatedly overlapped bank vegetation at intervals frequent enough to impede continuous peat formation, over a sustained period of time. Away from the stream banks, the lateral equivalent is dense black peat. The limited distribution of overbank sands contrasts with sheet flooding which deposited flushes of sand within the peat.

4.4.1.5 Modern soil

The uppermost horizon on the stream banks is an immature soil, 20 - 40 cm thick, with no evidence of profile development (Figs. 4.17 - 4.19). The base is sharp and apparently non-erosional. The preceding period of small-scale overbank deposition appears, therefore, to have terminated, as well as started, abruptly. The lack of lamination in the sediment in which the modern soil is developing suggests its deposition in a single event.

4.4.2   Buried pine stumps

Large *in situ* stumps of *Pinus sylvestris* are exposed at two points in the banks of the fan stream, one close to Section SF5, and the other on water-worn bedrock close to SF2 and 3. Adjacent to SF5, stumps about 1m below water level sit on rounded fluvial gravels beneath 1.75m of peat. No pine or other species was found higher in the succession. The age of these stumps would define the maximum age of the distal fan. The overlying peat thickness suggests they lived thousands, rather than hundreds, of years ago.

4.4.3 Subsurface Peat Erosion

In Sections SF4a, (Fig. 4.18) and SF5a, (Fig. 4.19), amorphous peat disintegrated when auguring and excavation were attempted. Overlying peat was compacted and coherent, so disintegration must have occurred *in situ* beneath the surface. Water in contact with peat has a low pH and low redox potential, leading to high concentrations of Fe and Mn in solution (e.g. Robins 1990). As a result, peat can produce flushes of iron solutes and colloids which stain stream waters, and precipitate as dark reddish-brown filaments and coatings on sediment and vegetation (Fig. 4.20). The circumstances, frequency and extent of these events have not been reported, but they must represent a volume loss in the peat column, and possibly sub-surface peat disintegration and erosion.

4.4.4 The Braulen Fan

The Braulen fan in Strathfarrar (NH 238387) is 2.5 km west of Allt Innis na Làrach (Fig. 4.10). Its source is a gully on the margin of the floodplain of the River Farrar which is about 500m wide at this point with a single terrace. *In situ* pine stumps lie below gravels and floodplain peats (Section SF6, Fig. 4.21). Pine can only have grown when the margin of the floodplain was relatively dry. But the sediment enclosing the pine stumps records a rising water table, culminating in a massive flow which buried the deposits beneath a fan-shaped spread of coarse, gully-derived debris.

4.4.5 Discussion and summary

The stratigraphy of the Allt Innis na Làrach and Braulen fans are comparable in that each sequence records a rising water table, accompanied by slope instability.

The Allt Innis na Làrach debris cone was twice incised and extended to form crude lower terraces, terminating nearly 50m above the outwash surface, possibly above the surface of a thinning valley glacier. The 3m high terraces bounding the mid and lower fan are probably glaciofluvial because they are below the outwash surface, but the upper terrace has the same elevation as the glaciofluvial terrace fragments bounding the River Farrar just upstream. The breadth and depth of the high terraces is a gross misfit with the power of the modern fan stream.

Three features indicate high energy flow beneath the fine sediments of the distal fan: imbrication in the large, matrix-supported gravel bar against which distal fan deposits terminate; water-worn bedrock; and armoured surfaces of rounded gravel. The base of the imbricated cobbles is 5m above the present day River Farrar, and they must be older

than the low energy sediments which abut them. High energy flow in an *up*-fan direction is indicated by gravel imbrication, and by the disappearance of these features before the rock-cut exit down-valley to Loch a' Mhuillidh, which could not have been cut by the small fan stream. High terraces below the outwash surface, which dip very gently towards the rock-cut exit, confirm large flows. These data are interpreted as indicating the course of the palaeo-Farrar which flowed south for a short distance up the depression later filled by the north-trending fan.

Incision of about 5m and the large gravel bar appear to have prevented re-invasion of the abandoned channel during subsequent floods, and peat accumulated on gravels and bedrock in damp hollows in the abandoned channel. Pine is intolerant of water logging, and was able to mature before the water table rose in parallel with distal fan aggradation, accelerated slope erosion, incursions of mineral sediment. and downstream, and development of thick peat. In the mid fan, sheets of angular gravel, giving way to sand and peat down-fan, can only have originated upslope. Since flow was predominantly water-dominated, an incising stream probably evolved in parallel with the mid and lower fan.

Intervals of uninterrupted peat growth indicate an absence of both mobile sediment and the erosive floods required to transport them. Section SF1 (Fig. 4.17) shows two periods when peat near the fan terminus was repeatedly invaded by millimetre scale sand sheets, and two alternating periods devoid of sand deposition. About 15m upstream, sandy peat in Section SF2 records wet, erosional conditions at the time when basal deposits were forming on the recently exposed bedrock of the former river bed. Augured sections SF4a and 5a (Figs. 4.18 & 4.19) confirm that mineral sediment was spread across the low angle fan surface. At times of high water table and slope erosion, coarse sand, pebbles and cobbles in the medial fan passed laterally downslope into peat with frequent thin sandy flushes. When slope erosion was minimal, the distal fan equivalent of sandy peat in the mid fan was peat with little or no mineral content.

Conditions leading to the grain flow which carried the eroded podzol can be inferred from Sections SF2 and 5a. The lateral equivalent of the grain flow in the upper reaches of the distal fan is waterlain pebbles and coarse sand, so it was a probably a single, groundwater-driven event. Downstream, peat was invaded by sand-bearing water. No information is available about any role played by the burning of slope vegetation.

The thickness of peat overlying the grain flow - 55cm - can be estimated from Section

4a. Since no peat growth rates for the Northern Highlands in excess of 7cm/100yrs have been reported (Binney 1996; but see also discussion in Chapters 5 and 6), the overlying peat suggests that the soil failure occurred at least hundreds, and possibly in excess of a thousand, years ago. The rarity of sediment flushes in the lower fan confirms it as a large, rare event, and its sharp upper boundary emphasises the sudden change to a lower energy environment when repeated minor flooding transported only fines in suspension. A further sharp boundary between overbank deposits and the modern soil indicates the cessation of the conditions which produced the gradually accreting overbank sediment sequence. Table 4.1 summarises the inferred sequence of palaeoenvironmental change.

Lateglacial	Debris cone formed between gully base and valley glacier
Lateglacial-Early Holocene	Cone surface lowering in 2 events; outwash incised by the glaciofluvial River Farrar which incised two wide 3m high terraces and incised a rock-cut channel through a cross-valley rock barrier.
?Early-mid Holocene	Avulsion changed the course of the River Farrar to a deep new channel running directly south. Probably early development of the mid fan.
?Mid-late Holocene	The abandoned channel was colonised by vegetation which formed peat in damp hollows but supported mature pine trees in drier areas.
?late Holocene	Wetter conditions and a rising water table initiated slope erosion and peat growth to form a distally fining fan in the elongate hollow of the former river channel. Waterlogging displaced the pine woodland. Massive podzol erosion occurred in association with burning. Surplus water was channelled in an incising stream originating in the gully (Allt Innis na Làrach). It changed course on at least one occasion.
Late Holocene	The abrupt onset of minor floods charged with silt and sand frequently interrupted growth of stream bank vegetation
?Recent	Unstratified sand and silt was deposited on top of the overbank facies, forming the matrix of the modern soil.

TABLE 4.1
Inferred postglacial history of part of Strathfarrar

4.5   GLEN CANNICH: EVIDENCE FOR HOLOCENE LANDSCAPE EVOLUTION

Glen Cannich runs roughly parallel to, and about 5 km south of, Strathfarrar (Fig. 4.10) The catchment includes a number of summits over 1000m, deglacial gullies are absent, and streams follow straight paths controlled by rock structure. Younger Dryas ice filled the upper and middle valley (Bennett and Boulton 1993) which has an area of 160 km^2. Four localities within 4.5 km of each other were logged (Fig. 4.22). They are described below from east to west.

4.5.1   The Liatrie Fan

Drainage from a col in the ridge between Strathfarrar and Glen Cannich is in a fracture which runs north to the Allt Innis na Làrach fan, and more steeply south to Liatrie in Glen Cannich, dropping 400m over a distance of 3km. Liatrie at 200m AOD, is surrounded by 800m summits dominated by ice-smoothed outcrops of schist, with widely spaced joints.

The Liatrie fan (NH252326) is at the base of the Liatrie Burn, and dissected by it. It has a maximum width of 100m and grades to the floodplain. It consists of coarse, matrix-rich debris with some pebbly, sandy zones (Fig. 4.23a) whose composition and texture match that of till exposed in the source gully. Fine floodplain sediments onlap the fan toe whose base is not exposed. Traces of imbrication and rare, crude stratification, indicate some vigorous water flow. However, the poor sorting, coarseness, matrix-supported texture, near absence of stratification, and lack of distal fining, combined with the presence of randomly distributed peat fragments and rootlets, show that deposition was mainly through energetic debris flows.

Two horizontal, blackish-brown horizons about 5cm and 20cm thick respectively, were found near the base of the exposure. Maximum continuous lateral exposure in the unstable stream bank was 1m. The horizons contained variable amounts of homogenised peat particles, and were moderately to poorly cemented by Fe/Mn precipitates. Their upper and lower margins varied over a depth of several centimetres because cement and peat particles had irregularly penetrated the highly permeable debris.

4.5.2 The Allt Coire Eòghainn fan

The fan (NH 238323, Fig. 4.22) emerges from a valley-parallel fault gully (Fenton 1991), and partly buries an older, very small fan at the base of a pair of structure gullies which rise vertically for 250m up the slope immediately above it (Fig 4.24). Fan stratigraphy is partially exposed in the banks of the small incising stream. The postglacial fault scarp disappears beneath peat towards Creag Feusag, reappearing briefly on its western flank above the east end of Loch Mullardoch (Fig. 4.22). From stream displacement and fault gouges (some enclosing peat), Fenton (1991) calculated several episodes of postglacial lateral movement amounting to 25m, as well as vertical displacement. He inferred faulting after ice retreat, but before blanket peat establishment, which he assumed to be mid-Holocene (c. 6000 BP), but may well be much earlier.

Some fan building postdated growth of at least 0.5m of peat (Fig. 4.25). The three elements of Figure 4.25 allow reconstruction of successive environments at this locality. Wooded ground (indicated by the abundant wood in the peat) gave way, in association with a rising water table, to an extended period of peat growth when, judging by the absence of mineral sediment in the peat, the slope was stable. Thereafter, run-off deposited 20 cm of water lain pebbles, sands and muds on the upper fan peat. Finally, the till-rich gully walls released debris in a flow up to 2m thick, covering all earlier deposits.

4.5.3 The Allt Crumaidh fan

Allt Crumaidh, NH230314, occupies a deeply excavated structural gully just east of Loch Mullardoch (Fig. 4.22). A debris cone at its base has been extended to form a low angle fan with maximum apron width of 100m. The dam has changed river flow and water levels immediately downstream, so that the extent of erosion and submergence of the terminal fan on the margin of Loch Sealbhanach is uncertain.

Shallow surface channels and lobes of distinct materials indicate several pulses of aggradation. The course of the entrenching stream changed relatively recently because it truncates a historical dry stone dyke which protrudes only 30cm above the fan surface. Exposure of distal fan stratigraphy is poor, but from an eroded section on the shore of Loch Sealbhanach the fan terminates without fining in unstratified debris derived from

till. Further exposures are visible in the banks of the incising stream. Bouldery fan debris consisting of 1.5m of partially winnowed sandy till is visible 120m upstream from Loch Sealbhanach. Some mainly sand-sized deposits up to 1m thick confirm fluvial processes. The structure is therefore a debris cone extended mainly by debris flow, but with fluvially reworked lobes and lenses.

On the low angle distal fan, the stream dissects a lobe of homogeneous, structureless, medium sand with silt, intensely stained by haematite (Figs. 4.26a & b). From limited exposure, the deposit is at least 20m wide, 15m long and 0.9m thick. Randomly scattered within the sediment are eroded peat lumps, or peat stains which pseudomorph the original material. These are up to 10 cm in length with irregular boundaries and a pointed tail, but no preferred orientation. Well-spaced, randomly oriented, black-coated moulds of *Calluna* roots precisely retain the shape of living roots on the fan surface. Rare charcoal fragments 1-2 centimetres long are randomly dispersed in the sediment.

Beneath the terracotta deposit, the top 30cm of a coarse mineral horizon with a fine matrix, and zones where cobbles or pebbles dominate, is exposed. Larger clasts show some fluvial rounding, but elongate particles point downstream in the opposite direction from an imbricated river bed. Together with poor sorting this feature is characteristic of debris flow. As at Liatrie, there is a thin layer of Fe/Mn cement near the base.

4.5.3   The north shore of Loch Mullardoch

A hydroelectric dam containing Loch Mullardoch obscures the stratigraphy and morphology of the low valley walls. During fieldwork, the level of the reservoir was temporarily several metres below the normal summer range, exposing a wide eroded wave zone, and exposing debris flows of bouldery, sandy till with a variable peat cover beneath which buried *in situ* tree stumps are common. Peat overlies sandy till which itself overlies laminated sands and silts, and a stiff, olive-brown, stoney, sandy, clay. This association of facies often represents blanket melt out till, proglacial lake deposits and lodgement till respectively (Richards *et al* 1996). The laminates consist of repeated, thin, fining up sequences, or couplets which are dominantly sand or dominantly silt respectively, and have a lateral extent of >1km consistent with annual meltwater deposition in standing water.

The sites described are within 1.5 km of the mapped western end of the Allt Coire Eòghainn postglacial fault trace, and extend from Benula Lodge, NH206318 to Allt

Mullardoch, NH216316 (Fig. 4.27). The following observations and discussion are organised so as to clarify any contribution of neotectonics to landscape response.

Evidence consistent with postglacial seismicity

i)   Ball and pillow structures and microfolds are a classic result of pressure release during liquefaction caused by ground shaking (Fenton 1991). They can also be seen in sedimentary rock adjacent to faults. Two such structures, forming an almost perfect sphere about 10 cm across, with concentric layers, were found in lodgement till (stiff, olive sandy clay with randomly dispersed angular pebbles) below normal reservoir level, at NH217316 (Fig. 4.28a). Small scale overturned folds in sandy till were also found (Fig. 4.28b).

ii)   Rock brecciated *in situ* is exposed on a gently sloping, ice-smoothed rock bench (NH 215316) and in an apparent moraine mound (Fig. 4.28c). Fresh, non-linear open fractures up to 20cm wide are unrelated to rock structures. The rock mass has been subjected to very high strain and has failed without being transported.

iii) Below the Benula Lodge boathouse, laminated sands dip beneath a mass of weathered clay-rich, stony till. This represents an overturning of strata, probably through slumping. Downslope, beneath peat stripped by wave action, a transition from gently dipping to near vertical laminae was seen (Fig. 4.28d). Similar exposures were found along a 60m stretch of the low loch margin, indicating widespread inversion and deformation.

iv) Millimetre to centimetre scale laminae in sands and silts overlying the olive till are folded. Microfaults displace folded laminae. Peat is folded with the laminates (Fig. 4.28e). The folding is therefore Holocene in age. From the top down, over a depth of 0.5m, sub-horizontal laminae give way to upright folds then overturned folds, with peat congruently folded with the mineral sediment, then chaotic convolutions with disrupted folds and fold limbs ending in a structureless mass or detached fold closures (Fig. 4.28f). Some layers are intruded by sub-vertical stringers of sand. Some folds in the homogenised peaty layers are pierced by very angular pebbles (Fig. 4.28e & h) similar to those in the overlying till.

v)   Laminated sands at the foot of a 10-12° slope of ice-scoured outcrop with no feeder gully close to the Benula Lodge boathouse (Fig. 4.28g) are emplaced on ice-smoothed bedrock, but folds are sharply truncated at their bases against the rock and cannot have

developed *in situ* (Fig. 4.28h). Their upper surfaces too are truncated by a 50 cm thick layer of sandy till with angular boulders.

vi) The fine matrix of the debris flow noted in v) is homogenised peaty, iron-rich sand. Its deposition therefore post-dates peat formation.

vii)   Some coarser laminae are crudely cemented with blackish Fe/Mn precipitates similar to those seen at Liatrie and Allt Crumaidh. Microfaults with a throw of 2 cm displace folded, cemented laminae. The inferred sequence of events is therefore:

a)   deposition of laminar sands and silts

b)   peat growth

c)   peat disintegration

d)   preferential cementation of coarser laminae

e)   folding

f)   microfaulting.

Steps e) and f) could have been caused by faulting in the early stages of the mid Holocene. Cementation must have occurred while the laminae were horizontal, and probably when they were subject to a high standing water table because the cement is not diffused through adjacent layers, but preferentially follows coarser, sandier layers.

viii) At the foot of Allt Mullardoch, which occupies the trace of a small fault (Fig. 4.22), a debris flow in the till sequence incorporates small patches and streaks of peaty organic matter. A large boulder has pierced a sand lens (Fig. 4.28i), and on the downslope side of the debris flow, the sand lens is displaced about 40 cm upwards. Along with finer debris, it appears to have toppled in to fill a gap formed when the sand lens was fluidised or faulted.

ix) Figure 4.28j shows a possible peat injection breccia. Peat appears to have invaded the matrix of a debris flow, leaving thin stringers in places, and elsewhere a mosaic of peat and sand with pebbles. Injection breccia is found in rocks fluidised during ground acceleration and liquefaction, but no descriptions of soft sediment injection breccias were found in the literature.

x)   Figure 4.29 shows a dissected, morphologically typical, moraine mound. But its internal features are incompatible with this genesis. The ground behind the mound is a long, gentle, irregular slope of <10° and the materials are not homogenised as would be

expected in a debris flow. Instead they indicate violent disruption of the succession of peat, laminates, till and bedrock.

Evidence consistent with non-seismic interpretation of slope activity

xi)   'Neotectonic' folding and microfaulting as described in iv) above, as well as sandy stringers and detached fold closures could be caused by liquefaction and slumping of sediment as it was rapidly invaded by a rising lake level (K. W. Glennie 1998, pers. comm.).

xii)   By analogy with previously described cements, the source of iron/manganese cement in some layers of laminated sand and silt is subsurface peat disintegration. An alternative source is subglacial, which would date the cementation as Lateglacial. However, iron and manganese expelled from water as it freezes during regelation under sliding glaciers usually form a thin, rock varnish. (Whalley *et al* 1990, V. M. Haynes, University of Stirling, pers. comm.).

4.5.4   Discussion and summary

Disruption of glacigenic and Holocene sediments has occurred along a 1200m slope cross-section from the Benula Lodge boathouse, where at one point, laminated sands can be seen dipping beneath olive-grey, pebbly clay-rich till, to the foot of Allt Mullardoch, where a debris flow contains vertical, elongate boulders and peat fragments, and overlies faulted sediment.

Some of the features described in i) - x) above are characteristic products of ground shaking and could be Lateglacial or very early Holocene in age. Others are consistent with, but not necessarily the product of, seismic activity, although the conjunction of features strengthens the case for seismicity. The most unambiguous evidence for neotectonics as the cause of the disturbances consists of *in situ* shattered bedrock and the ball and pillow structures in the lodgement till in proximity to the inferred surface trace of the fault. Plastic deformation of a stiff, impermeable clay requires liquefaction whose origins are difficult to envisage if dissociated from ground acceleration which may have occurred at an early stage of isostatic readjustment after deglaciation. Glaciolacustrine laminates are commonly subject to early post-depositional slumping and gravity flows due to rapid deposition, but folded laminates appear to have slid over bedrock and then been truncated by a debris flow which contains organic matter. On a

long, gentle slope where ice-smoothed bedrock outcrop is common, and gullies are not, the cause of debris flows is not obvious unless seismic shaking is invoked. The pseudo-moraine mound is also consistent with a seismic trigger. An injection breccia of peat in sand is most easily explained as the product of a brief, high energy pulse during which materials lost their coherence. Sudden deposition of a mass of debris on waterlogged sand and peat might disrupt and homogenise the substrate. But like the shattered *in situ* bedrock, the injection breccia strongly suggests ground acceleration typical of seismicity. The inclusion of peat and peat precipitates in debris flows and contorted sands by Loch Mullardoch confirms that at least some of these ground deformations occurred within the Holocene.

Neotectonics have been shown to shape the Holocene landscape of the Highlands in only a handful of localities, although there is substantial evidence of fault scarps, slope destabilisation and stream diversion from Finland and Sweden where continuing isostatic rebound of $1-11 \text{mm/yr}^{-1}$ (Pan & Sjøberg 1999, RWMAC 1988) is the legacy of land surface depression beneath the Fennoscandian ice sheet. Most work indicates that faulting to accommodate recovery from isostatic depression beneath the Devensian and Younger Dryas ice sheets in Scotland was restricted to the Lateglacial and very early Holocene (Fenton 1991, Firth et al 1993, Holmes 1987, Ringrose 1989, Sissons and Cornish 1982). A very small number of studies identify faulting in the mid and late Holocene (Davenport et al 1987, Fenton 1991, Ringrose 1989). The sites described above appear to offer further evidence of fault-driven slope disruption.

Groundwater saturation with the iron and manganese which precipitated as cement, preceded folding of the laminates. At the Liatrie fan, as at Allt Crumaidh, Allt Innis na Làrach, and in the Loch Mullardoch laminates, a possible source of the cementing fluid is a sub-peat erosional flush. The presence of homogenised peat within the cemented zones at Liatrie confirms the association of peat and cementing fluids. The proposed process at Liatrie is shown schematically in Figure 4.23b. Specific conditions, including a persistent high water table, are needed for the mineral cement to have precipitated in horizontal layers in the body of the fan. Surface percolation through unsaturated fan debris would have produced diffuse staining, and an irregular pattern in the very permeable debris, since percolation would have been very rapid. However, shallow groundwater rising to permeate fan sediments, and supersaturated with Fe and Mn, could have produced horizontal layers like those preserved. The presence of OM in the

form of dispersed peat promotes precipitation (Tucker 1981). Relatively impermeable fluvial silts, muds and peats (Fig. 4.23b) provide a barrier capable of impeding rapid dilution and dispersal of the concentrated fluid, and forcing iron-rich groundwater towards the surface within the body of the fan.

The layers of cement may therefore be attributable to a water table at least 5m higher than today at Liatrie. If so, the river may have formed a ribbon lake below the 200m contour in this wide, now boggy, reach, which lies between the currently widely separated Lochs Sealbhanach and Carrie (Fig. 4.21), just downstream of the rock gorge across which the Mullardoch dam was built.

There is no evidence that the sequence in the Allt Coire Eòghainn fan reflects fault movement. In Figure 4.25, fining up alluvium, can be seen as abruptly emplaced on peat which accumulated in an environment where slope sand had been mobilised. The rising water table inferred from this sequence was followed by substantial slope instability. At the fan toe, deposits of the River Cannich stacked against coarse fan debris obscure its full thickness. As at Liatrie, such deposits represent a shallowing up sequence with clean silts giving way upwards to low energy, muddy marginal deposits in what must have been an embayment. These sediments are several metres above today's floodplain and, as at Liatrie, appear to imply that the River Cannich reached a level near the 200m contour. If the rise in water level at the two adjacent sites is correlated, the ribbon lake was at least 1.2km long in a reach where the modern river is still unusually wide. Subsequently, water level fell, leaving the fans high and dry. If the thin layer of Fe/Mn cement in the Allt Crumaidh fan has a similar origin, the length of the ribbon lake was at least 2km. Unfortunately the dam just upstream has re-routed short section of river, lowering the level of Loch Sealbhanach into which the eroded fan margin now protrudes, and preventing verification of predicted fining up alluvium at the fan toe.

In the Allt Crumaidh fan there is evidence for distal debris flows and massive soil failure. Features of the 'terracotta deposit' point to sudden failure of a substantial area of intensely iron-stained, podzolic till which had supported heath vegetation. The vegetated surface was ripped up, fragmented and dispersed in a homogenised, fluidised flow. Heather roots and peat rip-up clasts without other obvious plant debris, in conjunction with charcoal, suggest failure of a heavily burned surface.

In summary, there is evidence for slope remodelling above Loch Mullardoch accompanied by violent *in situ* disruption and transport of materials on a low-gradient,

apparently stable slope. It was probably triggered by Lateglacial seismicity, reactivated during the Holocene. There is sufficient evidence for lake level changes to encourage further investigation. At Liatrie, Allt Coire Eòghainn and Allt Crumaidh, there is evidence for slope destabilisation associated with increased humidity and possibly higher water tables, after a period of woodland growth.

4.6   STRATHDEARN: EVIDENCE FOR HOLOCENE LANDSCAPE EVOLUTION

Strathdearn (425 km^2 from the headwaters to localities described below) is the central portion of the faulted Findhorn valley in the Grampian Highlands (Figs. 4.1, 4.30). Annual rainfall is less than 1000 mm (930mm at Aviemore c. 20km to the south). Sites described below are in the middle reaches near Tomatin, on either side of the arterial road (the A9).

Bedrock is steeply dipping biotite gneiss, with psammitic and pelitic schists and intrusions of pre- and syn-deformation granites (Stephenson & Gould 1995). North-east (downstream) of the A9 a narrow graben of Devonian conglomerate abuts granite. In contrast to the long, rugged, gullied slopes of the west coast and central Ross-shire, the valley walls rise steeply two to three hundred metres above the floodplain to rounded summits which coalesce in a plateau (Figs. 4.31a&b). An extensive mantle of grus is attributed to deep Tertiary weathering (Stephenson & Gould 1995).

Deglaciation was probably complete by about 13000 BP, although severe periglacial conditions may have affected the plateau to the south during the Younger Dryas Stade (Hinxman & Anderson 1915, Sissons 1974, 1979, Young 1980). The deglaciation history of the Streens gorge area just downstream of the logged sites was reinterpreted by Auton (1990), who noted that its Holocene history remained unexplored. One of the most complete Holocene peat sections so far described in Scotland (Birks 1975) is on the plateau above the River Findhorn at the source of a small tributary, Allt Fèithe na Sheilich, between Streens and the A9 road.

As discussed in Section 4.2.1.1 (Strathdearn), deglacial and structural gullies as previously defined are very infrequent. Coarse cones and fans, some of which have crudely incised fluvial terraces, appear to be devoid of buried organic horizons, perhaps due the dominance of fluvially- rather than debris-dominated processes since their initial (?Lateglacial/very early Holocene) formation. Dendritic drainage of thick plateau drift and peat reflects the absence of strong channelled flow. The contrast with the

gullied slopes associated with warm-based ice to the mountains to the west, is striking.

The floodplain of the River Findhorn between Tomatin and Streens (Fig. 4.30) was investigated during three geotechnical studies for road building, hydroelectric dam construction and water extraction. A large buried channel, with rock-cut benches visible (although not identified as such) in a geophysical study (Scottish Development Department 1973), underlies the area on either side of the A9 road. The modern river is reworking parts of the surface of a stack of up to 35m of sand and gravel with cobbles and boulders (Scottish Hydro-Electric boreholes 1957/8; Scottish Development Department 1973). A BGS borehole adjacent to Tomatin (NH 803273) did not reach bedrock, but confirmed buried valley fill of (from the base up): 20m well-sorted medium sand, 6m silty clay, 5m gravel, 1m peat/silt (Robins 1990). Scottish Hydro-electric boreholes within 1km downstream proved a continuation of the buried glacifluvial channel with sands and gravels up to 28m thick over weathered bedrock. The buried channel is floored by gneiss and schist which varies from 'slightly weathered' to 'rotten' (i.e. grus) down to a depth of at least 10m below buried rock-cut benches. From SDD (1973) borehole data it can be inferred that grus thickness exceeds 35m. Tentatively, the presence of a sheet of thick, permeable grus, with weakly channelled drainage beneath peat, may explain the coherence of the drift cover, and the apparent lack of Holocene fan aggradation by debris-dominated flows.

None of the boreholes encountered any buried organic horizons on the floodplain in the 5km between Tomatin and Ruthven (Fig. 4.30). Terraces with 3-5m risers fringe the modern river where it is currently reworking about 750m of a 1100m wide glacifluvial floodplain in the vicinity of Invereen (NH798315) (Young 1980) (Fig. 4.30). They are composed of matrix-supported, imbricated boulders visible in rare non-degraded exposures. There is no evidence of alternating slope and valley floor sediment wedges with bounding organic horizons, as in Strathfarrar and Glen Cannich. Sandy or silty Holocene fills deposited on the thick glacifluvial gravels and stratified sands, form two sub-facies: a massive structureless flood deposit forming a 2.5m fill terrace of fine sand and silt, as at Inverbrough, and overbank deposits up to 2m thick, with alternating mineral and organic layers representing gradual accretion, as in Section F3 and further downstream of Ruthven.

Upstream, at Dalarossie (NH76432408), some large, convex elongate, matrix-rich, boulder bars are buff coloured in contrast to, texturally and mineralogically similar, but

redder-coloured material found in adjacent glaciofluvial terraces. They also contain boulder to pebble sized grus particles which crumble under finger pressure. These could not have withstood collisions during bedload transport and must have been carried in coherent grain flows below water during very large floods. Archaeological remains provide an indication of their age. Hut circles on a terrace fragment at the same elevation at Dalarossie are 'of Bronze or Iron Age affiliation' (Highland Region Archaeological Sites and Monuments Records) demonstrating that floods of that magnitude have not occurred in the last 2000 years as a minimum, and possibly for much longer, notwithstanding the Great Flood of 1829 Lauder (1830).

4.6.1 Logged sections

Exposures of Holocene slope stratigraphy containing long datable sequences were not found despite extensive search, but three floodplain sections and one slope peat section all in the reach immediately downstream of Tomatin were logged.

4.6.1.1 The floodplain

Sections F1 and F2 (Fig. 4.32) sample peat and alluvium on the glaciofluvial terraces. Wood fragments are found throughout F2. Section F3 shows the sequence below part of the modern floodplain. Nearly 1m of overbank sands and silts with peaty stringers have a basal iron pan. They record frequent floods with a suspended load of sand not seen immediately before or after, since they have sharp upper and lower boundaries. They overlie more than 1m of peat whose lens-shaped exposure in the river bank indicates infill of a palaeochannel which ran at nearly 90° to the modern channel. Its original width cannot be determined from surface features due to bank erosion and deep, rapid river flow, but it appears to be a meander cut off during a large flood. Iron pan such as that beneath the overbank sequence takes hundreds of years to form, and basal peat from a column in excess of 1m thick is likely to be $\geq 10^3$ years older than the base of the overbank sequence. The abandonment of the channel in which the peat sits is older still and must have a minimum age well in excess of 1000 BP.

4.6.1.2 Slopes

Extensive areas of pine stumps buried beneath peat on the upper glacifluvial terrace in the Invereen area and on the plateau margin in the Streens area (NH 870379) have been found, but await investigation. Thick (>1.5m) overlying peat, whose thickness may have been reduced by peat cutting, suggests their age could extend to thousands rather

than hundreds of years.

The source of Allt na Fèithe Sheilich at 595m AOD (Fig. 4.30), is a key site for interpreting aspects of Holocene environmental change in the region (Birks H. H. 1975, Birks H. J. B. 1993). Combining pollen stratigraphy with dated *in situ* tree stumps, Birks (1975) described the palaeoecology of an almost complete Holocene succession in 3.25m of peat, dating back to 9800 BP. By 7500 BP pine-birch woodland was the dominant vegetation, but by 6900 BP *Sphagnum* had begun to replace pine.

In order to investigate slope evolution, a section was excavated (as far as practicable below the water table), adjacent to Allt na Fèithe Sheilich, at an altitude of 350m (Fig. 4.33). Enclosed hollows affected by very local changes in drainage were avoided. The section exposed >2m peat with a small *in situ* unidentified tree stump near the base, and three horizons where tree fragments were preserved. All the peat was black, non-fibrous, well humified, and devoid of visible mineral sediment. The *in situ* stump sits on peat, and confirms replacement of local woodland by peat at a depth of c. 1.8m. In the absence of pollen stratigraphic evidence, the presence of wood in only parts of the overlying peat column could reflect either fluctuating woodland cover, or oxidising conditions unfavourable to its preservation for most of the time represented. The continuous presence of wood fragments in the peat column on the lower glaciofluvial terrace some 50m downslope (Section F2), gives some credence to preservation as an important control.

4.6.2   Discussion and summary

The lack of lower slope landforms containing buried horizons limited investigation of Holocene stratigraphy in Strathdearn. Early debris cones are crudely fluvially incised or extended, but, with the exception of a single buried (or slipped) immature podzol on a steep slope made visible in a roadside layby (NH 746213), no evidence was found of debris flows occurring after periods of stability. The lack of fans aggraded by debris flows in Strathdearn, and perhaps, the absence of mineral sediment in the Allt na Fèithe Sheilich section, is associated with an absence of strongly channelled drainage in rock-cut gullies. Some indication of changing conditions in the past is provided by the replacement of trees by peat, interpreted in Birks' (1975) plateau section as a response to changing humidity. Human impact on late Holocene tree cover cannot be ruled out, given the archaeological sites upstream at Dalarossie.

Floodplain restructuring in response to changes in flooding and sediment transport is evident in the vicinity of Invereen. More than one thousand years ago a substantial meander was cut off and abandoned. After it had filled with more than 1m of sand-free peat, a period of moderate flooding began, and suspensions of silt and sand spread out over floodplain vegetation at intervals frequent enough to inhibit continuous peat formation.

4.7 OVERVIEW AND CONCLUSIONS

The objectives of the preliminary survey and of the detailed investigations were to locate Holocene landforms which preserved evidence of repeated environmental change; to understand the typical geological and geomorphological parameters of such sites; to describe their Holocene stratigraphy; and to select for dating representative sites where anthropogenic triggers for any patterns of change detected could reasonably be excluded in order to simplify interpretation.

The following responses have been inferred in the landscapes described above:

- channel avulsion
- episodic fan aggradation
- subsurface-peat degradation
- blanket peat erosion
- soil erosion
- enhanced slope wash
- till failure
- homogeneous fill terrace formation
- gradual floodplain accretion
- intercalation of slope and floodplain sediments at times of rising and falling water table
- temporary conversion of a section of river to river lake dimensions
- changes in peat humification and wood fragment preservation.
- evidence of fires on slopes.

Judging by the widespread and variable nature of the buried organic matter found, and processes inferred during the field survey, slope and valley floor activity has been common across the region during the Holocene. None of the evidence from the catchments discussed suggests that the east-west precipitation gradient emphasised or attenuated processes of landform evolution, although geological and glacial history appear to have influenced pathways of Holocene development through the inherited

style of slope drainage.

Evidence from the sites investigated in some detail is of rising water tables and increased slope instability after peat development, pine woodland spread, and intense soil leaching, continuing into the late Holocene. The expansion of rivers to form ribbon lakes, and a subsequent water table fall, have been suggested in Glen Cannich, and Strathfarrar. Lake level rise in the adjacent Glen Affric has also been detected (R. Tipping, University of Stirling, pers. comm.). Across the region, modern soils have a sharp lower boundary, with gradual accretion sediments which record a period of high frequency, sand-bearing, overbank floods immediately preceding the deposition of thicker sands and silts in which modern soils are developing. Overbank sequences in turn abruptly replace earlier strata, deposited when the frequency of flooding was lower, but when water tables were rising, peat had replaced pine woodland, and slope soils with burned surfaces were suffering mass failure. The close spatial proximity of eroded soils and failed till with Fe-Mn cemented fan sediments, suggests that erosion beneath the peat cover was associated with slope destabilisation and soil erosion. Such events are probably not rare on a Holocene time scale, since evidence of cementation was found at four separate sites. Subsurface peat disintegration associated with changes in channelled flow within the body of peat rather than run-off *per se*, is a plausible direct trigger for some slope failures, with excess water acting primarily on peat stability, and through positive feedbacks, on the mineral substrate.

Postglacial faults have reactivated lines of weakness whose origins go back to Caledonian mountain building, and many valleys lie in fault zones excavated during successive glaciations. Although only a very small number of faults is known to have caused surface disturbance after the Lateglacial, exposures along the shoreline of Loch Mullardoch suggest that bedrock brecciation, debris flows on low angle slopes, and soft sediment liquefaction could be locally important outcomes which are, in terms of landscape evolution, more significant than small fault scarps or stream diversions. Interpretation is complicated by the fact that not all the features of sudden ground acceleration are exclusively attributable to seismicity.

Characteristic geomorphological settings likely to yield useful data about Holocene stratigraphy incorporate

- a direct link between sediment source and sediment sink
- limited capacity for sediment storage between source and sink

- a sediment sink protected from fluvial erosion
- a potential for renewed sediment release
- a sediment and groundwater source area protected from anthropogenic modification
- minimal vegetation in the zone of infiltration.

An example of a source which embodies these features is a long bedrock gully system originating in a steep rock face, and terminating in a lower slope fan. Mid or lower slope bedrock depressions, and the margins of aggraded valley floors where slope and river (or lake) sediments interdigitate, form additional sediment stores. However, investigation of slope evolution in the Northern Highlands is circumscribed by accidents of preservation and exposure in an erosional mountain environment where difficult access and coarse sediment can make coring impractical. So even in theoretically favourable localities, access to the history of sediment cycling and landscape change is restricted. This may lead to underestimation of the true extent of Holocene land surface changes.

The preliminary site survey indicates that some areas are inherently more responsive to subsequent slope and valley floor remodelling than others, due to conditioning of the landscape by geological and glaciological processes, although the number of variables prevents a simple classification of 'robust' and 'sensitive' sites. Key system elements through which matter and energy are transferred, are water, rock, sediment and blanket peat. Perturbations which drive changes of state, producing new landforms and land surfaces, require variations in the state of one or more of these elements.

4.7.1 Selection of locations for dating

Both of the localities selected for dating (upper Gleann Lichd and a slope in lower Gleann Chorainn) are in small catchments with steep slopes which are in general less affected by long term sediment storage than large, complex catchments. This minimises the problem of long lag times and complex processes intervening between generation and mobilisation of unstable sediment and lower slope aggradation. Both localities have steep, rocky source areas directly affected by changes in runoff, infiltration and rates of rock weathering. These sites embody the general characteristics of locations which are sensitive to landscape altering processes triggered by groundwater, and are suitable for evaluating a climatic trigger.
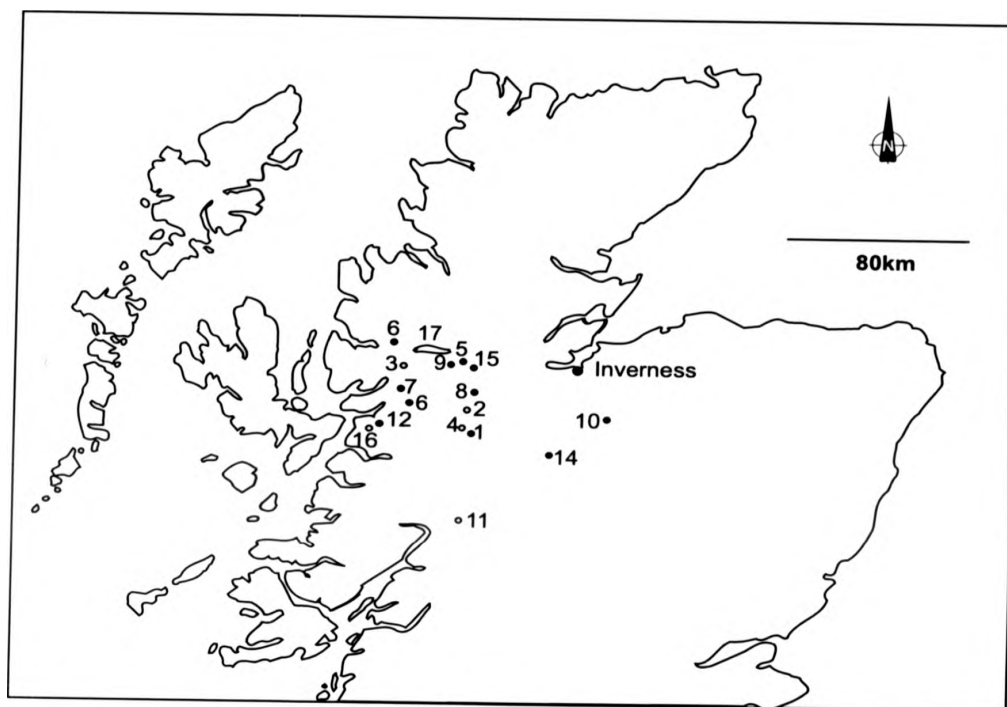
FIGURE 4.1
Locations of catchments surveyed

1	Glen Affric
2	Glen Cannich
3	Glen Carron
4	Gleann a' Choilich
5	Gleann Chorainn
6	Glen Docherty
7	Glen Elchaig/Srath Duilleach
8	Strathfarrar
9	Gleann Fhiodhaig
10	Strathdearn (middle Findhorn valley)
11	Glen Gloy
12	Gleann Lichd
13	Glen Ling
14	Glen Markie
15	Glen Orrin
16	Glen Shiel
[17	Loch a' Chroisg]
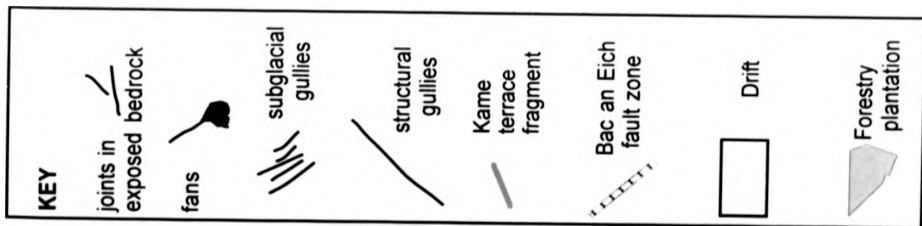
FIGURE 4.2

Air photo trace: Gleann Chorainn

v1 Section GChV1 (Fig. 5.3)   v3 Bac an Eich fan (GChV 3, Fig. 5.3)

v2 Section GChV2 (Fig. 5.3)   v4 Creagan a' Chaorainn paleogullies (Fig. 5.6)

KEY

joints in exposed bedrock

fans

subglacial gullies

structural gullies

Kame terrace fragment

Bac an Eich fault zone

Drift

Forestry plantation
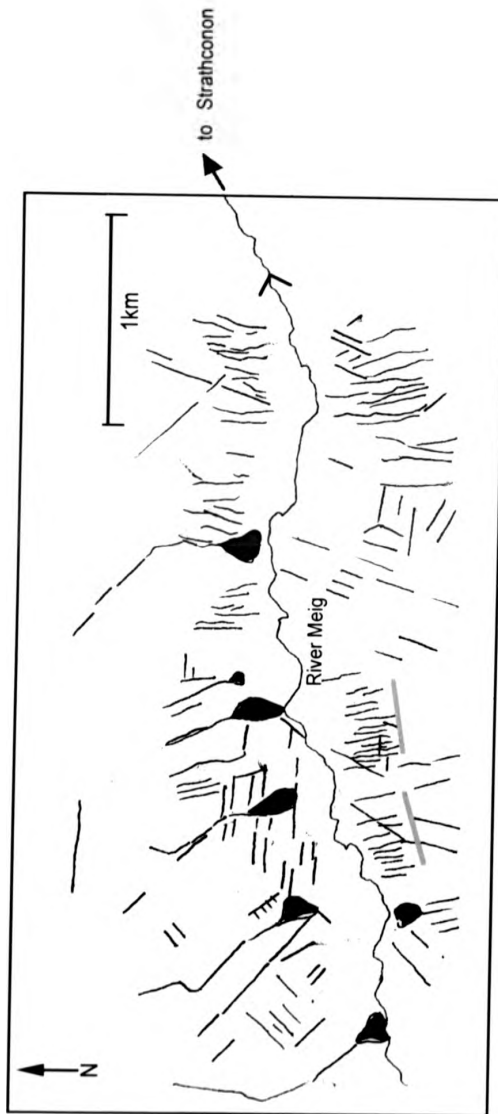
FIGURE 4.3
Gleann Fhiodhaig air photo trace

Key as for Figure 4.2

FIGURE 4.4
Looking south across Loch a' Chroisg

Parallel, subglacial gullies dissect the far slope.
Crescentic, retreat moraines overlie their bases (extreme lower right).
A well-defined raised shoreline can be seen in the middle distance.
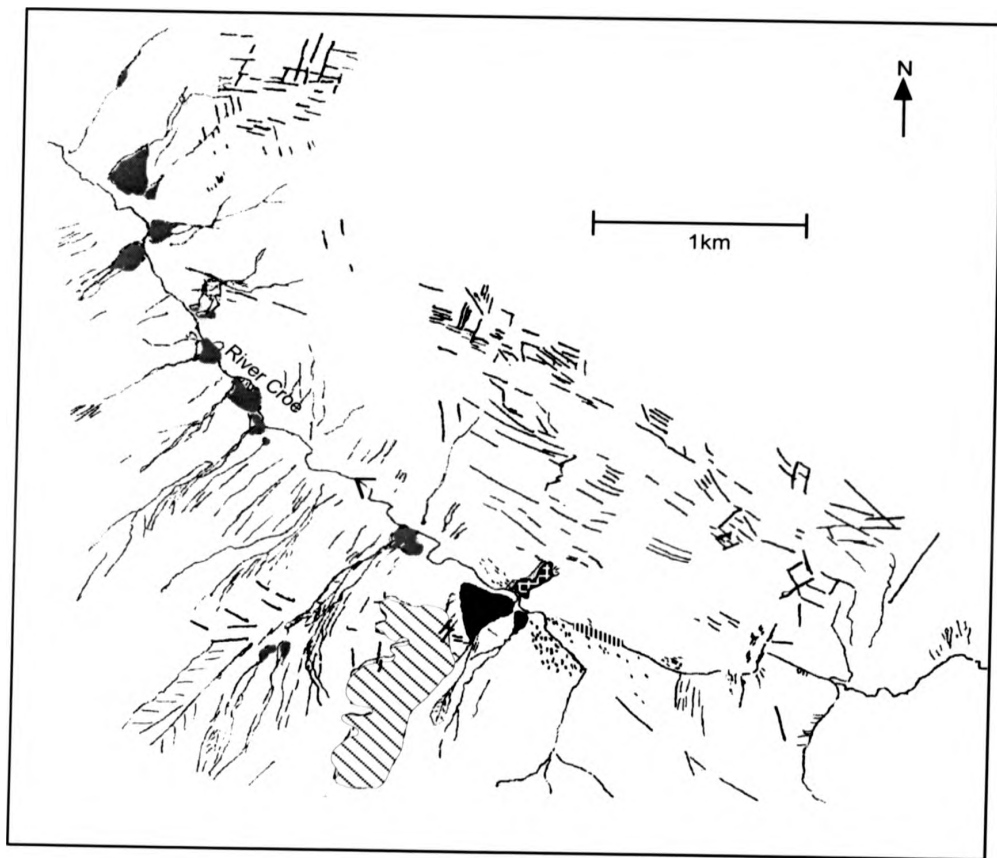
FIGURE 4.5
Upper Gleann Lichd air photo trace

Key as in Figure 4.2

ADDITIONAL KEY

▨	Aligned moraines
▦	Till failure
■	Dated fans
■	Undated fans
▨	Landslip (?pre-Holocene)
▪▪▪▪▪▪	Subglacial rock gorge
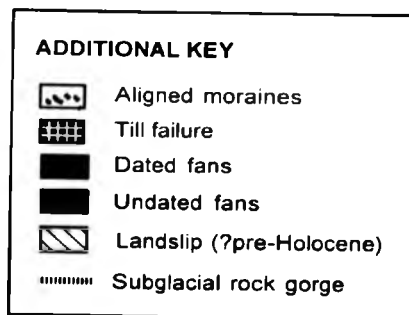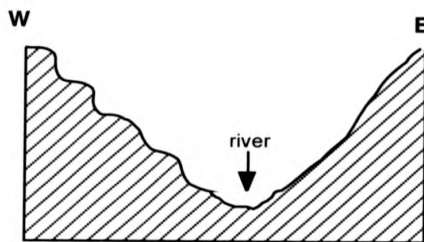
FIGURE 4.6
Schematic diagram to illustrate the effect of dipping rock structures
on valley wall stability.
On the west valley wall, benches and ledges interrupt sediment and water flow.
On the east wall, long, slope-parallel structures reduce sediment storage.

KEY

Dipping bedrock

Direction of dip parallel
to ornament
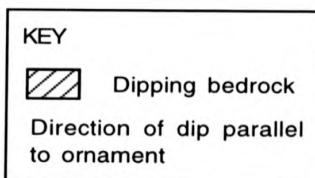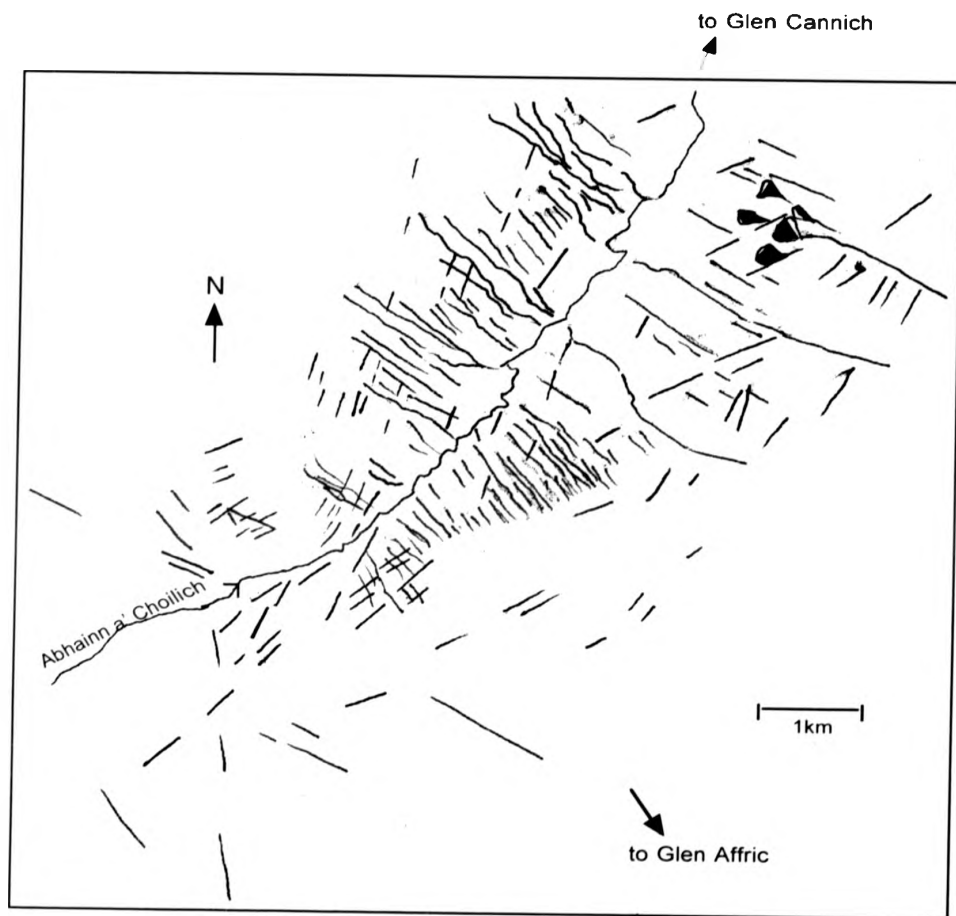
FIGURE 4.7
Gleann a' Choilich air photo trace

Key as for Figure 4.2
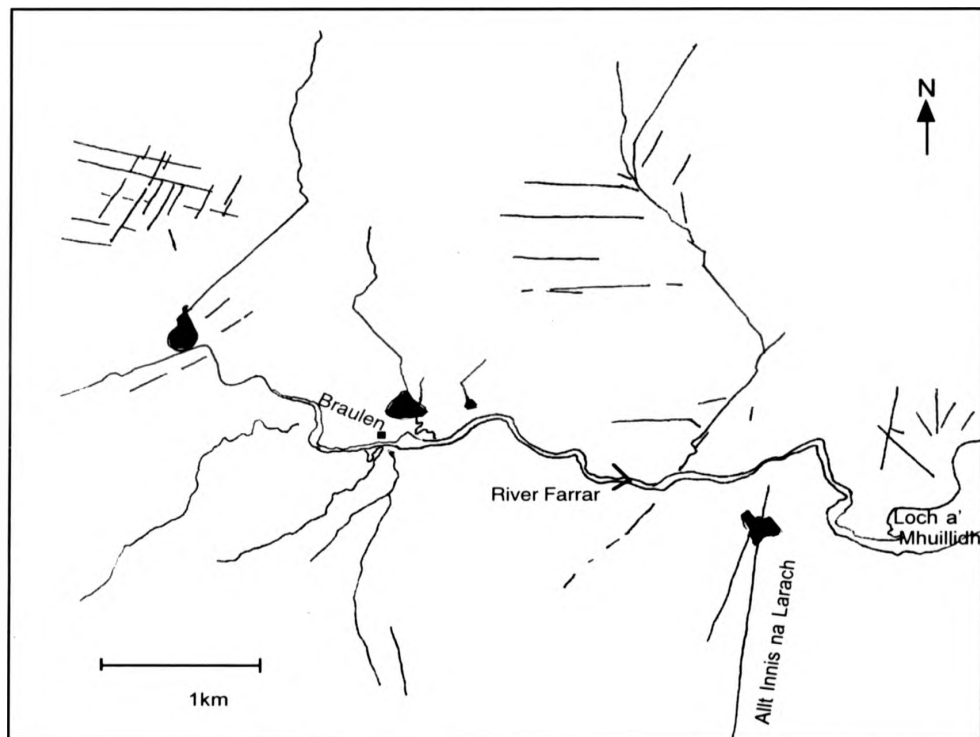
FIGURE 4.8
Strathfarrar air photo trace

Key as for Figure 4.2
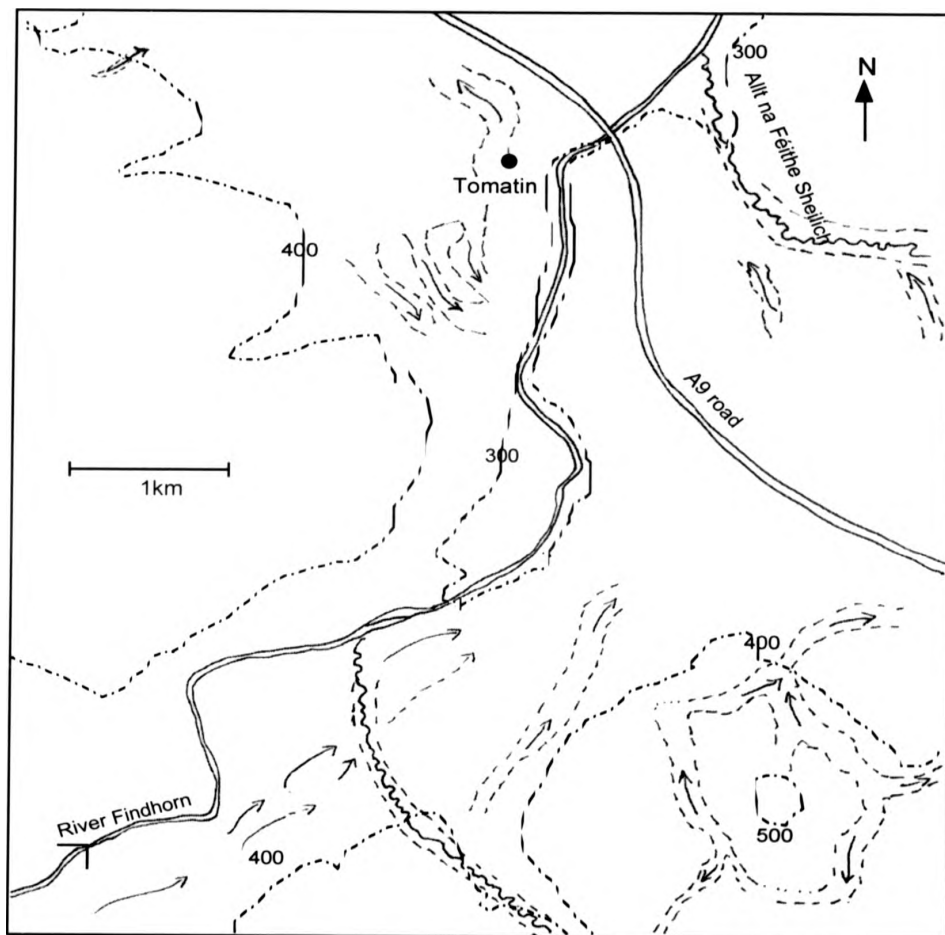
FIGURE 4.9
Air photo trace showing meltwater channel
control of Holocene drainage in Strathdearn

KEY

↗ meltwater flow
and channels

〰 Holocene drainage
streams contained within
meltwater channels

(500) height in metres
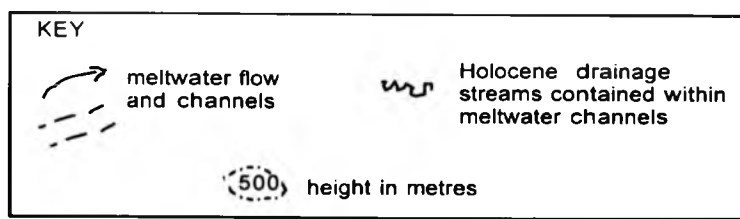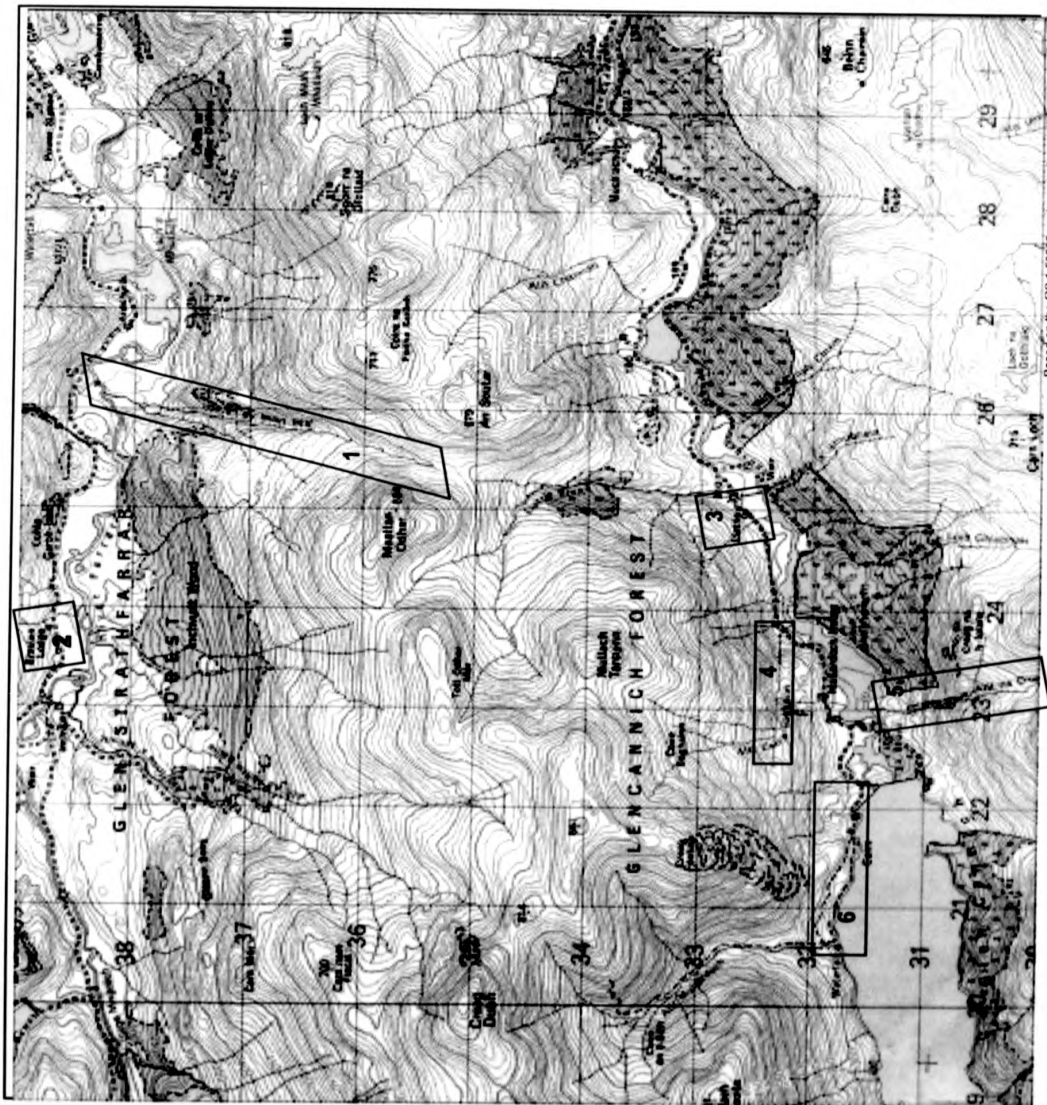
1 Allt Innis na Larach

2 Braulen fan

3 Liatrie fan

4 Allt Coire Eoghainn fan and fault trace

5 Allt Crumaidh fan

6 Shoreline of Loch Mullardoch: Benula Lodge boathouse to Allt Mullardoch

Figure 4.10
Strathfarrar and Glen Cannich: Location map

Based on the OS 1:50000 scale map, Sheet 25 © Crown copyright
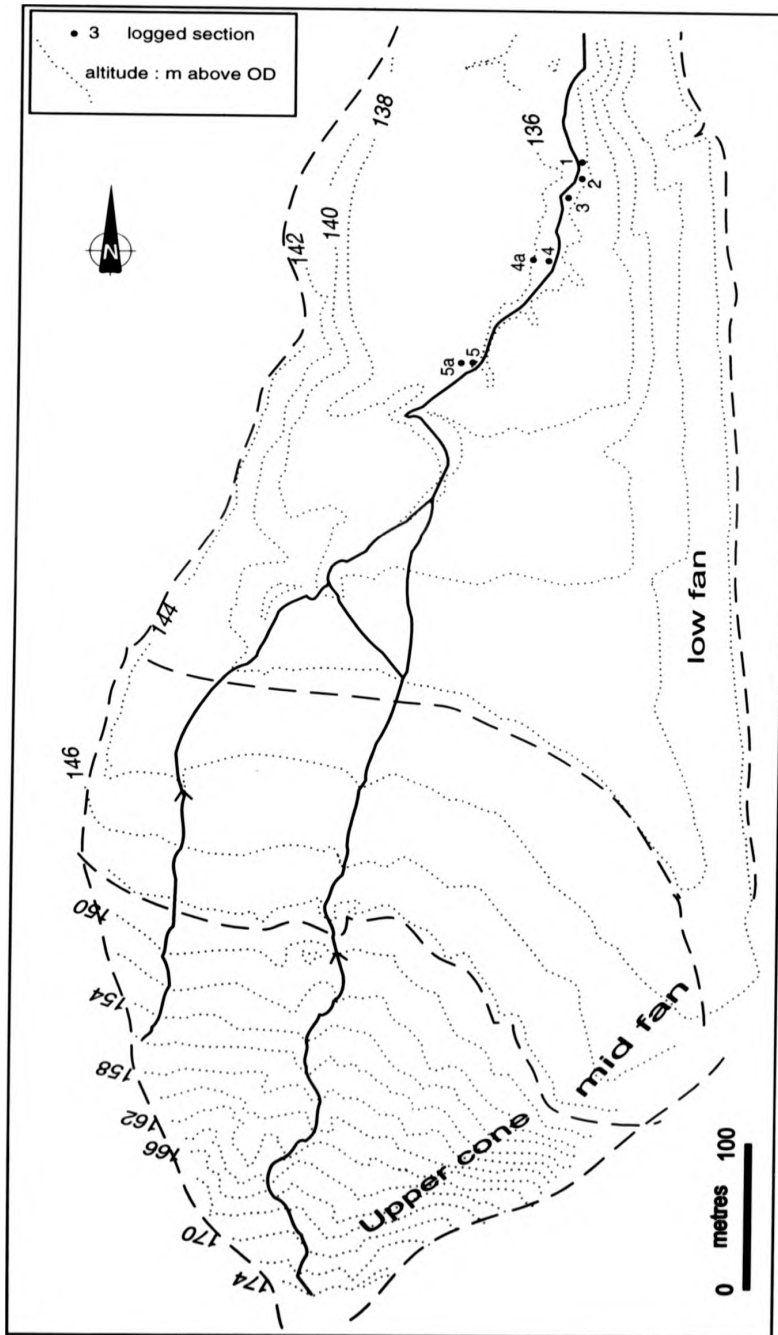
FIGURE 4.11
Allt Innis Na Larach fan, Strathfarrar
Logged sections SF 1- 5a are shown in Figures 4.17 - 4.19
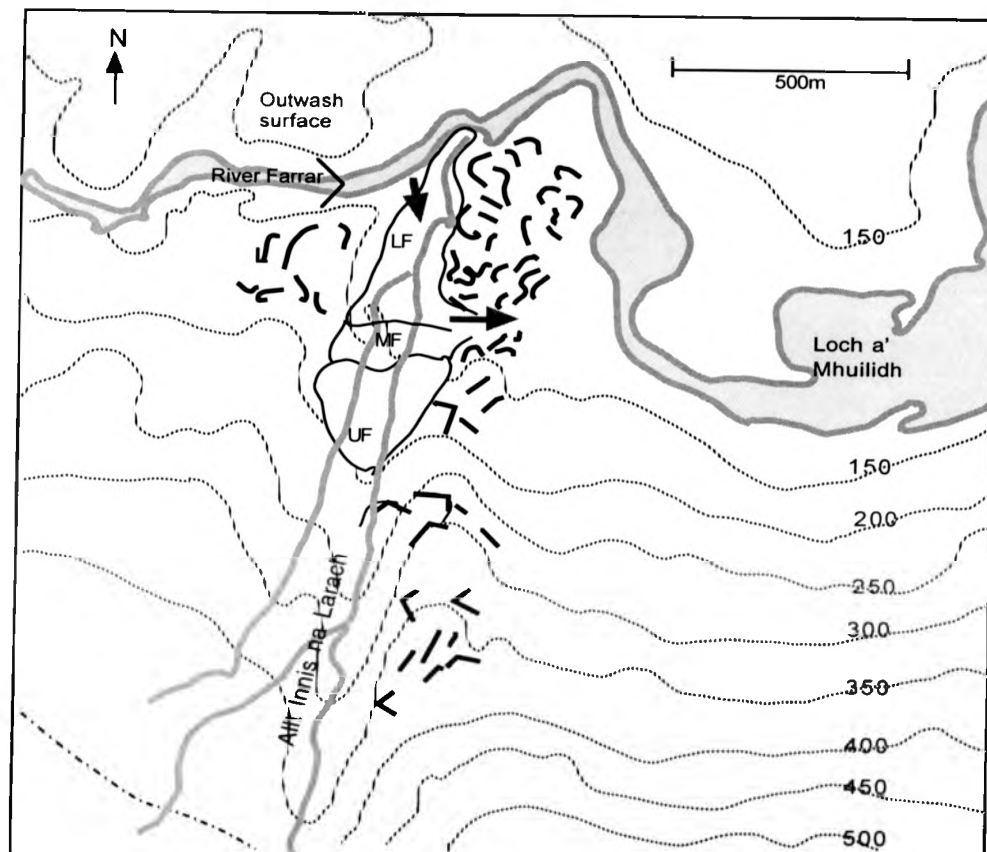
Based on OS 1:25,000 Pathfinder map 191; NH 23/33

FIGURE 4.12
Allt Innis na Larach fan

KEY

·· 200	contours/m AOD
	stream/river/loch
outcrop symbol	outcrop
→	inferred palaeo-flow

Approximate area of:

UF upper fan

MF mid fan

LF lower fan
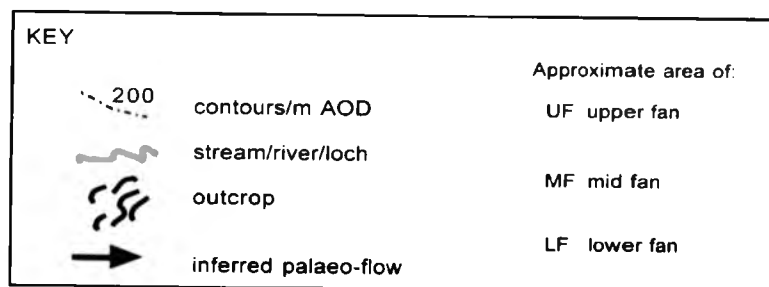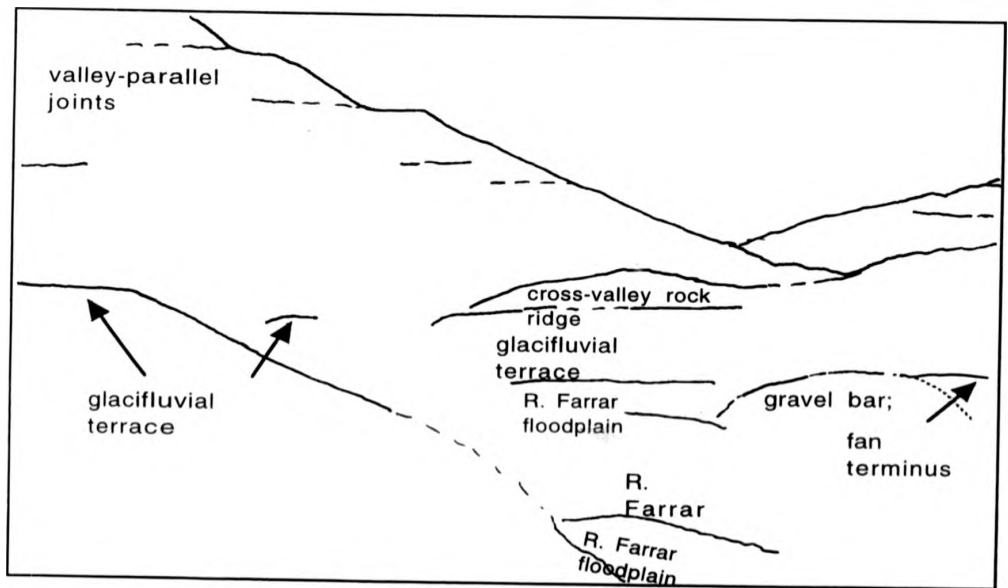
FIGURE 4.13b
Interpretation of Fig. 4.13a

FIGURE 4.14

The upper fan, Allt Innis na Lararch, showing two (?three) abandoned, incised surfaces, at successively lower angles
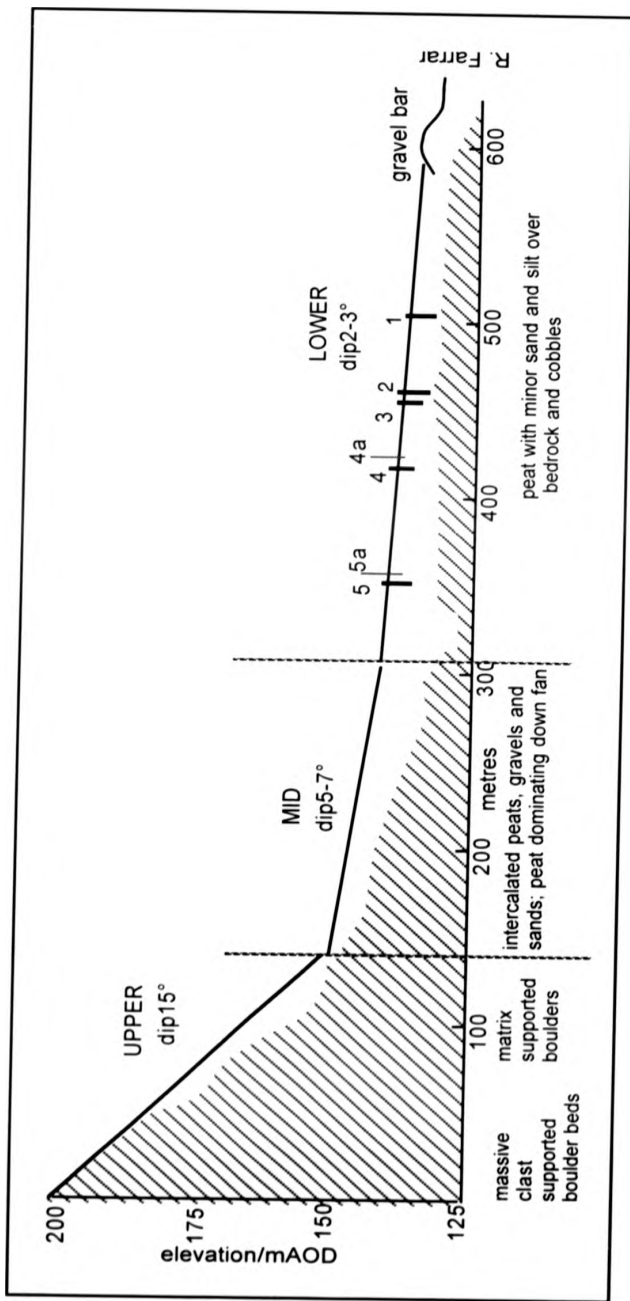
F1GURE 4.15

Schematic long section: Allt Innis na Larach Fan

Vertical exaggeration = 2.7

KEY

▎ logged Sections 1 -5 in the bank of the dissecting stream

▏ augurs on the undissected fan surface
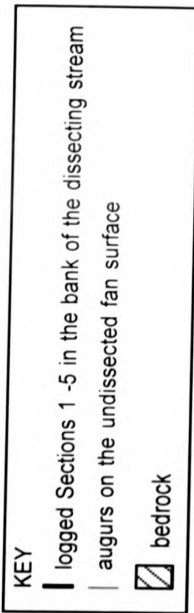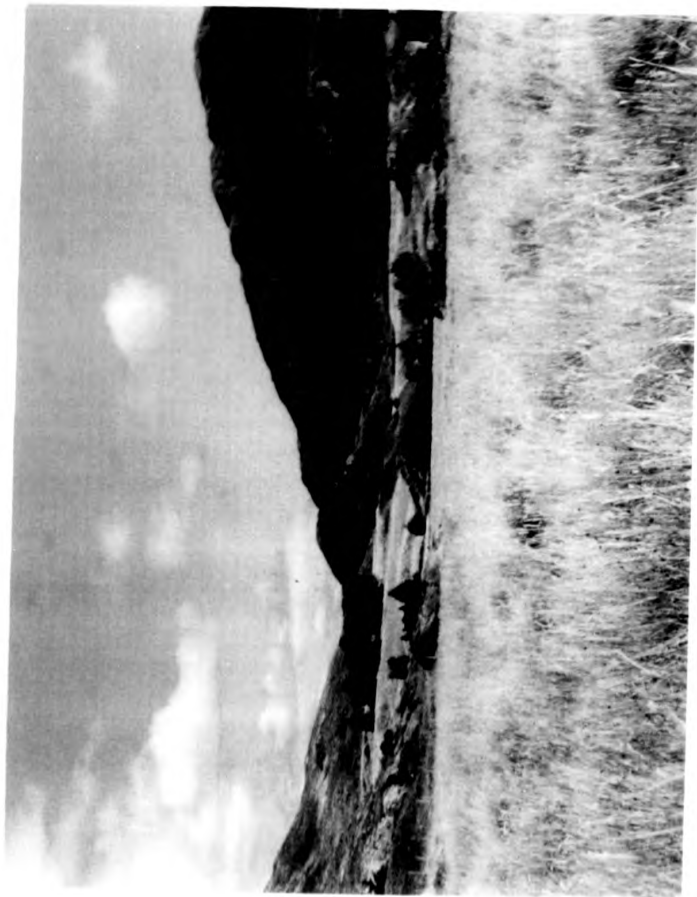
▨ bedrock

FIGURE 4.16

Looking across the lower/mid fan boundary to the inferred rock-walled exit of the palaeo-Farrar, which is at the end of the elongate low area on the margin of Figure 4.11.

depth/m

Fe

sparse small plant fragments
& unidentified OM

Section SF1

Section SF2

depth/m

Fe

c   grey plastic silt with variable sand
c

Fe
Fe
Fe
Fe
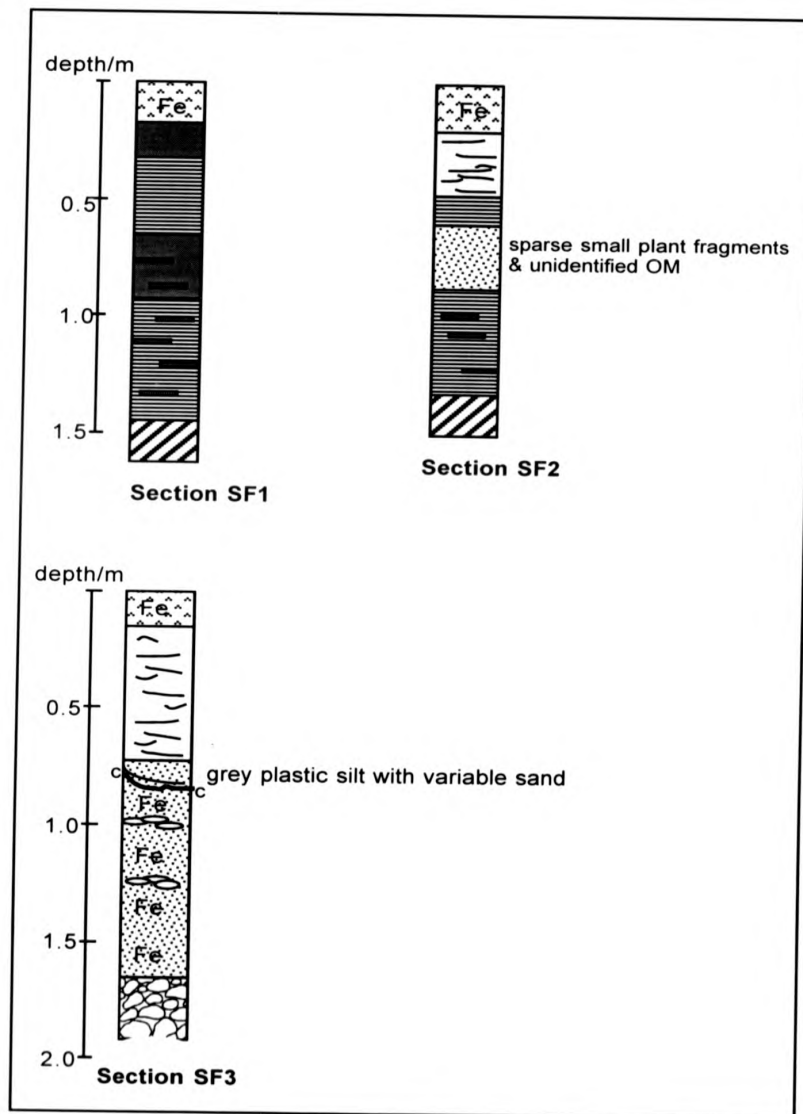
Section SF3

FIGURE 4.17
Lower fan stratigraphy, Allt Innis na Larach, Strathfarrar
Section SF1 - 3

depth/m

0.5
Fe.
1.0
1.5
2.0
2.5

3m

no sample

Pa/M

no sample

- ?

(10)

Section SF4
stream bank exposure

Section SF4a
gouge augur: fan surface
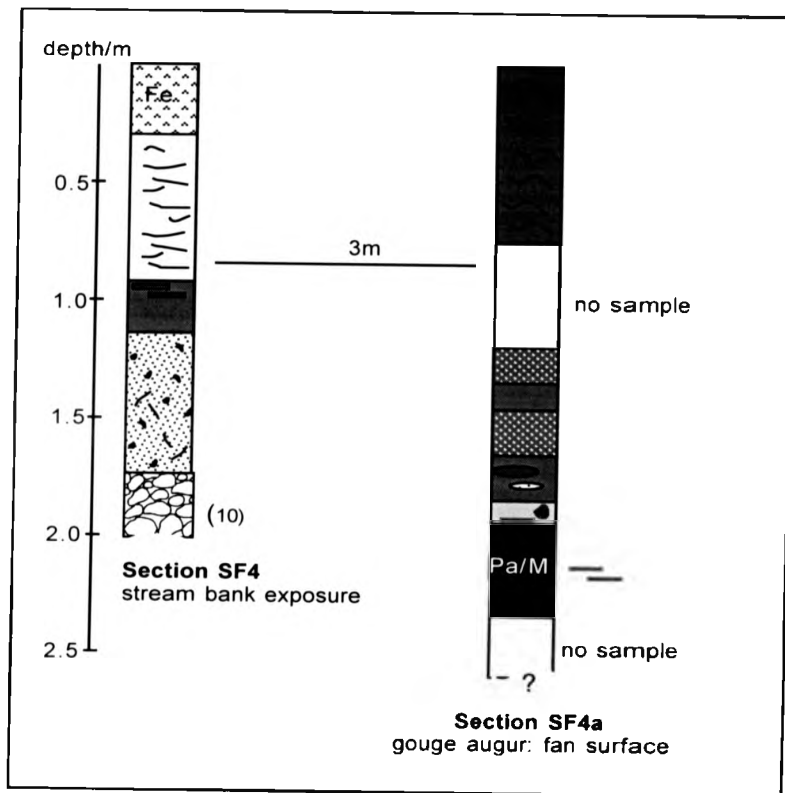
FIGURE 4.18
Sections SF4 and SF4a
Lower fan, Allt Innis na Larach, Strathfarrar

depth/m

2m

(10)

(10)

(10)

disintegrating peat

Section SF5
Stream bank exposure

disintegrating peat

Section SF5a
Gouge augur on the undissected fan surface
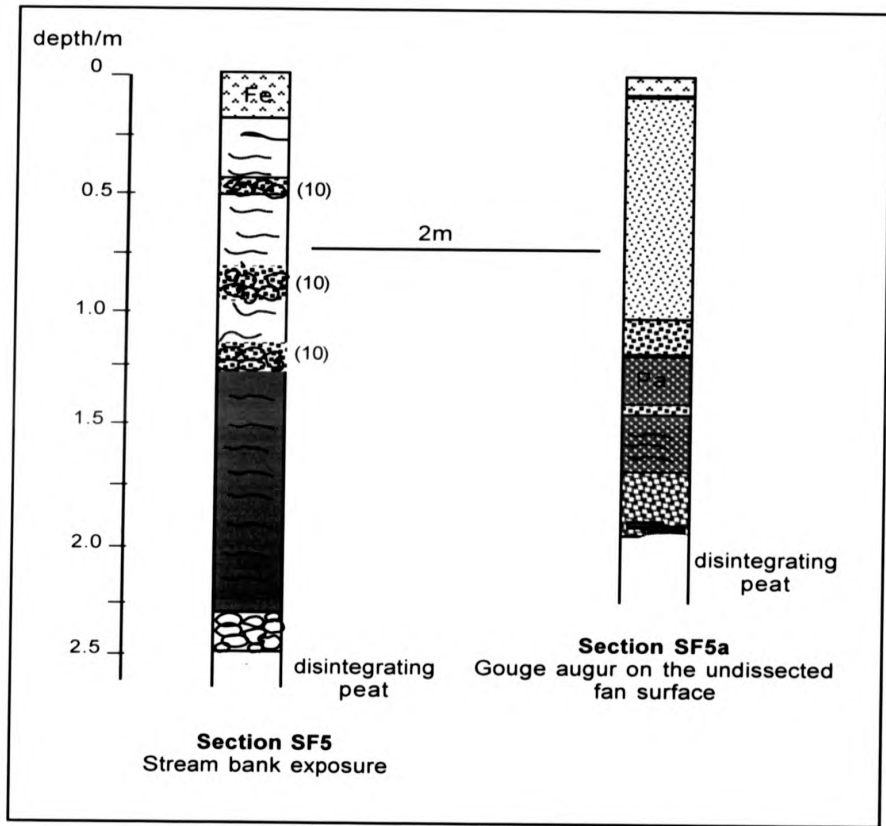
FIGURE 4.19
Sections SF 5 & 5a
Lower fan. Allt Innis na Larach, Strathfarrar

FIGURE 4.20
Iron precipitation in Allt Innis na Larach

depth/m

0.5

(10)

1.0

Calluna roots

1.5

Pinus sylvestris

?
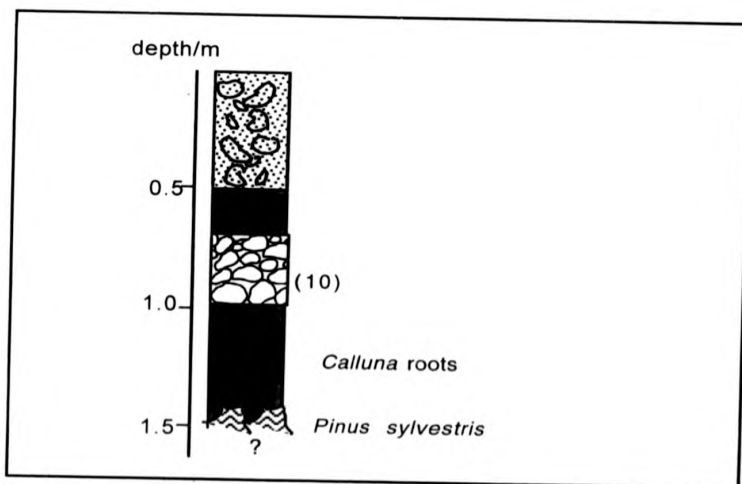
FIGURE 4.21
Braulen fan, Strathfarrar
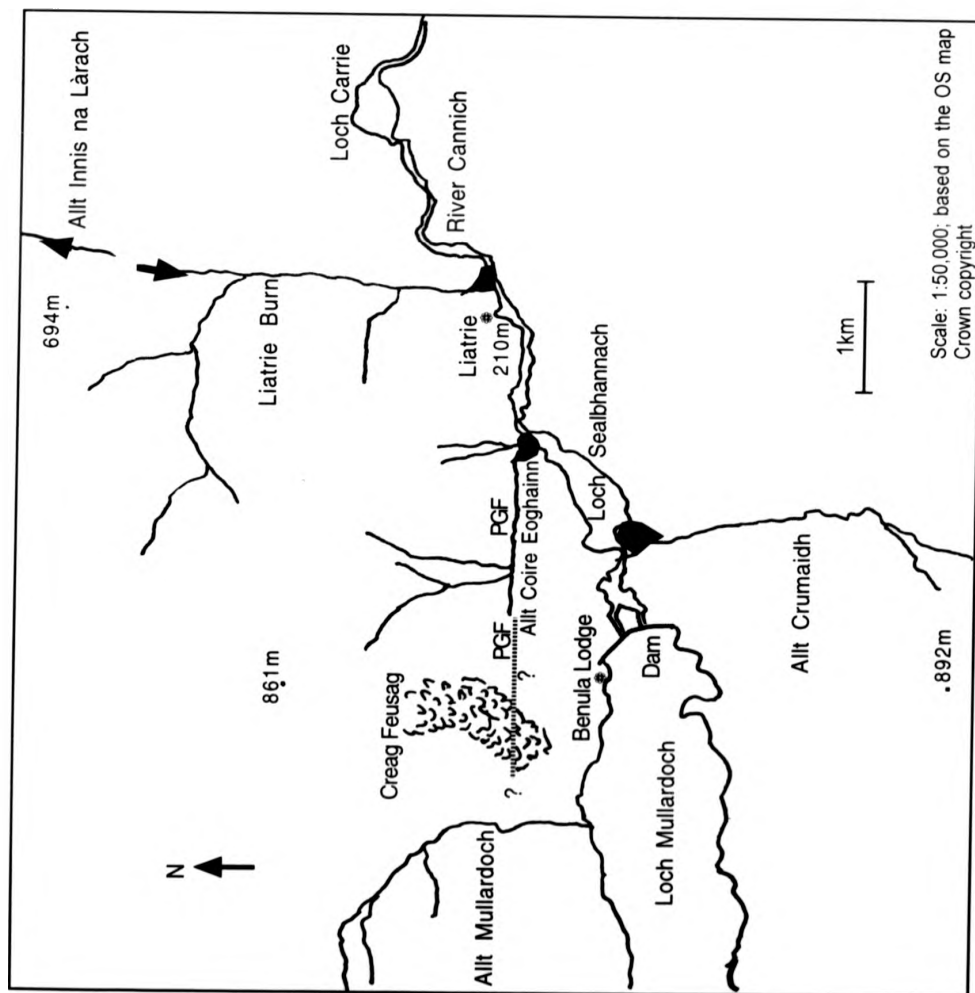
FIGURE 4.22
Glen Cannich Location Map

KEY

outcrop

PGF: post-glacial fault

fan

Scale: 1:50,000; based on the OS map
Crown copyright

1km

N

Allt Innis na Làrach

694m

Liatrie Burn

Loch Carrie

River Cannich

Liatrie
210m

861m

PGF

PGF

Allt Coire Eoghainn

Loch Sealbhannach

Creag Feusag

Benula Lodge

Dam

Allt Crumaidh

Allt Mullardoch

Loch Mullardoch

.892m

122

depth/m

1.0

0.5

1.0

1.5

2.0

peat erosion ?

KEY

till with blanket peat

permeable fan ⟶ subsurface flow of Fe-Mn rich water
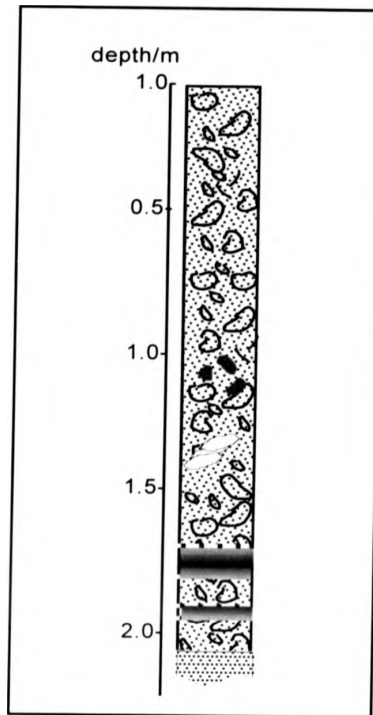
impermeable alluvium ⟶ former water table and cementation
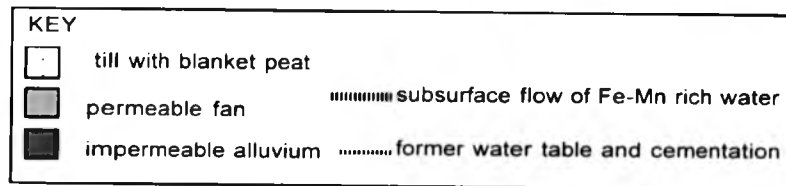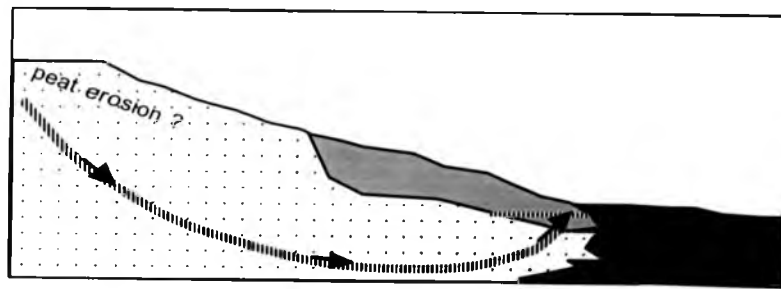
FIGURE 4.23 (b)
Schematic diagram of Fe-Mn cementation in the Liatrie fan
which consists of buff coloured, angular debris with no distal fining
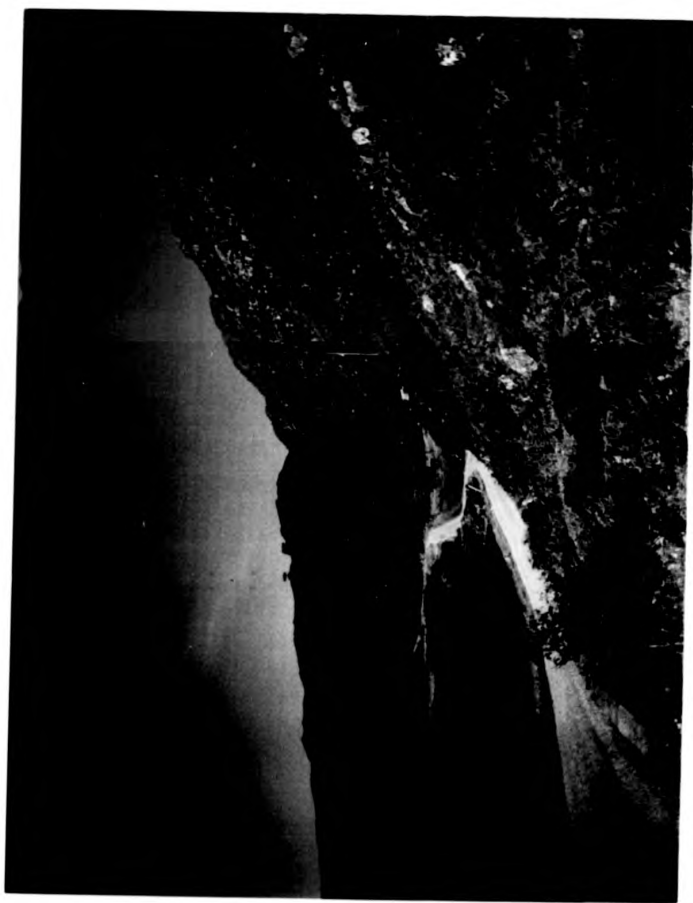
FIGURE 4.24
Allt Coire Eoghainn postglacial fault trace, Glen Cannich
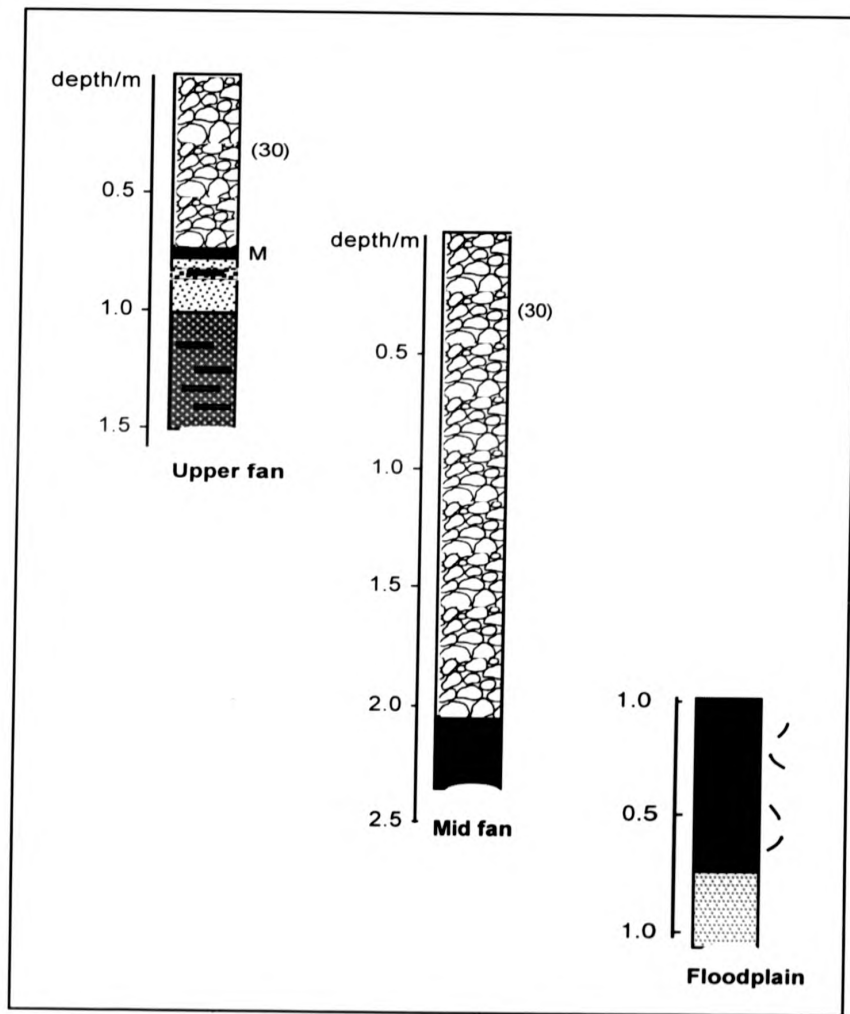
FIGURE 4.25
Stratigraphy of the Allt Coire Eoghainn fan, Glen Cannich

The floodplain mud is grey-brown, with some silt, and
occasional dark brown organic patches.
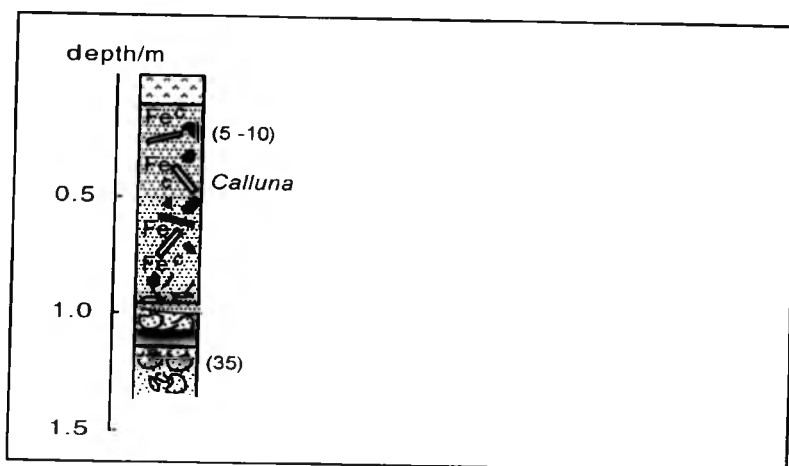
FIGURE 4.26(a)
Stratigraphy of part of the Allt Crumaidh fan, Glen Cannich
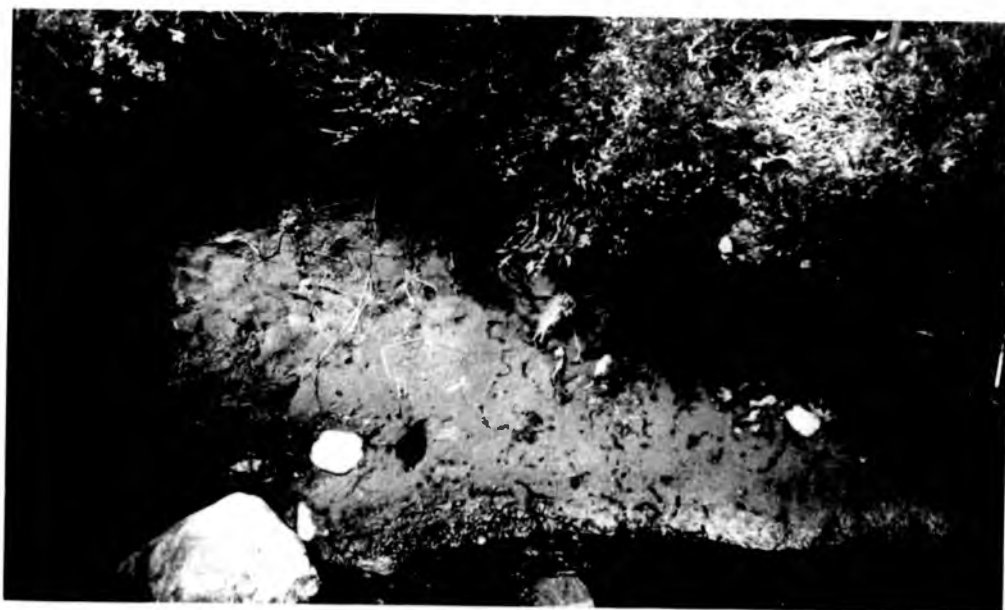
FIGURE 4.26b
The 'terracotta' debris flow, Allt Crumaidh fan.
Peat rip up clasts have pointed tails and are randomly oriented. Carbon stained 'ghosts'
of peat and Calluna' roots are widely dispersed in the sediment.

FIGURE 4.27
Eroded wave zone, Loch Mullardoch, Glen Cannich

FIGURE 4.28a
Cross-section of liquesfaction structure in stiff,stony till
on the shore of Loch Mullardoch

FIGURE 4.28b
Small scale overturned folds in stiff, stony till agjacent to
the Benula Lodge boathouse, Loch Mullardoch

FIGURE 4.28c
Brecciated bedrock on the shoreline of Loch Mullardoch

FIGURE 4.28d
Steeply dipping laminates beneath till, Loch Mullardoch

FIGURE 4.28e
Peat folded with laminates, Loch Mullardoch shoreline

FIGURE 4.28f
Chaotic folds in laminates, Loch Mullardoch

FIGURE 4.28g
The low slope above the folded laminates, by Loch Mullardoch

FIGURE 4.28h
Intense folds in laminates by Loch Mullardoch mimic folds in bedrock,
and are truncated against both the overlying debris flow and the bedrock
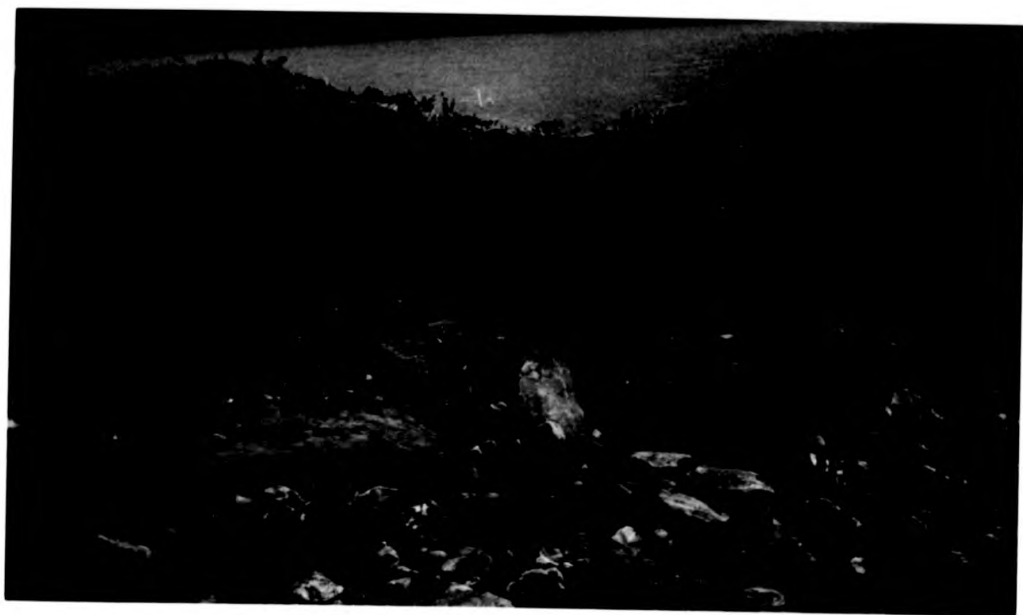
FIGURE 4.28i
Small fault and debris flow in the east bank of Allt Mullardoch

FIGURE 4.28j
Injection breccia(?) of peat and debris, Loch Mullardoch shoreline

6m

3m

a)

d)

no, or poor, exposure

KEY

brecciated bedrock with peat injected into the interstices

detached peat masses

rotated boulders

detached and rotated slices of silty-sandy laminates
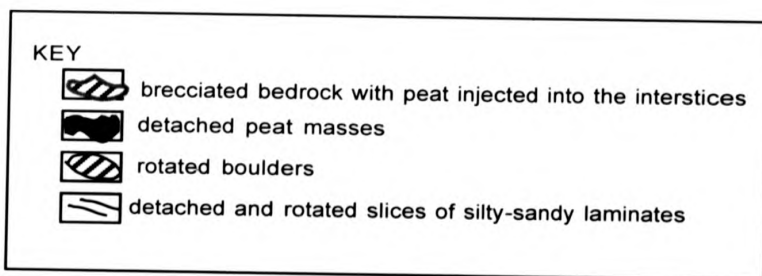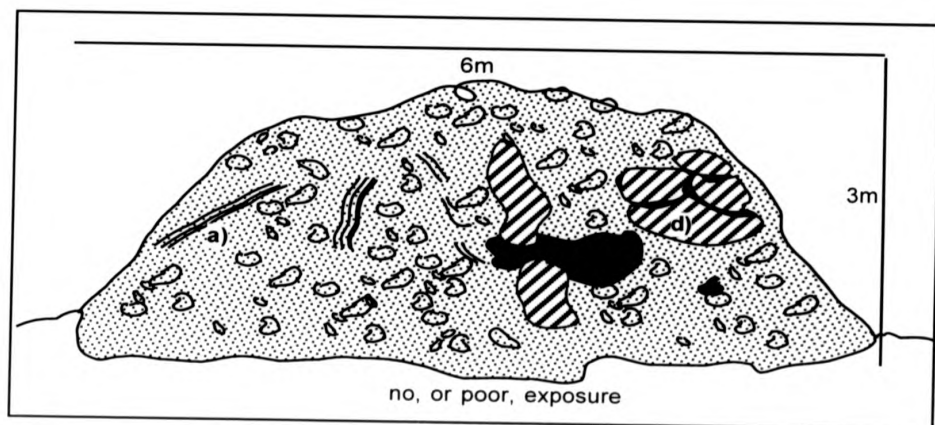
FIGURE 4.29
Apparent ice-contact mound, loch Mullardoch shoreline, Glen Cannich
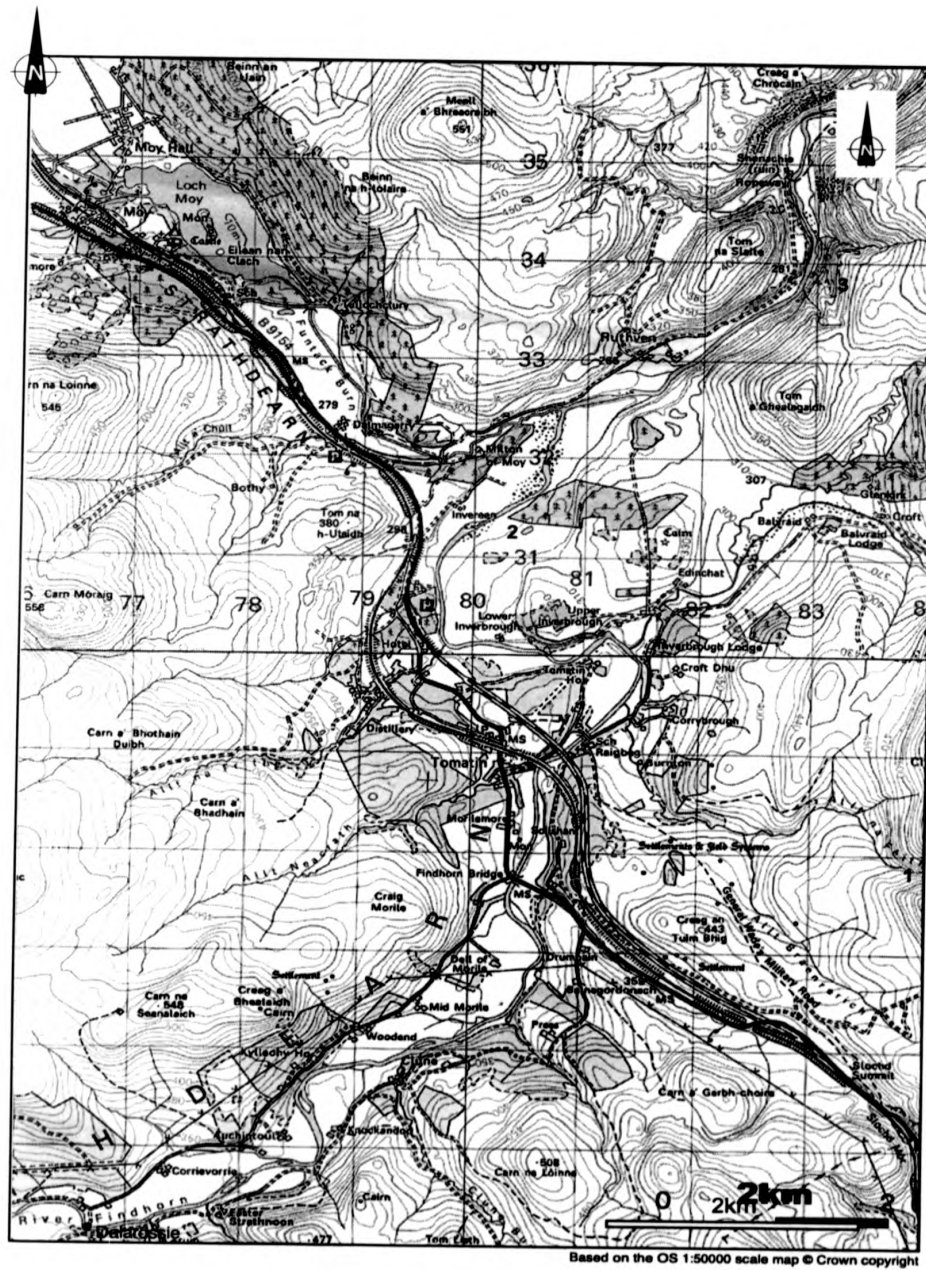
Figure 8.22   Strathdearn location map2km

Dalarossie

FIGURE 4.30 Strathdearn location map
1 Allt na Feithe Sheilich; **2** Invereen; **3** Streens gorge
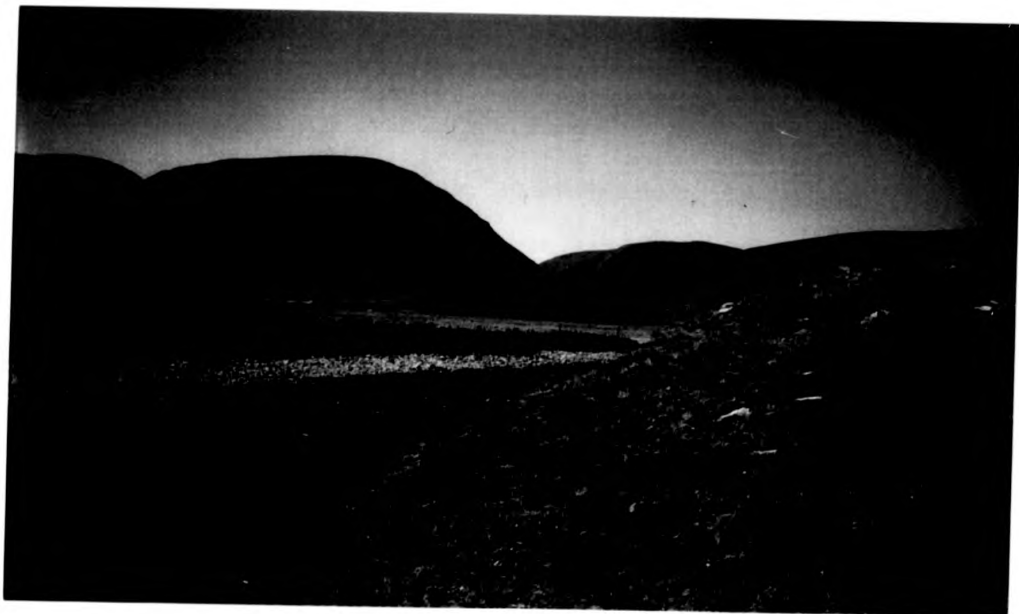
FIGURE 4.31a
Wide single gloodplain and high kame terrace, Strathdearn
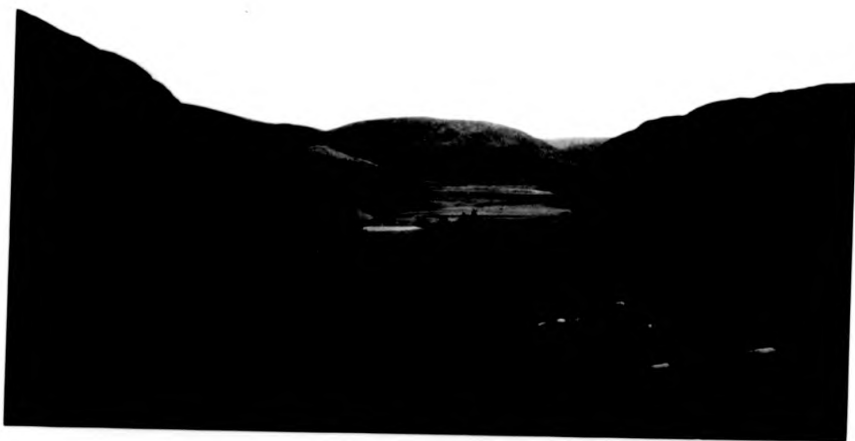
FIGURE 4.231b
The rounded plateau incised by the River Findhorn, uptstream of Tomatin

135
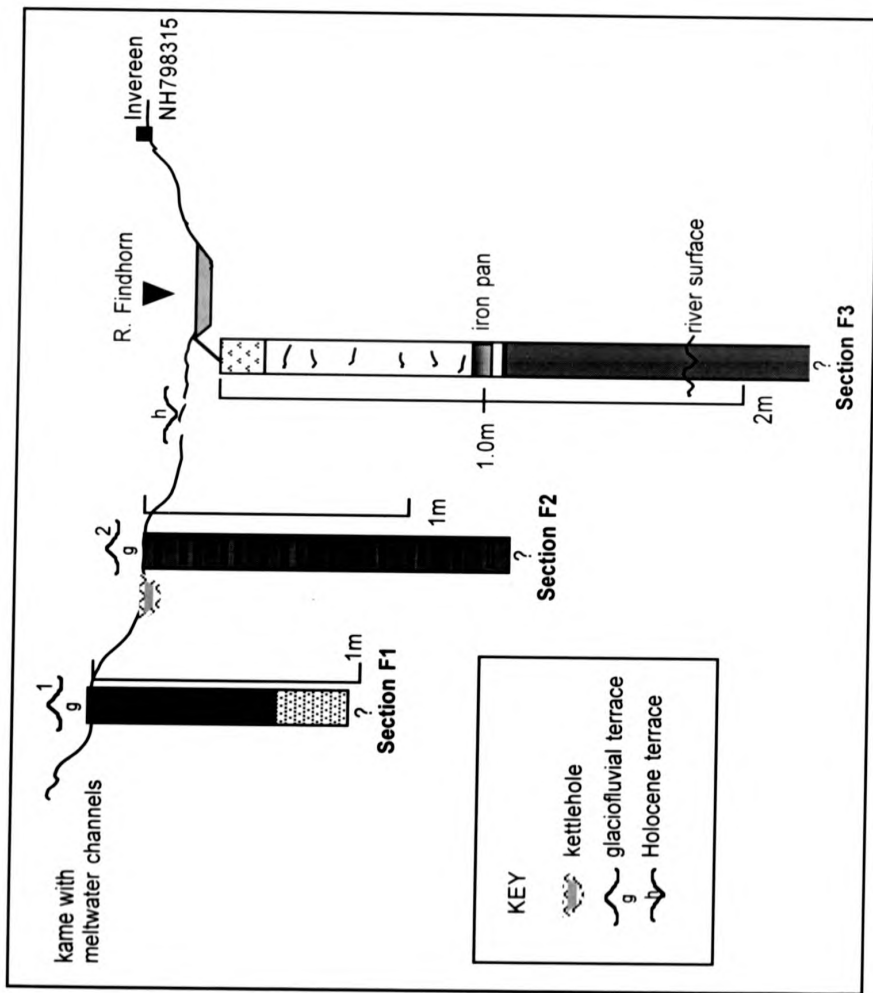
FIGURE 4.32

Schematic cross-section of Findhorn valley near Invereen showing known stratigraphy.

Section F3 is 500m downstream from F1 and F2.
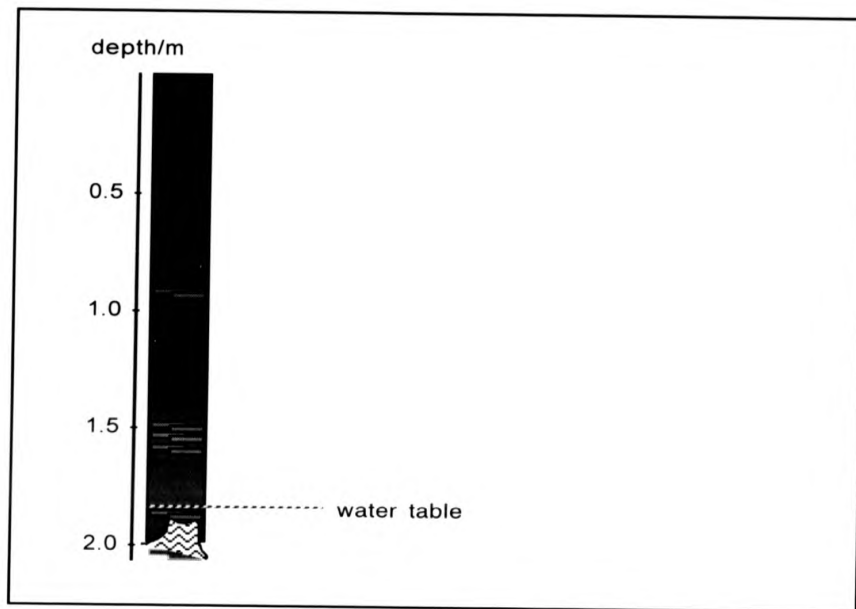
FIGURE 4.33
Stratigraphy of the slope adjacent to Allt na Feithe Sheilich, NH825284

CHAPTER 5: HOLOCENE SLOPE EVOLUTION ON CREAGAN A' CHAORAINN

5.0   INTRODUCTION

The field site is on the north-facing slope of Creagan a' Chaorainn, at the foot of Gleann Chaorainn in central Ross-shire (Figs. 1.1, 5.1). This small valley has an area of 8km^2 and a length of 4.5km. It is glacially excavated in the gneisses and schists of the shatter zone of the Strath Conon fault, one of the series of major NE-SW lineations of Caledonian age which dissect the Scottish Highlands. There is no local weather station, but it lies 15km east of the north-south watershed and is close to the 1600mm isohyet for mean annual rainfall.

5.1   GEOLOGY AND DEGLACIATION OF GLEANN CHORAINN

The valley is excavated in fault-shattered Moine Series pelitic and psammitic schists and gneisses with a sheared slice of Lewisian gneiss in the vicinity of the logged sections (Johnstone & Mykura 1989), and strong geological control of slope characteristics. The gneiss is intruded by amphibolite dykes and pegmatites introducing additional planes of weakness. Foliation dips steeply, generally sub-parallel to the upper valley wall on the south side, and hence into the opposing slope, creating contrasting conditions for sediment transport and drainage. Joints are well exposed on high ground and in near vertical cliffs which form the uppermost valley walls (Fig. 4.2). On the mid and low valley walls, bedrock is largely obscured by sandy, bouldery till. Locally intense deglacial gullying, trending obliquely to joints, dominates slope drainage. (Fig. 4.2).

Deglaciation of the north-west Highlands was probably complete before 13,000 BP (Chapter 1). Gleann Chorainn has been mapped by aerial survey of retreat moraines as lying just within the most easterly margin of the Younger Dryas ice front (Bennett 1994a) (Fig.5.2). Gleann Chaorainn contains abundant evidence for repeated postglacial slope adjustment accompanied by valley floor aggradation. Fans are more common on the north-west valley wall, and are associated principally with long, joint-controlled gullies, and with the Bac an Eich fault zone which was probably active after the retreat of Younger Dryas ice (Fenton 1991) (Fig. 4.2). In the field, smaller fans, which show up poorly in the aerial photographs, can be seen over-running uppermost Holocene river terraces. Others terminate above sections of glaciofluvial terrace and may be either Lateglacial or Holocene in age. Buried organic horizons are also found in riverbank

sections, indicating episodes of valley floor aggradation (Section GChV1, Fig. 5.3). The largest fans over-run probable glaciofluvial terraces and therefore postdate ice melt. Their Holocene age, and repeated aggradation, is confirmed by the presence of buried organic horizons in two dissected fans (Sections GChV2 & GChV3 Fig. 5.3).

5.2   CREAGAN A' CHAORAINN

The northwest-facing slope of Creagan a' Chaorainn which contains the dated sections lies at the foot of the valley close to its junction with Strath Conon (Figs. 5.1, 5.4). Below the slope, the Holocene river, Allt Gleann Chorainn, cuts down through stratified coarse gravels (probable glaciofluvial outwash) to incise bedrock, in what appears to be a re-excavated subglacial gorge. As a result, lower slope sedimentation on Creagan a' Chaorainn, unlike that in the main valley, is decoupled from fluvial processes.

The study site is a slope typical of the local area, and unusual only in that dissection by a modern gully (Figs. 5.4, 5.5a, 5.5b) provides extensive access to sub-surface materials. Slope structure is tripartite (Fig. 5.6) with a steep upper section dominated by a near-vertical cliff. Below this, bedrock dipping steeply at 25-40° has a only a thin patchy cover of bouldery sand. The mid slope has an angle of 15-18° and is dominated by a system of stabilised palaeogullies (Fig. 5.6). The palaeogullies terminate in the convex lower slope, which has a gentler angle of 10-12°, and is underlain by undulating bedrock with occasional protruding knolls. Drift cover consists of 10 cm (min.) to 4m (max.) of sandy till cut by linear zones of alternating peat and mineral debris which have accumulated in the palaeogullies. The thinnest till cover is found on the steep upper slopes where bedrock surfaces are parallel to dipping planes of cleavage. The thickest deposits coincide with the lower ends of the palaeogullies where hollows in the bedrock surface create a sediment sink. The upper slope has a cover of thin, eroding, blanket peat. Lower down, grasses have largely replaced heather due to intense grazing. Thick peat appears mainly at the base of the slope where the till cover thins over rising bedrock. The fact that the lower slope is a sediment sink protected from fluvial erosion and deposition simplifies the depositional history.

The investigation of slope evolution was based on seven sections (GCh4-9) which are exposed in the walls of the modern gully and contain the history of palaeogully fill, or of events which affected intergully zones, together with one (GCh10) further downslope in the wall of a minor incision of a palaeogully (Fig. 5.7). The first three sections were

not dated because of their limited scope, but confirm the presence of buried Holocene surfaces and repeated slope remodelling (Fig. 5.8). Bank collapse sections of uncertain orientation in the gully floor contained peat with fragments of wood with well-preserved bark identifiable as birch and hazel between thin layers of debris. The latter tree species is no longer common (or present?) in upper Strath Conon (pers. obs.), although it is found at altitudes of about 100m near the foot of the valley. Hazel nut (*Corylus*) shells were also found in peat beneath a Holocene terrace in lower Gleann Chaorainn (NH256497).

The alternating mineral/organic couplets found in some of the logged sections were unexpected. Commonly, slope exposures reveal only blanket peat over till. The discovery of these gully-fill deposits, and of extensive thin charcoal layers within reworked till between gullies, suggested the location could provide new information about the timing and frequency of slope stability and instability. In some places the number of mineral/organic couplets further indicated that the preserved slope history might cover a long time period.

5.2.1 Slope hydrology

Groundwater flow on the north-west slope of Creagan a' Chaorainn depends on the interaction of geological and geomorphological controls. The bedrock is foliated, jointed, and in places closely fault shattered, allowing groundwater infiltration through innumerable hydraulically linked fissures. Sustained groundwater seepages through outcrop are maintained even in the driest summer (personal observation). This indicates a minimum groundwater residence time of weeks, and moderately long flow paths. The parallel set of palaeogullies forms a series of linear depressions attracting run-off and lateral flow (Figs. 5.5a, 5.6,). Despite that, they remain today almost entirely infilled and vegetated with only a minor scattering of debris at their upper extremities. Exposures of blue-grey and yellow inorganic plastic clay emplaced on bedrock are occasionally visible beneath a cover of predominantly sandy, cobbly, locally bouldery till. Judging by the absence of clay exposures in the modern gully and variations in surface wetness, these plastic clays locally impede surface drainage on the mid and lower slopes but are not widespread.

Drift cover and modern soils are logged and described in detail below. In summary, the drift is a permeable sandy till of variable thickness overlain by modern soils ranging from sandy peat to thin mineral soils with a low organic content. Since the till in the

main directly overlies shattered bedrock, generally good hydraulic connectivity between the two materials is likely.

5.2.2 Origins of the Modern Gully

Incision of till to bedrock to a depth of 2-4m in the modern gully (Fig. 5.5b) occurred after 1970, since the gully is not marked on the revised 1:10560 Ordnance Survey map of that date, but before 1988, since it is visible in the 1988 Scottish Office air photographic survey. The gully originates from the side of a footpath which is marked on the 1970 map (Fig. 5.4), but has since been widened to accommodate all-terrain vehicles. Field investigation showed that a drain from this track was fortuitously sited so that it greatly enhanced run-off from a small rock face some 2m high, causing a deep, and currently widening incision which cuts down vertically in an excavated joint, then diagonally across the slope. The modern gully terminates, as it starts, in bedrock, in a narrow, pre-existing bedrock gully cut in fault-shattered rock.

The top of the gully is at an altitude of 270m and the base at 195m (Fig. 5.7). Above this, Creagan a' Chaorainn rises to an altitude of 550m. The maximum width is around 10m but changes with seasonal bank collapse. Its length is c. 400m, but vertical sections were logged only in the central portion above the 200m contour, where it traverses the system of palaeogullies at an angle approximating to a cross-section (Fig. 5.5a).

5.2.3 The dated sections

This section briefly introduces the dated sections which are numbered GCh4 to GCh10 (Figs. 5.9 - 5.15) and are 1.5 to 3.0m in height. They are within and between palaeogullies and are exposed on both walls of the modern gully. Samples within each section are identified as, for example, GCh 9/1 and GCh9/2, with sample 1 stratigraphically lower than sample 2. Sections GCh4 - 10 are at successively lower altitudes because the modern gully cuts across the slope in a downward pointing diagonal, at a high oblique angle to the trend of the palaeogullies (Figs. 5.5a, 5.7).

In cross-section, the palaeogullies have comparable dimensions to the modern gully, with intergully zones 8-18m wide. Differential mass movement in the gullies and intergully zones has resulted in laterally impersistent horizons, which could not be lithostratigraphically correlated. However, buried organic materials offered the possibility of chronostratigraphic correlation, that is, an event stratigraphy.

Some exposures contain alternating mineral and organic horizons which represent palaeogully fills. In intergully zones, only reworked till with some thin charcoal horizons and occasional thin soils was found. Given published peat accumulation rates of 2-7cm/100yrs (discussed in detail below), and the thickness of the peats, basal organic horizons seemed likely to be thousands, rather than hundreds, of years old. A variety of organic materials was therefore sampled and dated. Two sections were found to contain a sequence of palaeopodzols. Other organic matter was in the form of peat, wood and charcoal. Sediment analysis provided complementary information about the nature of mass movement.

The scarcity of high resolution information about natural mountain slope evolution over long time periods in the U.K. and in Europe as a whole, indicated that this could be an informative site. The nature of the exposure also raised the question of whether other poorly incised slopes in the northern Highlands with stabilised palaeogullies, conceal equally complex histories (c.f. Hinchcliffe 1999).

5.2.4 Reliability of radiocarbon dates

Twenty-five dates obtained for Creagan a' Chaorainn are shown in Table 5.1 overleaf. The thin sequences encompass a time span of over 7000 calendar years.

A problem affecting dates in this setting is the mobility of carbon through vertically and laterally permeable sandy diamicton during thousands of years in which the water table on the slope may fluctuate on short and long timescales. Modern soils are about 20cm in depth, varying from sandy peat at the base of the slope, through to friable sandy slopewash up slope. This raised the issue of contamination of shallow samples by surface carbon. As a result, uppermost peat surfaces (with one exception) were not dated. The exception was Sample GCh10/2 (Fig. 5.15) where the soil is a sandy peat and thus less permeable than its upslope equivalents. There is therefore a bias in the record obtained against the most recent slope movements, and the dates when critical thresholds for the establishment of modern soils occurred.

Lab Code	Sample	Material	¹⁴C age	cal BP	δ13C/‰
SRR-6051	GCh4/1	charcoal (humin only)	5335±45	6265 - 6030	-28.2
SRR-6052	GCh4/2	peat (humin + humic acids)	5510±45	6395 - 6280	-27.6
SRR-6053	GCh4/3	peat (humin only)	2495±50	2715 - 2360	-25
SRR-6054	GCh4/4	peat (humin only)	2455±50	2710 - 2355	-25.3
SRR-6055	GCh5/1	peat (humin only)	5660±50	6490 - 6360	-26.4
SRR-6056	GCh5/2	peat (humin only)	460±45	525 - 500	-28.5
SRR-6057	GCh5/3	charcoal (humin only)	275±45	420 - 285	-28.4
SRR-6058	GCh6/1	charcoal (humin only)	5290±45	6170 - 5945	-25.6
SRR-6059	GCh7a/1	podsol (humin only)	5450±45	6290 - 6200	-28.1
SRR-6060	GCh7a/2	podsol (humin only)	1820±45	1820 - 1630	-31.4
SRR-6061	GCh7a/3	podsol (humin fraction)	4560±45	5440 - 5050	-28.6
SRR-6062	GCh7a/3	podsol (humin + humic acids)	2435±45	2705 - 2355	-27.9
SRR-6063	GCh7a/4	charcoal (humin only)	6405±45	7380 - 7300	-26.4
SRR-6064	GCh7b/1	peat (humin + humic acids)	5970±45	6890 - 6745	-27.2
SRR-6065	GCh7b/2	charcoal (humin only)	6615±45	7525 - 7395	-26.4
SRR-6066	GCh 8/1	charcoal (humin only)	5795±45	6705 - 6490	-27.3
SRR-6067	GCh 8/2	peat (humin + humic acids)	6165±45	7160 - 6950	-28.5
SRR-6068	GCh8/3	peat (humin + humic acids)	2140±45	2295 - 2010	-28.6
SRR-6069	GCh9/1b	soil (humin + humic acids)	6260±45	7220 - 7030	-27.6
SRR-6070	GCh9/1a	wood (humin only)	6205±45	7200 - 7010	-26.7
SRR-6071	GCh9/2	charcoal (humin only)	5925±55	6855 - 6665	-28.1
SRR-6072	GCh9/3	peat (humin + humic acids)	3850±45	4350 - 4150	-28.5
SRR-6073	GCh9/4	peat (humin + humic acids)	1260±50	1260 - 1085	-27.9
SRR-6074	GCh10/1	peat (humin + humic acids)	3340±45	3635 - 3475	-28.3
SRR-6075	GChS10/2	peat (humin + humic acids)	1410±45	1345 - 1285	-27.7

TABLE 5.1 Radiocarbon and calibrated dates: Creagan a' Chaorainn

In order to assess the timing of debris flows, assumptions about the age relationship between them and the peat surfaces on which they rested had to be made and tested. The number of inter-peat debris flows logged (17) greatly exceeded the number of dated peat surfaces from which their timing could be derived (5). The micromorphology of peat/debris flow boundaries (see below for details) showed that erosion of underlying peat was on, at most, a millimetre scale. Most debris flows are therefore treated as instantaneous events coeval with the underlying peat surface, and are taken to be the events which interrupted peat growth, although a time gap between the formation or deposition of the organic matter dated and the arrival of overlying debris, cannot be excluded. Further uncertainties affecting the precision of radiocarbon dating, but not specific to this slope, were discussed in Chapter 3, Section 3.2.2.

Charcoal dates were used to cross-check dates derived from peat and soil organic matter from the same stratigraphic level. Preservation of drifts of fragile charcoal fragments up to 15mm in length between debris flows on Creagan a' Chaorainn (Fig. 5.16) suggests rapid burial, not transport and repeated abrasion. On a steep slope such as this, fragments of macroscopic charcoal are susceptible to transport by rain, wind and snow melt. The likelihood of them persisting at the surface for lengthy periods on a mobile slope is probably less than in stable lowland environments. So while the problem of persistence is recognised (Chapter 3, Section 3.2.2), in the absence of evidence to the contrary, charcoal is here taken as a reasonable indicator of the age of its confining surface.

Comparison of the peat sample GCh7b/1 - age 6890-6745 cal BP - and the apparently older charcoal enclosed in it - sample GCh7b/2, age 7525-7395 cal BP - in Figure 5.13 illustrates aspects of the problems of establishing a secure event stratigraphy on Creagan a' Chaorainn. A possible explanation for the age discrepancy is that the charcoal persisted for some 600 years on the slope, before small degraded fragments were washed into the peat. An alternative explanation may be that the pre-treatment for the two samples differed (Chapter 3, Section 3.2.2). Different fractions of peat have been shown experimentally to differ in age by more than 2000 years (Shore *et al* 1995). The more reliable date is therefore taken to be that for humin only (the charcoal) since it is less mobile in the environment than humic acids (Chapter 3, Section 3.2.2) The age of the thin, poorly preserved peat may therefore be skewed by the inclusion of younger carbon. These explanations are not mutually exclusive, and interpretation is

consequently subject to uncertainty.

The contrast in age between Samples GCh8/1 (charcoal) and 8/2 (peat) poses a comparable, but opposite problem (Fig. 5.11). In this instance, the humin-only date for the charcoal is *younger* than that of the overlying peat (humin + humic acids). (Table 5.1 above). If the same argument applies, and the charcoal date is more reliable, the peat must in this instance, be contaminated by older, not younger carbon since the charcoal particles are too large, and form too coherent a layer to have percolated through the peat. They lie on the upper surface of partially water-sorted till, presumably washed on to the till surface by run-off, and the three horizons appear to be in the correct stratigraphic order.

In Section GCh5 (Fig. 5.10), charcoal and the peat on whose surface it lies are in the expected sequence. Both are humin-only dates (Table 5.1). This is consistent with experimental findings that, as discussed above, pre-treatment can have a significant effect on the precision of dates.

Dating of palaeopodzols was known to pose particular problems (Chapter 3, Section 3.2.2). Sampling was therefore planned to allow some cross-checking. Sample GCh7a/4 is a charcoal lying between two podzols in a section disrupted by shearing (Fig. 5.13). If preservation of fragile, 10-15 mm long, fragments of charcoal between debris flows in a coarse, sandy till, was the result of rapid burial, the age of the charcoal defines the timing of truncation of the lower podzol and the emplacement of the overlying debris flow within which a further podzol profile subsequently formed. However, a literature search found no information on charcoal preservation in cool, humid slope environments, and the conclusion is therefore untested.

5.3   RESULTS

The following sections describe the nature of mineral sediments on the slope (5.3.1, 5.3.2); the nature of the boundaries between mineral and peat horizons (5.3.3); peat growth in palaeogullies (5.3.4); macroscopic charcoal (5.3.5); and palaeopodzols (5.3.6).

5.3.1   Mineral Sediments

The mineralogy, and the textural and mineralogical maturity of slope sediments, were analysed in order to test field interpretation of two distinct facies. Sandy till covered the slope and formed the substrate of gullies and the material of intergully zones. The

hypothesis to be tested was that the debris intercalated with peat as gully fill originated in till. Raw data (mass per fraction + fractions as percentages) are shown in Tables 5.2 (till) and Table 5.3 (inter-peat debris flows) overleaf. The two sediment types are compared in Table 5.4 and Figures 5.17 a) & b). [*For convenience, these figures are placed following Table 5.4 in the text, not at the end of the Chapter.*] Six features of each fraction and facies were considered (Chapter 3, section 3.2.3).

5.3.1.1 Till

Till samples (Table 5.2) were collected from intergully zones in the upper and mid sections of the walls of the modern gully where no evidence of sediment movement or water sorting was apparent in the field. In grain size distribution, mineralogy and textural maturity, they closely resemble the indurated material (fragipan?) which is exposed at the base of the sequence in sections 10-30 cm thick. Fragipans are considered to result from soil induration at the top of a former permafrost table (FitzPatrick 1956, 1986). On Creagan à Chaorainn, till could have been affected by periglacial processes during the retreat of glacier ice at the end of the Younger Dryas Stade its the margins lay just to the west of Gleann Chaorainn (Section 5.1).

Small variations in the distribution of particle sizes between the indurated till and the unconsolidated tills (Table 5.2 overleaf) are well within sampling error. Mineralogy, weathering and angularity are almost identical. Small patches of cobbly/pebbly diamicton, with finer material winnowed out, can be found in the palaeogullies, but water reworking of till is a small scale localised phenomenon on this slope.

Rock types in the till are representative of the geology of the surrounding area. Particle sizes greater than coarse sand are dominated by rock fragments, a sign of limited transport and weathering. Individual grains of quartz, with fresh feldspar, mafic minerals and small fragments of muscovite begin to appear in the medium to fine sand fraction where resistant quartz grains begin to exceed 50% by volume, reflecting a degree of weathering. All the sediment fractions have angular to very angular grains.

Sample No.	Size/ mm	mass/g	%	cum. %	Sample type
1	30-40	232.72	24.4	24.4	basal,
	20-30	103.80	10.9	35.3	indurated till
	10-20	135.64	14.2	49.5	
	1.5-10	189.06	19.8	69.4	
	1.0-1.5	51.99	5.5	74.8	
	0.5-1.0	25.59	2.7	77.5	
	0.2-0.5	91.94	9.7	87.2	
	<0.2	<u>122.49</u>	<u>12.9</u>	100	
		953.23	100.1		
2	30-40	169.01	33.1	33.1	unconsolidated till
	20-30	53.98	10.6	43.7	
	10-20	33.42	6.6	50.3	
	1.5-10	95.91	18.8	69.1	
	1.0-1.5	29.74	5.8	74.9	
	0.5-1.0	15.24	3.0	77.9	
	0.2-0.5	40.83	8.0	85.9	
	<0.2	<u>72.01</u>	<u>14.1</u>	100	
		510.14 100			
3	30-40	484.68	32.6	32.6	unconsolidated till
	20-30	178.41	12.0	44.6	
	10-20	165.03	10.6	55.2	
	1.5-10	281.00	18.9	74.1	
	1.0-1.5	55.01	3.7	77.8	
	0.5-1.0	32.71	2.2	80.0	
	0.2-0.5	124.89	8.4	88.4	
	<0.2	<u>165.03</u>	<u>11.1</u>	99.5	
		1486.76	99.7		

Mean values	% of mass	cum. %	all samples
30-40	30.0	30.0	angular to very angular in all size ranges;
20-30	11.2	41.2	
10-20	10.6	51.8	mineralogy: mica schist (c. 70%; pink
1.5-10	19.1	70.9	(minor grey) gneiss; pegmatite; quartz
1.0-1.5	5.0	75.9	grains approach 50% only in <0.5mm
0.5-1.0	2.7	78.6	fractions; fresh mica laths in fine
0.2-0.5	8.7	87.3	fractions; no organic matter.
<0.2	12.7	100	

TABLE 5.2 TILL ON CREAGAN A' CHAORAINN

Samples from the upper and mid sections of the modern gully walls

No.	Size/mm	mass/g	%	cum. %	comment (all samples)
1	50-150	404.86	45.8	45.8	All fractions angular/ v. angular
	30-50	135.91	15.3	61.1	rare sub-rounded, weathered
	20-30	51.73	5.9	66.9	clasts of >30mm.
	10-20	56.52	6.4	73.3	Mineralogy of local derivation:
	1.5-10	72.42	8.2	81.5	pegmatite, pink gneiss,
	1.0-1.5	21.30	2.4	83.9	amphibolite, grey gneiss, schist;
	0.5-1.0	10.73	1.2	85.1	individual feldspar and
	0.2-0.5	64.87	7.3	92.4	ferromagnesian grains appear in
	<0.2	65.80	7.4	100	<10mm fraction;
		884.14	99.9		<1mm fraction contains c. 40%
					free quartz + muscovite+ rock particles
2	50-150	1501.04	46.3	46.3	quartz is >50% of <0.5mm
	30-50	672.49	20.8	67.1	fraction plus mica and feldspar
	20-30	411.44	12.7	79.8	with minor rock particles.
	10-20	148.97	4.6	84.4	
	1.5-10	191.53	5.9	90.3	Debris is coarse, with fine tail
	1.0-1.5	36.89	1.1	91.4	and poor sorting.
	0.5-1.0	25.63	1.0	92.4	
	0.2-0.5	141.1	5.3	97.7	
	<0.2	86.67	3.3	101	
		3239.02	101		
3	50-150	2539.57	55.7	55.7	
	30-50	666.82	14.6	70.3	
	20-30	448.33	9.8	80.1	
	10-20	261.05	5.6	85.7	
	1.5-10	250.61	6.5	91.3	
	1.0-1.5	45.60	1.5	92.3	
	0.5-1.0	93.96	2.1	94.4	
	0.2-0.5	156.90	3.4	97.8	
	<0.2	97.09	2.2	100	
			100		

Mean values

Size/mm	%	cum. %	comment
50-150	49.3	49.3	>50mm - 49.3%
30-50	16.9	66.2	
20-30	9.5	75.7	
10-20	5.5	81.2	>10mm - 81.2%
1.5-10	6.3	87.5	
1.0-1.5	1.5	89.0	>1mm - 89.2%
0.5-1.0	1.4	90.4	
0.2-0.5	5.3	95.7	
<0.2	4.3	100	

TABLE 5.3 INTER-PEAT DEBRIS FLOWS ON CREAGAN A CHAORAINN

A number of features confirm the till as little altered since deposition. Chemically and physically unstable particles of muscovite, amphibole and feldspar are well preserved, while even large particles remain unrounded by impacts such as occur with water transport. Rock particles are predominantly derived from the mica schist with minor mafic intrusions which is characteristic of local geology. The high percentage of rock and mineral particles other than quartz in medium to coarse sand, with mica schist forming the majority, reflects rapid breakdown of rock particles. The angularity is characteristic of ice transport.

5.3.1.2 Inter-peat debris

Samples of the debris found between peats were taken from the base, middle and upper horizons of Section GCh9, spanning a time period of some 5000 years (Fig 5.14). They are visually indistinguishable from inter-peat debris in other logged sections. Despite wide age separation the three samples are very similar in mineralogy, textural maturity and range and relative proportions of clast sizes (Table 5.3 above).

The mineralogy reflects very local bedrock with a high proportion of pink gneiss, pegmatite and amphibolite, which predominate over schist. Angular rock fragments form the majority of particles in all grain sizes over 0.5mm (medium sand). Large angular fragments are a particularly good indicator of brief and/or impact-free transport, since the greater the mass and velocity, the more rounding occurs. The angularity, mineralogy and poor sorting of the sediment are therefore consistent with a very local source and brief entrainment. The largest debris clasts match the size and extreme angularity of occasional contemporary surface particles, produced by recent erosion of fault-shattered bedrock.

The debris includes a small amount of weathered material. Rare sub-angular clasts in the 20-30mm size range show surface weathering, and are likely too have the same origin as occasional particles visible today, which have lain on the slope for lengthy periods of time. There are also some randomly distributed small masses of fine rootlets in samples 1 and 3.

5.3.2   Comparison of till and inter-peat debris

Clast size, mineralogy, sorting and angularity or till and inter-peat debris flows are compared below.

Size range/ mm	% in debris flows	% in tills
50-150	49.3	0
40-50	16.9	0
30-40	30	30
20-30	9.5	11.2
10-20	5.5	10.6
1.5-10	6.3	19.1
1.0-1.5	1.5	5
0.5-1.0	1.4	2.7
0.2-0.5	5.3	8.7
<0.2	4.3	12.7

TABLE 5.4

Comparison of till and inter-peat debris flow clast sizes
Creagan a' Chaorainn
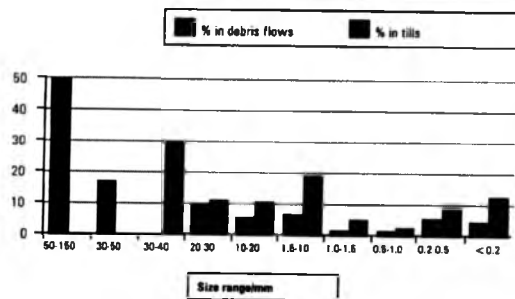
FIGURE 5.17a

Comparison of inter-peat debris and till clast sizes
Creagan a' Chaorainn
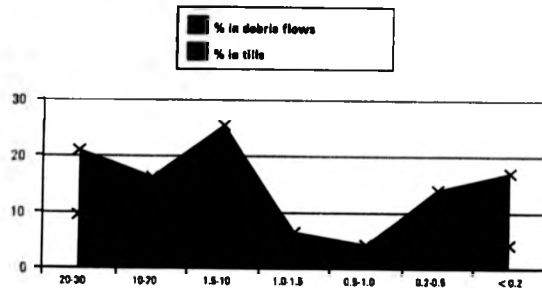
FIGURE 5.17b

Comparison of clast sizes in till and inter-peat debris flows (finer fractions only)

Four features distinguish inter-peat debris from till:

- It contains particles coarser than any in the till.
- The percentage of >10mm mica schist fragments derived from regional rocks is <20% (compared to 50-60% in till) due to the predominance of pink gneiss and pegmatite from outcrop local to Creagan a' Chaorainn.
- The >20mm fraction contains a few sub-angular, weathered clasts.
- There is a small amount of macroscopic organic matter in the form of rootlets.

More than 50% of the mass of the inter-peat debris flows (based on mean cumulative percentages), is in the form of coarser clasts than are found in the till (Fig 5.17a). These clasts are unlikely, therefore, to be derived from the till. Inter-peat debris also has a much lower percentage of fines than till (Fig. 5.17b). The relative depletion of finer material varies from 50% to 70% for every fraction smaller than 20cm in diameter.

Till and inter-peat debris are distinct in respect of clast size distribution, mineralogy and extraneous material. Their genesis is therefore different. If the coarsest, very angular debris is not sourced in till it must come from another very local source. Outcrop on the upper slope together with fault-shattered bedrock on gully floors (Figs. 5.5b, 5.6) is the only obvious source and closely matches inter-peat debris in terms of lithology, angularity and fault shattering.

The small percentage of weathered clasts and rootlets are likely to be surface material entrained in the moving debris. Scattered weathered rock fragments can be seen emerging from beneath thin peat today on the upper slope of Creagan a' Chaorainn. Attack by humus-rich, very acid groundwater in blanket peat (pH 2.5-3.5, Burt *et al* 1990, Fuchsman 1986), causes alteration of rock fragments beneath eroding peat throughout the Scottish Highlands. Hydrolysis of mafic minerals leads to 'bleaching' as a result of production of clay minerals and solutes. The length of time required for mineral alteration in these conditions has not however been reported.

5.3.3 Micromorphology

The micromorphology of peat/debris boundaries was examined to establish the extent of disruption of underlying peat surfaces. Debris flow-peat boundaries appeared generally planar and sharp in the field but if the youngest surface material has been removed, uncertainties in dates would be increased, and the timing of debris deposition made more uncertain. A further sample was taken to validate field observations of an abrupt

change in peat humification. Procedures and limitations are those outlined in Chapter 3, Section 3.2.4. Photomicrographs are shown in Figures 5.18 - 5.20 and the three samples, identified by the symbol * in the corresponding graphic logs, are described in turn below.

Peat boundary in Section in GCh4 (*Fig.6.10= 5.9)

This sample spans the boundary between the lower well-humified peat and the overlying, very poorly humified, layered peat. The lower peat consists of mid-brown AOM (amorphous organic matter) with occasional partially decomposed to well preserved, horizontally oriented, woody fragments (Figs. 5.18a & b). The upper peat lacks both woody fragments and the sub-horizontal linear structures associated with intense compaction (Fig. 5.18c). Black particles are sparsely scattered throughout the upper peat (Fig. 5.18c). From Kemp (1985) they are ovoid (?) faecal pellets and plant fragments. Small blebs of light brown AOM with 'floating' silt are present. The microscopic boundary between the two peats is sharp, marking an abrupt change in plant material (woody material giving way to *Sphagnum* and grasses) visible to the naked eye. This vegetation shift is accompanied by an invasion of fine sand to silt sized particles (Fig. 5.18d).

The poorly humified peat contains small aggregates of light brown AOM with silt which are eroded soil aggregates (Kemp 1985). There are also rare grains of pitted quartz whose surface is etched to form scattered irregular indentations visible under high (x 400) magnification.

Peat/debris flow boundary: GCh5.2 (*Fig. 5.10)

The sample is from the top of an approximately 1m thick peat, itself overlain by a coarse debris flow. The peat is composed of mid to dark brown AOM with some decomposed plant remains, disseminated flecks of (?) charcoal and a small amount of carbonised woody tissue up to several millimetres long showing clear cell structures (Fig. 5.19a). It is almost entirely free of mineral grains. Close to the upper surface are large (up to 10mm) charcoal fragments (Fig. 5.19b) with residual cell structure visible.

Coherent peat gives way upwards to an irregular zone of mixed peat and mineral grains 10-20mm thick (Fig. 5.19c) in which wood fragments from the peat are chaotically mixed with coarse sand containing strained polycrystalline quartz derived from a metamorphic source (Fig. 5.19d). Above that is a mineral layer consisting of unsorted

angular muscovite, feldspar, quartz clasts and rock fragments with rare chlorite and biotite, matching bedrock mineralogy. Micas are fragile and fracture easily under compression. Mica laths visible in thin section have therefore avoided compression. There is no sign of Fe/Mn grain and vug coatings typical of eroded soil (Bertran 1993, Kemp 1985). The poorly sorted silt to coarse sand grade mineral debris is typical of the basal zone of reversely graded debris flows with preservation of fragile mineral grains attributable to impact-free grain flow. The upper peat in Figure 5.19 has been mixed with mineral grains, but only to a maximum depth of 20 mm below a 470 mm thick debris flow with clasts up to 30 cm in diameter. There are no sheared fabrics (Kemp 1985), so peat has not been laterally displaced. The reworking has been both of very limited vertical extent, and muted enough to preserve fragile, macroscopic charcoal.

Since the top few millimetres of the peat surface are homogenised, the age of 525 - 500 cal BP for sample GCh 5/2 is possibly slightly too old. The charcoal age (420 - 285 cal BP, Fig. 5.10) affords a comparison. But allowing for error bars, the measured age difference between peat and overlying charcoal may be as little as 80 years. The precise timing of debris emplacement is uncertain, but the charcoal date and micromorphology suggest a date after about 420 cal BP.

*Peat/debris flow boundary: Section GCh8/3 (*Fig 5.11)*

This thin section, like the previous one, samples the zone between peat and an overlying debris flow. The peat is well humified and reddish-brown in colour, with partially decomposed woody plant residues and scattered ovoids (faecal pellets?). Angular grains of fine sand are very rare and may be wind blown (Fig. 5.20a). Also seen in Figure 5.20a are a number of horizontal elongate vugs. They may have formed when plant material decayed during peat compaction. The vugs die out quickly away from the immediate boundary with the debris flow.

Figure 5.20b shows how these horizontal discontinuities within the peat give way laterally and vertically to disrupted layers with complex folds. Woody fragments in the peat are rotated to a sub-vertical position. These structures are found only in the peat and are absent from the light brown AOM above, which has not therefore been subjected to the same processes. The light brown AOM encloses millimetre-scale irregular masses of remobilised peat.

Above the folded and rotated zone, the light brown AOM, whose colour contrasts with

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the darker peat, is liberally peppered with small angular quart grains. Within this area is a folded, dislocated, strand of plant material (Fig. 5.20c). When seen at 100x magnification, it is clearly segmented, with no variation in diameter along its length. High order birefringence colours and 'Mexican wave' extinction are consistent with a non-crystalline (organic) origin (Kemp 1985). Groups of segmented plants to which it might belong include *Juncus* and *Equisetum*.

The mineralogy and grain size of the mineral fraction in the thin section corresponds to that in Figure 5.19, but the boundary is different. Confirmation of the different origins of the light brown AOM and mid to dark brown AOM is provided by Figure 5.20d. A 3-4 mm deep, wedge-shaped re-entrant of light coloured AOM ranging from silt to coarse sand-sized mineral grains has been driven into the darker brown peat surface. On the left-hand margin of the wedge the lower end of a partially decomposed length of woody tissue is deformed into a V shape. It mirrors the impact, which also formed the re-entrant.

The vertical transition from horizontal structures to chaotically folded material is typical of soft sediment deformation caused by hydrostatic overpressures or minor deformation associated with ground shaking during earth tremors. Peat appears to have behaved in similar ways when subjected to comparable stresses. A reversely graded debris flow with grains in a fluidised state will move with little friction over a soft substrate. Evidence for this has been seen in samples described above. But when the flow is about to come to a halt, dewatering results in a rapid loss of buoyancy, and hence an increase in the shear stress exerted by the flow debris on the substrate. On defluidisation, which must occur as the flow comes to rest and reverts to a rigid state (Iverson 1996), frictional forces on the substrate will briefly be increased. Peat has a high water content but very low vertical transmissivity. This means that sudden compression as a debris flow loses momentum will create high pressures in surface-parallel pore spaces such as elongate vugs. If the water cannot escape from them as fast as it is injected, deformation will ensue - hence chaotically folded vugs and rotation of competent particles. This inference remains to be experimentally verified. The downward force of the terminal debris flow in Figure 5.20d was probably small and/or very brief, since it requires little energy to penetrate the surface of soft peat by a few millimetres, or to bend a sliver of waterlogged, semi-decomposed wood with finger pressure. As in the previous example, deformation has occurred only on a millimetre scale invisible in the field.

The segmented probable plant stem has not originated in the peat, which is well humified and consists of mid to dark brown AOM with woody fragments. Textural relationships, colour, a relatively high content of fine sand, and lack of structures in the light brown AOM, together with its similarity to material in Figure 5.19, strongly suggest that the plant fragment is a delicate rip-up clast derived from a surface soil.

5.3.3.1 Peat-debris boundaries: implications for dating and interpretation

Two effects of debris flows on peat surfaces have been described - mixing and penetration. They can be related to a sudden change in hydrostatic pressures in the substrate as these small flows came to rest. However, the most striking feature is the restricted scale of the disruption. Coarse debris flows up to 500 mm in thickness on Creagan a' Chaorainn have affected peat surfaces to a depth of less than 20 mm. Reverse grading means that during movement, mainly sand and silt size particles have come in contact with the peat substrate. These appear to have acted like a conveyor belt on which large (up to boulder-sized) particles floated. As a result, where peat and soil rip-up clasts occur, they are no more than 2-3 mm in diameter and delicately preserved in a matrix of mineral grains. If the top few millimetres of peat are expanded by an equal volume of sediment, the radiocarbon age will differ little from the bulk age derived from a thin slice of peat, although homogenisation may contaminate the surface with slightly older material.

There is no evidence that peat has been laterally displaced beneath the flows, or that vertical mixing took place on a scale adding substantially to errors inherent in radiocarbon dating. Microscopic investigation has not produced evidence which would invalidate using the age of the underlying peat surfaces to date inter-peat debris flows on the scale seen on Creagan a' Chaorainn. But the possibility of an increase in the error bars on some dates is recognised. Although the peat substrate was remarkably stable beneath debris flows, there is evidence of entrained soil (identified on the basis of light brown AOM). A credible explanation is that thin, friable soil was dislodged by impacts from large angular clasts on the upper slope, close to the source of debris, before reverse grading developed within the flow. This explanation is consistent with the macroscopic evidence for inclusion of rare weathered clasts and rootlets.

The micromorphology of horizon boundaries appears to be a rich and little explored source of information about slope processes, also capable of contributing to understanding of contrasts in peat humification.

5.3.4 Peat growth in palaeogullies

The earliest organic deposits on palaeogully floors consist, in one case, of eroded soil a few centimetres thick (GCh4, Fig 5.9), and elsewhere peat (GCh 5, 7b, 8, 9 and 10). Different types of peat were defined in the field using a x 10 hand lens. Poorly humified peat was mid to light brown in colour and contained distinct layers of sedges and grasses and/or *Sphagnum* mosses. Moderately well humified peat was fibrous and mid to dark brown in colour. Well humified peat was amorphous in texture, and dark brown to black in colour. Twigs and root fragments were found only in moderately and well humified peat.

The timing of initiation of peat growth on formerly unstable gully floors as defined by radiocarbon dates is shown in Table 5.5 below.

Sample No.	^{14}C Age	cal BP
GCh4/2	5335±45	6395-6280
GCh5/1	5660±50	6490-6360
GCh7b/1	5970±45	6890-6745
GCh8/2	6165±45	7160-6950
GCh9/1	6260±45	7220-7030
GCh10/1	3340±45	3635-3475

(Increasing altitude — arrow)

TABLE 5.5
Palaeogully stabilisation
(Basal peat dates on disturbed till on palaeogully floors.)

Three lines of evidence suggest that the gully floors were unstable before these dates: a poorly exposed, indurated, but uncemented, basal layer which directly overlies bedrock but is very thin (?reworked by debris flows); minor water sorting in till, indicating a degree of reworking; and the absence of organic horizons below those recorded in Sections GCh4-10 (?frequent instability). Sample GCh10/1 is much younger than the

others, and is from the lowest, peat-dominated section of the slope below the modern gully. With that exception, basal peat growth on palaeogully floors which consist of till over bedrock, appears to have progressed upslope over a time interval of between 735 and 1000 calendar years. From the history of deposition logged in Figures 5.9 - 5.1, palaeogully history can be divided into two major phases: net erosion between the end of the Younger Dryas Stadial and about 7.5 cal BP, and net accumulation from that time until gullying was recently reinitiated.

5.3.4.1 Peat accumulation rates

Quantitative studies of rates of peat growth in Northern Scotland indicate variations related to temperature, microtopography, plant community and the height of the water table (Charman 1994, Clymo 1984), but deal with patterned fen and raised mire communities respectively. Little is known about accumulation rates for blanket peat on slopes. Rates have been quoted as averaging 2.4 - 4.0cm/100yrs in the northern Highlands on Beinn Dearg, which lies 30 km due north of Creagan a' Chaorainn, and in the Cairngorms (Binney 1997); and 1.4 - 3.4 cm/100yrs in the Cairngorms (Pears 1975). The use of uncalibrated radiocarbon dates in the earlier study, together with estimates based on varying thickness of peat, and secular variations such as altitude, temperature, vegetation type, precipitation and local drainage, conceal the range of uncertainty.

Growth rates for lowland bogs, like those for small basins (Anderson 1995, Tipping 1995), may not be applicable to conditions in palaeogullies on Creagan a' Chaorainn. They are however, broadly comparable: c. 4.0cm/100yrs (Anderson 1995) and 2.9 - 4.9cm/100yrs (Charman 1994). A lower range of 0.2 - 3.3cm/100 yrs was calculated for upland peats in northern England (Shore *et al* 1995). The latter authors pointed out that their rates were calculated on the basis of mean ages for contiguous 1cm slices in peat columns, and that if different fractions were used, different accumulation rates would result. Such high resolution dating is exceptional, and unavailable in this, or any other upland study in northern Scotland.

On Creagan a' Chaorainn, paired upper and lower surface dates which allowed calculation of average growth rates were available for only two peats (Table 5.6 below). Peat accumulation rates are here expressed as maximum - minimum ranges based on calibrated age ranges (2σ) and are at the lower end of published values. On the basis of only two samples, spurious uniformity cannot be discounted. Additional factors

potentially influencing the marginally lower growth rate in GCh5 are the longer time period (>5800 years, as opposed to c. 4000 years), and continuing mass loss with time. Variable growth rates may however be concealed within both these long records.

Section/Fig. No.	Peat description	Thickness	Growth Rate in cm/100yrs	
			Max.	Min.
GCh4/2-4/3	black amorphous peat	62 cm	1.7	1.4
GCh5/1 - 5/2	dark brown amorphous	86 cm	1.5	1.4

TABLE 5.6
Peat accumulation rates: Creagan a' Chaorainn
(calculated from calibrated age ranges)

Because the number of debris flows greatly exceeded the number of dated surfaces, consideration was given to the use of calculated and/or estimated peat accumulation rates from which estimates of debris flow dates could be derived. However, the thickness of the two peat columns (66cm and 83cm respectively) and consequently, the very large time spans measured in both peats, potentially introduces large errors if the averaged rate is applied to thin peats. Local controls on peat growth which vary from vegetation type to topography, substrate and humidity raise questions about validity of comparisons between sites. Estimates of this kind were therefore not made.

5.3.5 Macroscopic Charcoal

Macroscopic charcoal dates on Creagan a' Chaorainn have an uneven distribution (Table 5.7). With one exception in historical times (sample GCh5/3), charcoal is restricted to between approximately 7500 and 6000 cal BP. Charcoal ages are not correlated with depth beneath the surface, emphasising the lateral impersistence of deposits.

Sample No.	^{14}C Age	age/ cal BP	depth/m
GCh7b/2	6615±45	7525-7395	1.6
GCh7a/4	6405±45	7380-7300	1.5
GCh9/2	5925±55	6855-6665	2.6
GCh8/1	5795±45	6705-6490	1.4
GCh4/1	5335±45	6265-6030	1.9
GCh6/ 1	5290±45	6170-5945	1.3
GCh5/3	275±45	420-285	0.6

TABLE 5.7 Macroscopic charcoal ages, Creagan a' Chaorainn

The dates obtained for charcoal sample GCh7a/4 and the oldest humin in the superimposed podzol are widely separated: 7380 - 7300 and 5440 - 5050 cal BP respectively. Possible explanations are discussed below.

At the bases of Sections GCh4 and 9 (Figs. 5.9, 5.14) particles of charcoal are randomly dispersed in thin slopewash, a few centimetres thick, which has the appearance of a disrupted, mineral soil in which minor organic matter has been carbonised. The reworked soils also contain thin, laterally impersistent, but more coherent, charcoal horizons. This charcoal lacks large, woody fragments and is easily reduced to a paste in response to light finger pressure. It may have formed when material low in lignin, such as herbaceous plants, was burned.

Beneath the charcoal in Section GCh9 is a small *in situ* tree stump, fragments of whose scorched surface yielded a date of 7200 - 7100 cal BP (Sample 9/1a). The preserved root spread is consistent with that of a small tree. The radiocarbon date for the tree wood is therefore probably not much older than the death-age of the tree. The thin soil around the stump contains randomly distributed specks of organic matter and lacks any layering. It does not appear to be *in situ*. Since its age (7220-7030 cal BP) is almost identical to that of the tree stump whose outermost layers are missing, and if the radiocarbon age is correct, the soil was developing upslope before the tree was burned down. But the age of the charcoal layer in the soil (GCh9/2) at 6855-6665 cal BP is some 400 years younger than that of the soil and the tree stump. It probably, therefore, captures the timing of the fire and subsequent soil erosion.

Charcoal clarifies the basal stratigraphy of Section GCh9 as follows:

i) The till mantled slope was colonised by herbs and small trees with thin mineral soils.

ii) Fire damaged the vegetation cover, and surface soil together with charcoal was washed downslope in saturated conditions.

iii) Drainage became poorer with charcoal possibly contributing through blockage of soil pores, and peat replaced trees and herbs as the vegetation cover on a stable slope.

The areal distribution of charcoal provides further information about the evolution of the slope. In Section GCh7a (Fig. 5.12) Sample 7a/4 (7380 - 7300 cal BP) is from a

thin layer of charcoal, which was traced laterally for over 3 metres on a dipping surface beneath the upslope inter-gully zone. Charcoal on the opposite side of the modern gully, about 10m downslope, lay at a level consistent with a projection of the dipping surface, but was not dated. It appears therefore that between 7380 and 7300 cal BP a fire produced charcoal which covered at least $30m^2$ of permeable sandy till.

5.3.6 Palaeopodzols

The development of podzols is closely related to hydrology and climate (Avery 1990, Bain *et al* 1993, Bech et al 1997, Caseldine & Matthews 1985, Catt 1979, Ellis & Matthews 1984, Haynes *et al* 1997, Knight 1990, Schaetzl & Isard 1996). Podzols only form when precipitation exceeds evapotranspiration for a significant part of the year, and in a medium which favours leaching. These conditions need to be maintained on timescales of hundreds of years for full profile differentiation. The most favourable conditions for podzolisation are in stable, coarse permeable sediments in a humid climate with little frost, combined with a vegetation of heath and conifer which produces acid litter, since soil acidity enhances the formation of organic chelation complexes and hence mobilisation and translocation of carbon, iron and aluminium. Past conditions of this kind are reflected in repeated humus-iron podzol formation on Creagan a' Chaorainn (Figs. 5.12, 5.13, 5.21).

Very few buried Holocene podzols on upland slopes have been radiocarbon dated and described in Scotland. A study of surface podzols on river terraces in Glen Feshie on the western margin of the Cairngorms 85 km south-east of Gleann Chorainn (mean annual rainfall c. 1050mm; altitude c. 350m) concluded on statistical grounds that soil age was positively correlated with profile depth (Robertson-Rintoul 1986). Some support for inferred podzol age was gained from subsequent radiocarbon dating of two buried podzols in the same area by Bain *et al* (1993) who did not report separation of organic fractions before dating. The dates may therefore be averages of materials of widely separated age. Other authors have suggested that developmental hiatuses and reversals of podzolisation can occur, and that rates of profile development vary with environmental conditions (Caseldine & Matthews 1985, Gerrard 1992, Haynes *et al* 1997, Matthews 1993).

Some of the undated podzols in Sections GCh7a and 7b (Figs. 5.12, 5.13) are truncated to the level of the eluviated, bleached horizon. Podzol horizon thickness in Section G7b varies by up to 100%, but the original thicknesses are unknown, because none of the

organic surface horizons are preserved, and some A horizons are also missing. Profile truncation by one or more debris flows with concave basal profiles can be seen in Section 7a which lies on the upslope side of a partially re-excavated palaeogully. Section 7b is exposed on the opposite side of the same palaeogully where there are no visible signs of till failure, but four stacked incomplete profiles indicate truncation. Beneath the modern soil in section GCh7a, a thin 'ghost' podzol profile (Soil No. 3 in Fig. 5.12) may either be in an early stage of development, or have been disturbed after profile formation.

Three of the A horizons in Section GCh7a were dated. Podzol ages are based on humin (the most stable fraction) from approximately the basal centimetre of A horizons, where the oldest carbon is likely to be concentrated.

Sample No.	^{14}C Age	cal BP
GCh7a/2	1870±45	1820 - 1630
GCh7a/3	4560±45	5440 - 5030
GCh7a/1	5450±45	6290 - 6200

TABLE 5.8
Dates for the initiation of podzolisation

Podzol dates in this environment are subject to unknown error, and the stratigraphy may conceal some erosion surfaces. Dating palaeopodzols is complicated by the fact that different fractions of organic matter may have markedly different mobility in natural environments. Dating trials have shown very large discrepancies of 2000 years or more between humin and humic fractions of the same soil samples (Bol *et al* 1996, Shore *et al* 1995). An extreme example of this problem occurs in the present study where separate analyses of different fractions of sample GCh7a/3 (Table 5.1) gave dates separated by 3085 - 2345 calendar years (at 95% confidence levels). Since humin is much less mobile than humic acids (Shore *et al* 1995), the older age of 5440-5050 cal BP, representing humin alone, is considered more representative of the time when the podzol started to form. This result indicates the high degree of mobility of humic acids on the slope.

Podzols 1 and 2 (Fig. 5.12) appear in the field to be sequential. However, about 4500 calendar years separates the ages obtained for humin in the basal centimetre of their A

horizons (Table 5.1). If the humin dates are correct, either the soils were extremely long-lived, or the stratigraphy is incomplete. The downslope section of Figure 5.12 confirms that two podzols of unknown age developed during this interval. They are not *in situ*, and appear to have slipped downslope, so the latter hypothesis is preferred.

Rapid podzol formation has been variously defined as taking about 350 years (Charman 1992), characteristically, 1000 years (Fitzpatrick 1986), or 'many thousands of years in an Arctic-Alpine environment' where processes are much slower (Caseldine & Matthews 1985). On Creagan a' Chaorainn, below 300m, conditions are damp, soils are acid, and the sediment is permeable. Podzols may therefore have formed relatively rapidly during intervals when humid, non-erosive conditions pertained. This puts a probable minimum limit on the return frequency of the debris flows which truncated their profiles at about 350 years. However, podzolisation on a 15-18° slope could be less rapid than on a level surface, due to an increased rate of leaching associated with downslope drainage. The minimum time for profile development, and the intervals between debris flows, may therefore be longer. With truncation as deep as the base of A horizons, and in the absence of evidence to suggest repetition of horizons through shearing, the stacked sequence of podzols may therefore indicate surface soil erosion at intervals of hundreds of years during a lengthy period of general stability.

Analysis of iron and aluminium concentrations in the reddened, illuviated horizons of four palaeopodzols provides some additional information about their environmental history and significance. Results are presented as the mean of three assays for each sample in Table 5.9 (p. 164). Some data from contemporary upland soils on slightly gentler slopes (Avery 1990) are included in Table 5.9 for comparison. However, the sparsity of data in settings comparable to that of the buried soils on Creagan a' Chaorainn, means that an absolute standard for 'high' or 'low' concentrations is not available, and the small number of samples allows only tentative analysis. Preliminary conclusions are:

i) Both Al and Fe concentrations increased from about 6.2 to 5.2 cal ka BP, with a further increase in the late Holocene. This may indicate an intensification of leaching.

ii) If the undated podzol (GCh7b/X) follows the same trend, it is late Holocene in age.

iii) There does not appear to be any systematic relationship between depth below the

surface and Fe and Al concentrations. This is expected, given evidence for displacement of horizons through till failure.

iv) Fe (but not Al) is greatly enriched in the palaeosols compared to the modern soils.

5.4 DISCUSSION

This section integrates results under the related themes of changes in slope hydrology and stability through time.

Occasional water sorting of till within palaeogullies and downslope orientation of clast long axes (Fig. 5.21) provide evidence of till remobilisation and transport between the formation of a basal indurated horizon, possibly under periglacial conditions, and about 7.5 cal ka BP, when the earliest preserved organic horizons are found. No earlier intra-till boundaries or processes were detected. But it is unlikely that the slope was not organically productive before 7.5 cal ka BP, since there is abundant evidence for plant growth elsewhere in northern Scotland before 7.5 cal ka BP (Chapter 1). The absence of organic material older than 7.5 cal ka BP is unlikely to be due to lack of exposure, since the modern gully samples a number of palaeogullies, and the undulating bedrock surface on the mid and lower slope nowhere suggests greater thicknesses of till than those investigated. As far as can be ascertained, therefore, the lack of OM older than 7.5 cal ka BP is real. This date therefore marks the first known change in system thresholds.

Systems do not necessarily require large inputs to reach threshold conditions. Small changes in values such as the frequency of erosional run-off, the average height of the water table, and the balance between precipitation and evaporation, may be sufficient. Just such a small adjustment can be envisaged between a system state, where organic matter was lost through till failure and rapid oxidation, and a state where the balance tipped, on average, towards long term preservation of summer vegetation as peaty layers on gully floors. Once established, peat itself modifies slope hydrology by further raising the water table and promoting lateral, rather than vertical drainage, thus creating a positive feedback loop in favour of lateral and vertical expansion. In this way, modest changes in forcing factors may account for abrupt, self-perpetuating changes of state in the slope system.

Dated A horizon	^{14}C age/(calkaBP)	Thickness/m (Bh horizon)	Depth of sample /m	Al mg/l	%Al	Fe mg/l	%Fe
GChS7a/2	1820±45/(1725)	0.10-0.12	0.60-0.58	49.28	0.49	171.19	1.71
GChS7a/3	4560±75/(5245)	0.25-0.30	0.95-0.92	47.08	0.45	143.71	1.36
G ChS7a/1	5450±45/(6245)	0.20-0.25	0.80-0.78	40.07	0.41	86.8	0.88
GChS7b/X	not dated	0.10	1.0-0.98	69.25	0.68	185.01	1.83

Author	Location	Material	Age		%Al		%Fe
Avery 1990	Cairngorms	surface soil (slope 12°)	not reported		0.60		0.20
Avery 1990	Cumbria	surface soil (slope 11°)	not reported		0.20		0.30

TABLE 5.9

Fe/Al weathering in Bh horizons of palaeopodzols on Creagan a' Chaorainn on a slope dipping at 15-18°.

Notes

See Figures 5.12 & 5.13 for the stratigraphic position of each palaeopodzol.

Samples were obtained from the basal 1-2 cm of each reddened, illuviated B horizon.

Some data for surface soils on slopes elsewhere in Britain are given for comparison

The oldest radiocarbon date obtained is for charcoal (sample 7b/2), at 7525-7395 cal BP (Fig. 5.1). Four other samples in Sections GCh7a, 8 and 9, all on the lower part of the mid slope, where sediment thickness is at its maximum, contain material formed before 7ka cal BP. In addition to peat, these early organic materials testify to the existence of thin mineral soil, tree growth, and fire which may have occurred in two separate episodes, although the 2σ calibrated age ranges are separated by only 15 years.

The second half of the 7th millennium BP therefore saw a mosaic of slope environments with tree growth, immature organic soil formation, and minor peat accumulation on a well-drained slope, stable on a minimum timescale of decades, in the absence of severe erosive run-off. If the switch from non-preservation to preservation of OM is not an artefact of sampling, a major system change from net erosion and rapid oxidation to overall stability, accompanied by organic accumulation, was achieved though a subtle change in humidity over a period of 0.7 to 1.0 ka. The low mineral content of the peat accumulating on palaeogully floors indicates an almost complete lack of water- or gravity-transported sediment.

By about 7160-6950 (sample 8/2, Fig. 5.11) and 6855 - 6665 cal BP (sample 9/2, Fig. 5.14) persistent peat growth began to replace the earlier more varied mosaic of slope environments.

5.4.1   Changes in slope hydrology

The evidence that slope hydrology underwent long-lasting change in parallel with preservation of buried organic horizons, replacement of trees and thin mineral soils by peat, establishment of peat in the formerly erosional environment of palaeogully floors, and changes in peat humification, is reviewed below.

The preservation of wood fragments in some peats (Sections 4, 5, 10), and well-preserved birch and hazel bark in undated bank collapse sections, indicates at least periodic woodland cover over long stretches of mid and late Holocene time. The presence and absence of wood in peat can be explained in terms of tree and shrub growth, and/or wood preservation. Peat humification is inhibited by a rising water table, which elevates the surface of the anaerobic catotelm within the mat of decaying vegetation, thus reducing the rate of bacterial degradation (Clymo 1984, Fuchsman 1986).

A relatively low water table with extensive oxidative degradation of plant litter in the acrotelm will inhibit preservation of recognisable macroscopic wood fragments in peat. Conversely, a relatively high water table increases the thickness of the anaerobic zone, reducing the time available for decomposition in the acrotelm of the most resistant, tannin- and resin-rich material, that is, wood, which can survive to be incorporated in the catotelm.

But if conditions were too wet, the growth of woody plants would be inhibited. At 2.5 cal ka BP an abrupt change from well humified to very poorly humified peat is recorded in Section GCh4 (Fig. 5.9). Root and twig fragments are incorporated in the upper few centimetres of the lower, well-humified peat, but entirely absent in the overlying peat which is fibrous, layered and so little decomposed as to contain easily identifiable filaments of *Sphagnum* and grass blades. It appears, therefore, that trees died out rapidly as slope conditions became too wet at the humification boundary.

Although peat humification can be a very local phenomenon, on a slope of 15 - 18° in well drained, sandy till, and with rapid lateral drainage from intergully surfaces to gully bases, the peat, though protected from active erosion, is unlikely to have been a hydrologically isolated sump. However, a detailed reconstruction of the relationship of changing woodland cover to changing slope hydrology awaits pollen analysis. The abrupt change in water table in Section GCh4 at 2.5 cal ka BP is accompanied by the sudden appearance of millimetre-scale eroded soil aggregates and etched quartz particles, which must have been sourced upslope in newly exposed, eroding till. Mineral sediment is extremely rare in the underlying peat, indicating a stable, vegetated slope before 2.5 cal ka BP, with little or no exposed till.

Palaeopodzols provide additional information about slope hydrology and stability, although radiocarbon dates from humus in A horizons are subject to greater uncertainty than those for peat or charcoal. From available evidence (Table 5.8), the earliest podzol formed shortly before 6.2 cal ka BP. Two others were dated, but till failure, with the accompanying possibility of repetition of truncated profiles, made the number of separate soils preserved uncertain. The earliest date for podzol development shortly before 6.2 cal ka BP (Sample 7a/1) coincides with the end of the time (7.2 - 6.2 cal ka BP) when gully floor colonisation by peat had replaced erosive run-off (Table 5.5). If the sequence of dates represents a real effect, development of the earliest podzol occurred in a system where low energy seepage and leaching was prevalent, and where

till failure recurrence frequencies were less than the minimum 350 years needed for profile differentiation (Section 5.3.6). Podzol 7a/1 also exhibits relatively low intensity of leaching of Fe and Al complexes (Table 5.9). The start of podzol development is therefore in tune with a transition from an erosive, high intensity flow regime, to stable, gently leaching conditions.

If the three podzol analyses available reflect a real trend, the relatively high values of oxalate extractable amorphous and organic bound Fe and Al in the younger soils show an increase in rates and intensity of podzolisation and leaching between about 6.2 and 5.2 cal ka BP, with a further increase in the late Holocene. The age of the youngest, most intense leaching at about 1.7 cal ka BP, falls within the late Holocene period of enhanced slope instability described in Section 5.4.2 below.

In summary, slope hydrological system change is indicated at about the times listed in Table 5.10 below.

7.5 cal ka BP	Earliest preservation of OM (charcoal, trees, thin mineral soils); till failure and water sorting on palaeogully floors begin to give way to peat.
6.2 cal ka BP	Palaeogullies stabilised; podzol profile differentiation.
6.2 - 5.2 cal ka BP	Intensification of leaching
2.5 cal ka BP	Rapid rise in water table with silt, sand and soil entrained in run-off; disappearance of wood from peat
1.7 cal ka BP	More intense leaching than previously recorded in podzols.

TABLE 5.10
The timing of changes in slope hydrology on Creagan a' Chaorainn

5.4.2 Changes in slope stability

Dated evidence for slope instability gives a misleading impression of the real frequency of events, because logged sections contain the following additional, undated horizons:

i) Mid to late Holocene: nine inter-peat debris flows (Sections GCh 8, 9,10)

ii) Late Holocene: up to four episodes of slope wash which interrupted the evolution of thin peaty soil (Section 6); seven accumulations of silty sand in which modern soils have formed (Sections 4 - 9).

Slides and debris flows in till also truncated podzol profiles, but their timing is less certain (Section 5.4.2.1). Some may have occurred in the early mid Holocene before about 5.9 cal ka BP; others are late Holocene, post-dating the podzol dated at about 1.7 cal ka BP (Sample 7a/2) and the 'ghost' podzol lying between it and the modern soil.

Four different types of sediment were identified on Creagan a' Chaorainn: till, inter-peat debris, thin mineral soil, and silt or sand particles in peat. Information about the nature and timing of their redistribution is brought together in the following sections.

5.4.2.1 Till failure

The extensive charcoal sheet dated at about 7.3 cal ka BP in Section 7b marks a buried surface. Downslope-pointing clasts in the till between it and the modern surface (Fig. 5.21) confirm mid-slope aggradation through till transport as water-saturated debris. Known dates for till failure are listed in Table 5.11 below.

Sample No.	Age range 2σ cal BP	Approx. date of debris flow
6/1	6170 - 5945	6.1 cal ka BP
7a/3	5440 - 5050	5.2 cal ka BP
8/3	2295 - 2010	2.2 cal ka BP
7a/2	1820 - 1630	1.7 cal ka BP

TABLE 5.11
Dated till failures on Creagan a' Chaorainn

Podzols developed in till were also stratigraphically disrupted by circular and/or shallow slab failures (Sections GCh7a and 7b). The features of their failure on Creagan a' Chaorainn do not match the theoretical conditions proposed by (Brooks & Richards 1994) in terms of the modelled plane of failure within the soil profile or the slope steepness. Disruption of podzols probably continued until historical times, since truncated podzols immediately underlie modern soils (Sections 7a and 7b).

Only maximum ages for the timing of the slides which truncated podzols can be derived from the age of underlying organic matter because there is no information on how long the soils remained *in situ* after profile development. Given a possible minimum development time of a few hundred years, and adding this to the age of the oldest sampled humin in order to allow time for soil profile development, known slides of shallow slabs of till which maintained the integrity of partial podzol profiles occurred at some point after 7.1 cal ka BP (Sample 7b/2); 7.0 cal ka BP (Sample 7a/4); and 5.9 cal ka BP (Sample 7a/1). However, the large gap between the age of Sample 7a/4 (7.0 cal ka BP) and the age of the humin in the immediately overlying podzol (5.2 cal ka BP, Sample 7a/3) fails to confirm that these dates necessarily closely reflect the timing of failure of podzolised till. They may have been stable for long periods of time. Dates for podzol failure are added to Table 5.11 above, with the proviso that they may be affected by large errors, with failure in fact taking place much later.

Sample No.	Age range/2σcal BP	Approx. date of debris flow	Podzol failure after
7b/2	(7525-7395) - 350yrs		7.1 cal ka BP
7a/4	(7380-7300) - 350yrs		7.0 cal ka BP
6/1	6170 - 5945	6.1 cal ka BP	
7a/1	(6290-6200) - 350yrs	5.9 cal ka BP]	
7a/3	5440 - 5050	5.2 cal ka BP	
8/3	2295 - 2010	2.2 cal ka BP	
7a/2	1820- 1630	1.7 cal ka BP	

TABLE 5.12
Dated till and podzol failures

The requirement for damp, leaching, stable conditions for podzolisation establishes broad palaeoenvironmental parameters. If stability thresholds at shallow depths on the slope are exceeded, podzols will either not form due to frequent disruption, or if established, will not survive for long. Neither will they form if conditions are so dry as to inhibit at least repetitive seasonal translocation of carbon, and metal complexes.

Where truncated podzol sequences are stacked on top of each other, (provided they are not repetitions of a single, sheared soil profile), this indicates sustained slope surface stability with widely separated erosional events. Critical thresholds separating steady leaching and till failure are recorded in the shearing evident in Sections GCh7a and 7b (Figs 5.12, 5.13, 5.16).

5.4.2.2 Inter-peat debris flows

Upper slope outcrop was inferred as the source of inter-peat debris in Section 5.3.1.2. Processes which provide a plausible mechanism for generating and mobilising this debris have previously been described in the northwest Highlands. Hinchcliffe *et al* (1998) concluded that 27-30% of rockwall retreat above a talus slope in NE Skye since deglaciation was due to granular weathering which produced fines, leading to a build up of porewater pressures in talus during rain storms. Salt & Ballantyne (1997) reported a similar combination of rockfall clasts and fines due to Holocene granular weathering of outcrop in a talus accumulation in the Assynt area, accompanied by periodic redistribution of talus by debris flow. The principal agent of chemical weathering in Scotland is hydrolysis (Bain *et al* 1993) which takes place in the presence of groundwater circulating through bedrock as well as on the surface. Together, these observations suggest the steep rock outcrop in the upper section of Creagan a' Chaorainn and rock gullies as the likely source of much of the inter-peat debris.

The peat, which is virtually free of mineral grains, must have formed when the slope was persistently damp, but at the same time stable, and free from erosive events. From charcoal and tree root remains in the peat, the slope appears to have been vegetated with woody plants before the debris flows occurred. However, there is no clear evidence that trees actually grew on the peat however - that is, that the peat surface had dried out. At about 500 cal BP, a debris flow ended peat growth at this site. The sand and silt at the base of the debris flow contains fragile, elongate, mineral grains, which would have been broken down and rounded during repeated downwashing of sediment, or moderate energy mixing, but would have retained their freshness and angularity during grain flow in which impacts are minimised. They probably, therefore, formed during granular disintegration of the rockface and were carried in the debris flow below large angular clasts. Mixing with the uppermost peat may only have occurred briefly when the flow lost momentum and grain flow gave way to frictional stress.

After about 6.2 cal ka BP, transported sediment in the palaeogullies consisted mainly of rock debris. Sections 5, 8 and 9 provide information about its timing and volume, although the majority of events remain undated.

Sample No.	Age range /cal ka BP	Max. age of debris flow
9/3	4350 - 4150	4.3 cal ka BP
8/3	2295 - 2010	2.3 cal ka BP
9/4	1260 - 1085	1.3 cal ka BP
5/2	525 - 500	0.5 cal ka BP

TABLE 5.13
Dated inter-peat debris flows

The thickness of the three debris flows recorded in the palaeogully sampled by Section GCh8 increases upwards (Fig. 5.11). The flow which occurred after 2.3 cal ka BP is more than twice as thick as the mid Holocene flow which truncated the earliest phase of peat growth, initiated at about 6.6 cal ka BP. This trend is confirmed in Section GCh9 (Fig. 5.14) which samples the main sediment sink on the slope and records the largest number of peat/debris couplets (nine). Between 6.8 and 4.3 cal ka BP the slow accumulation of peat was interrupted by only small volumes of debris on three occasions in the space of 2.5 ka, and mineral debris forms less than 20% by volume of the stratigraphic column - a ratio of 1:5. Between 4.3 and 1.3 cal ka BP incursions of debris occurred on 6 occasions and the debris/peat thickness ratio is reversed to nearly 3:1 - 4:1 above a depth of 1.1m. In other words, during the late Holocene, the relative volume of debris to peat is up to twenty times greater than before 4250 cal BP, with rock debris transported down the gully both more frequently, and in larger volumes.

The lateral and vertical impersistence of events makes calculation of flow volumes difficult. However, an approximation can be made by summing debris flow thicknesses in all sections. A simplified calculation based on the scale of slope movement seen in the field, is that rock debris flows contained in gullies are 3m wide and 20m long, with a volume of about 150m³. A deduction of, say, 10% can be made for entrained material leaving 135m³ derived from a source area of about 5000m². From aerial photographs, an estimated 25% of the surface area of the whole 2 km² catchment has been capable of generating rock debris. So some 54,000m³ of bedrock may have been mobilised in this

small catchment over the last 7.3 ka, but very largely in the late Holocene.

5.4.2.3 Soil erosion

Soil erosion occurred both in association with, and independently of, macroscopic charcoal. In Sections GCh4 and 9, palaeosols were washed downslope, forming centimetre-scale layers. Incorporated charcoal has an age of

- Sample GCh4/1   about 6.2 cal ka BP
- Sample GCh9/2   about 6.8 cal ka BP

The remaining evidence for soil erosion is undated. In Section GCh8 (Fig. 5.11) more than one layer of water-eroded soil and till may be recorded although repetitions of sheared fragments of the same soil cannot be excluded.

In Section GCh9, the date of the soil organic matter itself is 7.1 cal ka BP (Sample 9/1). This may be younger than the real date of the earliest SOM, since carbonised fragments of vegetation contemporaneous with the fire are likely to be incorporated in it. The age of the charcoal is therefore taken to be more representative of the date of soil erosion. The two available dates for mineral soil erosion coincide with the period when palaeogullies were gradually ceasing to transmit sediment and run-off. The fire which burned the tree and surrounding vegetation was closely followed by small scale soil erosion but stable conditions quickly resumed: immediately overlying the soil and charcoal are several decimetres of mineral-free peat which could only have formed in the absence of mobile slope sediment. The peat does, however, signal local environmental change. Before about 6.7 cal ka BP, the slope surface was wooded with thin mineral soil. After the fire, there followed an extended period of peat growth, interrupted by only two very thin layers of rock fragments, until about 4.3 cal ka BP (Sample GCh9/3). The rise in water table, leading to a change of style in the accumulation of biomass (peat instead of trees and thin, herb-covered soils), can be explained by an increase in effective precipitation, and/or the sudden removal of the pumping effect of water into the atmosphere by trees, enhanced by reduced soil permeability as a result of charcoal blocking pores.

5.4.2.4 Sediment in peat

The sand content of peat on gully floors is imperceptible except in microscope slides (Section 5.3.3 above). Large changes in relative proportions of organics and minerals

must occur during peat formation and compaction because, on average, less than 15% of vegetable matter contributing to peat survives leaching and bacterial degradation (Fuchsman 1986). On this basis, the mineral content of the surface vegetation mat would be enhanced by a minimum of 600% during the process of its transformation to a well-humified peat. A sparse scatter of individual mineral grains in peat is therefore likely to be the product of minor wind transport, not of inwashing from eroding slope surfaces during snow-melt or rain storms. Gully peats with very low mineral content must therefore have formed when run-off was virtually sediment-free. This implies extreme slope stability between flows, and a lack of a common trigger for the mobilisation of till and outcrop-derived debris.

Visible silt and sand is restricted to late Holocene peats: the very poorly humified peat in the middle of Section GCh5 which began to form at about 2.5 cal ka BP, and the sandy peat which forms the uppermost horizon of Section GCh10. Sand and silt began to reach the lower slope at about 1.3 cal ka BP (Sample GCh10/2). Sand in the unhumified peat contains pitted (etched) quartz grains (Section 5.4.1). Quartz is resistant to solution at normal earth surface pressures and temperatures. It can suffer marked corrosion in some soils (Pettijohn 1975) although this requires high pH (Deer *et al* 1966). In an area of acid rock and blanket peat, such conditions are more likely to reflect subglacial weathering. The release of chemically etched glacigenic grains to a peaty sump, is consistent with the onset of mineral inwash coinciding with exposure of till, beneath a degrading vegetation cover. The combination of a sharp rise in water table, with erosion of fine sediment from beneath formerly undisturbed soil cover, over a very short period of time, indicates a sudden switch to a wetter, less stable slope environment.

Other evidence for inwashing of sand-grade sediment onto peat surfaces is from contemporary soils. These are peaty in Section GCh10, (Fig. 5.15) in the footslope sump area, and increasingly sandy upslope. They consist of 10-20 cm thick layers of sandy slopewash of a type not seen in the stratigraphy of the last 7.5 ka. Degradation of peat cover to expose till is the most likely reason for recent fine sediment availability. Currently, on the upper slope beneath the rock outcrop, sandy till is in places exposed beneath degrading peat, and small downslope washes of sediment can be seen.

5.4.3 Charcoal distribution through time

Charcoal distribution on Creagan a' Chaorainn is highly skewed, with a complete absence of samples dated from about 6.1 - 0.4 cal ka BP (Table 5.7). Either the distribution is an artifact of sampling or it is a real feature of slope history. Sampling bias cannot be excluded due to both limited exposure and the fact that microscopic charcoal was not investigated. But if logged samples reflect fire frequency, their absence after about 6.1 cal ka BP is difficult to explain.

Macroscopic charcoal appears in the early mid-Holocene when the slope was stabilising in the presence of non-erosive run-off, and while it was covered by a mosaic of trees, herbs and fresh mineral soils, some podzolised, with minor peat. Speculatively, as peat cover increased, charcoal disappeared. Although wood fragments appear in some peat horizons, the slope water table was generally higher (Section 5.4.1 above), limiting tree growth and the production of woody, macroscopic charcoal until tree planting in the last few centuries.

Between 7380 and 7300 cal BP (GCh7a/4) a fire produced charcoal which covered at least $30m^2$ of permeable sandy till. It seems unlikely that the charcoal is *in situ*, since the till above and below it is devoid of other organic matter, and the palaeosurface appears to have been subject to debris flows. This charcoal is therefore interpreted as washed or blown downslope. Before about 7.3 cal ka BP, intergully areas may have supported only a thin cover of low herbs on periodically mobile till.

5.4.4 Environmental triggers

The pattern of slope evolution has five elements, albeit with generally poorly defined boundaries subject to the errors already noted.

1. Before about 7.5 cal ka BP: no organic preservation.

2. 7.5 - 6.2 cal ka BP: stable slope with some trees and thin mineral soils subject to minor disturbance after fire; gradual palaeogully stabilisation with runoff becoming sediment-free at the same time as peat established itself in former erosional sites.

3. 6.2 - 2.5 cal ka BP: spread of blanket peat with very rare (dated) till and bedrock failures; intensification of leaching in podzols and some increase in the frequency and volume of debris flows in the second half of this period.

4. Marked increase in debris flow frequency after 4.3 cal ka BP

5. 2.5 - 1.3 cal ka BP: water table rise and soil erosion; intensified leaching of soils; larger, more frequent flows containing weathered rock debris.

6. Within the last 1.3 cal ka BP: blanket peat cover breached, permitting erosion of fines from underlying till.

Early postglacial relaxation is likely to have established the tripartite slope structure with drainage dominated by a gully system (Chapter 1, Section 1.5). It may have continued to influence slope evolution through further adjustment and lowering of slope gradients, although only the indirect evidence of lack of preservation of organic horizons supports this hypothesis. Continuing postglacial adjustment cannot, however, account for the very low incidence of slope failures during the mid Holocene, and the subsequent marked increase in volume and frequency of rock debris flows, which interrupted peat accumulation in the late Holocene. Potential environmental controls were discussed in Chapter 2. They are: vegetation change, itself triggered by climate and/or human impact; random intense meteorological events interacting with intrinsic factors; and progressive, intrinsic change which can move systems to critical thresholds.

Charcoal provides a record of vegetation disturbance, and burning probably made thin mineral soils vulnerable to erosion by disturbing the stabilising root mat, partially filling soil pores, and hence reducing permeability. Human occupation in upper Strath Conon near Gleann Chorainn is recorded in a single Bronze age artifact found in slope peats, and an Iron Age crannog (Highland Region Archaeological Records), both of which postdate the charcoals (Table 5.7) by several thousand years. The absence of charcoal in the mid to late Holocene is difficult to explain in the light of human occupation.

The alternative source of charcoal is natural fire. Pine woodland with its high resin content is much more susceptible to lightning fires than other tree species native to the Highlands and is possibly the only woodland type in the British Isles that might have burned naturally (K.D. Bennett 1998, pers. comm.). Unlike other native species, *Pinus sylvestris* is well adapted to fire (McVean 1963). Although the pollen stratigraphy of the Strath Conon area is uninvestigated, it lies only about 30 km south-east of classic areas for the study of pine woodland expansion and decline of Loch Maree and Torridon

(Anderson 1995, 1998, Bennett 1995, Birks 1972, 1996, Kerslake 1982, Pennington *et al* 1972). There, the pollen stratigraphy (in radiocarbon years) suggests that a pine decline started slowly at c. 7 ka BP, accelerating at about 4 ka BP (Birks H. J. B. 1996). The latter date matches macrofossil evidence (Bridge *et al* 1990, Gear & Huntley 1991), and pollen evidence, (Anderson 1998, Bennett 1995, Birks H. H. 1972, Kerslake 1982), for pine decline. A pine decline at 5.3 cal ka BP on Beinn Dearg 30 km north of Creagan a' Chaorainn, was described by Binney (1997). The latest mid Holocene charcoal on Creagan a' Chaorainn is also dated at 5.3 cal ka BP (Table 5.7). Speculatively, macroscopic woody charcoal on Creagan a' Chaorainn could be linked to the presence of open pine woodland which declined as peat cover spread after about 7.0 cal ka BP.

Some additional information about woody vegetation on the slope is available (Section 5.2), although the pollen stratigraphy has not been investigated. Preserved bark of *Betula* and *Sorbus* together with *Calluna* stems, were found in buried peats in bank collapse sections in the mid and upper sections of gully fill so could not be dated. In a meander of Allt Gleann Chorainn, whole *Corylus* nuts were found in peat below a late Holocene terrace. These remains, which do not include pine, are probably all mid to late Holocene.

Circumstantial evidence for the association of woodland clearance and/or vegetation disturbance by human populations and their grazing animals, and slope mass movement, has been a recurrent theme in Holocene geomorphology in the Highlands (Chapter 2). However, it has been difficult to establish causal links. The stratigraphy of Creagan a' Chaorainn adds to the debate because it fails to strengthen the evidence for an anthropogenic trigger. Anthropogenic fires are not supported by current evidence, and an association of fire with pine woodland which died out to be replaced by birch, hazel, rowan and heather is untested by pollen stratigraphy. The significance of the incidence of macroscopic charcoal horizons on Creagan a' Chaorainn is therefore uncertain, although dating pine stumps in an expanse of eroding peat on the col between Gleann Chorainn and Glen Orrin, less than 1 km from the research site could help to resolve the issue.

If vegetation stabilises slopes (Chapter 2, Section 2.2.3), the corollary is that its removal should be associated with stratigraphic evidence for slope instability. If erosion did not follow fires in the past, the stratigraphy should therefore record continuing slope stability. These propositions were tested by examining the relationship between charcoal and overlying mineral and organic deposits in Sections GCh 4 - 9 as set out in Table 5.14. This table reveals no correlation between fires on Creagan a' Chaorainn and slope instability since only two out of seven sequences show the material overlying charcoal to be debris flows. Moreover, slope mass movement was frequent in periods when macroscopic charcoal was absent. Evidence for Creagan a Chaorainn does not therefore support fire as an independent driver of slope instability.

Section GCh	Stratum overlying charcoal
4	2 centimetre thick sandy, silty slopewash overlain
5	by thick peat
5	debris flow
6	debris flow
7a	organic (podzol)
7b	organic (podzol)
8	organic (peat)
9	organic (peat)

TABLE 5.14
The association of charcoal and slope instability on Creagan a' Chaorainn

Episodic failure of weathered bedrock which accumulated as inter-peat mineral layers, originated in the rocky upper slope, and can have been little influenced by slope vegetation. Twigs and roots of woody plants appear in some peat sections immediately below flows of rock debris, showing that rock flows occurred irrespective of the presence of such vegetation on the slope. Charcoal derived from woody plants is absent during periods of stability as well as instability, so no correlation between major slope instability and tree and shrub removal can be inferred. Furthermore, water-driven erosion of thin soils, with or without fire damage to vegetation, was not the precursor of intensified or repeated slope instability. Instead, in Sections GCh4 and 9, eroded soils are overlain by thick, sediment-free peat, which can only accumulate in the absence of slopewash derived from substrate erosion and energetic run-off. Where fires led to thin mineral soil erosion, it was possibly as a result of disruption of a shallow root mat,

which held the friable sandy matrix in place. There is no evidence that this led, in turn, to further, deeper-seated slope mass movement.

If non-random slope evolution was not driven by changes in vegetation cover, alternative hypotheses are that it was driven by intrinsic processes and/or random, intense meteorological events, and/or climate variability. The timing of the various events which contributed to slope evolution is set out in Figure 5.22.

Slope evolution driven primarily by gradual processes intrinsic to the slope system is not well supported by the evidence. Erosion of immature mineral soils did not progress to deeper till failure. Some such events were followed by the spread of sediment-free blanket peat, and a mid Holocene period of slope stability which lasted for two to three millennia (Fig. 5.22). Moreover, debris/groundwater interactions and the observed pattern of a time gap of several millennia between redistribution of Lateglacial frost-weathered debris in gullies and intergully zones and the late Holocene resumption of rock debris production, are open to alternative explanations.

Evidence for hydrological changes on Creagan a' Chaorainn (Table 5.10) can be compared with evidence for slope activity (Fig. 5.22). The late Holocene increase in several types of slope movement is matched by evidence for a sharp rise in water table at 2.5 cal ka BP whose persistence is reflected in 12cm of minimal peat humification (Section GCh4). The most intense leaching and podzolisation occurred in the late Holocene (Table 5.8) and was accompanied by more frequent mass movement. Some intensification of the degree of leaching in podzols from about 6.0 cal ka BP, suggests more persistent humidity but perhaps less intense precipitation. Persistent leaching requires soil matrix stability on a time scale of at least hundreds of years, and there is little evidence of slope movement at this time, or in the subsequent two millennia when palaeogullies were filling with virtually sediment-free peat. Till at this time may have been protected by a complete vegetation cover.

The concentration of flows including bedrock-derived debris in the late Holocene could also be explained by progressive weathering, initiated in the early Holocene, which produced effects only after several millennia, and noticeable only when the masking effect of postglacial relaxation had been attenuated. Since weathering rates remain unquantified, this explanation cannot be dismissed, however, they do not vary independently of climate.

A possible explanation for late Holocene bedrock erosion lies in changes in subsurface flow. The fines in inter-peat debris consist of fresh, angular and unweathered particles of biotite, muscovite and amphibole, all of which are unstable in contact with acid water (Deer *et al* 1966). They cannot therefore be the products of surface weathering in the presence of acid, peaty water. But the steep, fault shattered, dyke-intruded, glaciated bedrock on Creagan a' Chaorainn is susceptible to weathering (Chapter 4, Section 4.2.1) and offers large areas for mechanical and chemical breakdown, both where gullies are incised to bedrock, and in the subsurface, where some infiltrating rainwater has a residence time of at least 6 weeks (personal observation). Subsurface flow could therefore be implicated in causing fragmentation of bedrock.

The principal process of chemical weathering in northern Scotland in the Holocene is hydrolysis whose reaction rates increase with the volume and rate of groundwater flow (Bain *et al* 1993), and mechanical disruption of rock increases with increasing volumes of flow, as does ion exchange (e.g. Erikkson 1985, Robins 1990). Rock weathering and debris generation should therefore be faster when groundwater flow is persistently at a high level. In addition, recent work on the composition of talus slopes in Northern Scotland (Hinchcliffe *et al* 1998, Salt & Ballantyne 1997) suggests that granular disaggregation of rock faces has been a significant contributor to rock debris. The timing and volume of inter-peat debris flows is therefore most straightforwardly related to changes in infiltration. In the rocky recharge zone of Creagan a' Chaorainn the main control on infiltration is precipitation.

This is not to suggest that individual flows of debris are climatic signals. Each event can only be interpreted as a response to the co-incidence of the availability of unstable rock debris and the meteorological conditions capable of triggering its mobilisation. Debris flows originating in masses of disaggregated bedrock could have been triggered either when a destabilised rock exposure suddenly failed, or when a mass of eroded fragments in a temporary store was subject to hydrostatic overpressures. In both cases there is likely to be a time lag between the onset of conditions leading to accelerated rock weathering and debris transport.

Based on average peat accumulation rates quoted in the literature, and calculated here (Section 5.3.4.1), the thinnest peats between layers of rock debris in the palaeogullies may have taken of the order of a hundred years to form. The high relative proportion of non-anthropogenic, late Holocene, rock debris to peat in the upper part of Section

GCh9, together with the reduced thickness of the peats at this time compared to those formed in the mid-Holocene, is consistent with a late Holocene increase in frequency and volume of erosion. In the absence of evidence for an anthropogenic trigger, and since rates of bedrock weathering and erosion are both responsive to enhanced groundwater flow, a climate trigger for enhanced late Holocene mass movement best fits currently available evidence.

Drivers of earlier slope evolution are more ambiguous and probably interlinked. The slow transition from erosion to peat accumulation in the palaeogully system may have been a subtle response to changes in both water and sediment supply, influenced by feedbacks from increasing peat cover, with a feedback loop linking the stabilising effect of blanket peat on till to changes in groundwater and sediment supply. The principal source of run-off in the palaeogullies is the steep upper slope outcrop and the rocky recharge zone above (Fig. 5.4). In the past, sediment-free peat filled gully floors between rock debris flows, presumably moderating high energy run-off. Sediment was incorporated in peat only when the till underlying upslope peat was exposed by erosion. The contemporary environment forms a strong contrast: the modern gully is rock- and sediment-floored with small patches of summer vegetation regularly scoured away by intense flows. Between about 6.2 - (?)3.0 cal ka BP, slope hydrology and sediment supply was generally unfavourable to bedrock and till erosion, but favourable to long term podzol profile development with intensified leaching, peat accumulation and virtually sediment-free run-off. This points to humid conditions, accompanied by full vegetation cover and the absence of both destabilising meteorological events or human impacts.

The timing of release of fines from exposed till can be quite precisely defined in one case - 2.5 cal ka BP - while maximum dates can be established in others. Peat immediately below the surface soil in sample GCh10/2 has an age of about 1.3 cal ka BP. This date marks the boundary between an earlier period when fines were largely unavailable, and a subsequent on when they were. Conditions for this to happen are exposure of sandy till beneath degrading vegetation, especially peat, leading to persistent slopewash, but without major slope disturbance. Further up the slope in Section GCh7a, at a better drained site, the surface soil is a slightly peaty sand, rather than a sandy peat. Between it and the palaeopodzol dated at about 1.7 cal ka BP a 'ghost' (disturbed or immature) soil profile, both developed in the till matrix, required time to

form. So in Section GCh7a, as in Section 10, the sediment in which the modern soil developed was emplaced no earlier than about 1.3 cal ka BP. This date indicates the start of vegetation cover degradation, and therefore possibly the first major human impact. The 'ghost' podzol may have been disrupted by enhanced ground freezing during the late Holocene. Cryoturbation is known to disrupt profile differentiation (Schaetzl & Isard 1996, Haynes *et al* 1997). Another possibility is trampling and wallowing by grazing animals.

There is a lack of information about when modern sandy, (in places peaty) soils, which show no sign of podzolisation, began to form. These soils occupy a 10 - 20 cm thick horizon of unstructured sandy, silty sediment of a type not seen lower in the stratigraphy. Early non-podzolised soil horizons (Sections GCh4 and 9) are thin and immature, and palaeopodzols are developed in little-altered till, not a separated fine fraction. A hypothesis which fits these observations, is that the most recent phase of slope evolution was an unprecedented erosion of fines from till as a result of disruption of the vegetation mat. A contributor to this process may be the trampling by deer and sheep, which can today be seen to have broken up the peat cover. Despite this, there is no accumulation on the modern slope of either rock debris or failed till. As in the past, the environmental conditions which today promote erosion of fines, do not appear to be those which promote larger scale failure of till and bedrock.

5.5   SUMMARY AND CONCLUSIONS

The pattern of debris flows, till failures and soil erosion on Creagan a' Chaorainn was non-uniform, and apparently non-random. Slope materials and processes record responses to complex interactions of intrinsic and extrinsic environmental conditions. The principal processes for which evidence has been found on Creagan a' Chaorainn are rock debris generation and transport, till failure, repeated podzol formation and erosion, soil erosion, fluctuations in the water table, fire, and changes in vegetation cover. Changes in slope hydrology and slope stability through time and interacting triggering mechanisms have been inferred. The data have been brought together in the form of a broadly dated pattern of change, which however, lacks the precision inherent in the concept of an event stratigraphy.

Postglacial relaxation, progressive intrinsic processes, human activity, and fire-induced changes in vegetation cover, are unable to account for variable rates of rock debris production, but do provide a convincing explanation for other aspects of slope evolution. The primary control on slope wash of fines up to small pebble size appears to have been the integrity of the vegetation mat. Once established, blanket peat was probably a controlling factor in slope stability, with its disruption by animal trampling at some point after 1.3 cal ka BP exposing till to erosion of fines. The primary control on release of bedrock-derived debris may reflect hydrological changes. Although individual flows of bedrock-derived debris and till may be random events, which like till and podzol failures are responses to individual meteorological events, their frequency appears to have increased with rising humidity and water table.

The Holocene stratigraphy of Creagan a' Chaorainn broadly reflects the processes of postglacial relaxation, slope stabilisation in humid, but non-erosive, conditions, renewed destabilisation in a late Holocene environment when rates of rock weathering were enhanced, and finally blanket peat degradation.

1. Before about 7.5 cal ka BP conditions were inimical to preservation of organic horizons. Mass movement arising from slope adjustment to deglaciation, combined with relatively warm, dry, oxidising conditions may account for this feature.

2. After this time, the reduction of till reworking and transport in gullies, combined with an absence of intense erosive flows, allowed peat to replace mineral sediment on gully floors. Only modest-scale changes in slope hydrology may have been required to effect this change in system state.

3. Gradual peat accumulation, virtually complete vegetation cover and rare erosive events indicate slope stability which lasted until the late Holocene (possibly about 3 ka BP).

4. The late Holocene saw a rise in slope water table, and, in palaeogullies, an increase in the volume and frequency of debris flows which originated in weathered bedrock.

Podzol horizons were truncated by shallow slides and debris flows, and thin peaty soils were disrupted.

5. In historical times, possibly since the 'Little Ice Age', widespread disruption of peat by grazing animals, exposed the till to erosion of fines, creating a superficial deposit in which modern soils are developing.

In conclusion, the stratigraphy of Creagan a' Chaorainn shows the Holocene evolution of this typical slope in the northern Highlands to have been lengthy, episodic and complex. The evidence suggests a climatic driver combined with the intrinsic effects of blanket peat and human activity. However, the resolution achievable from this study is subject to a variety of cumulative uncertainties.
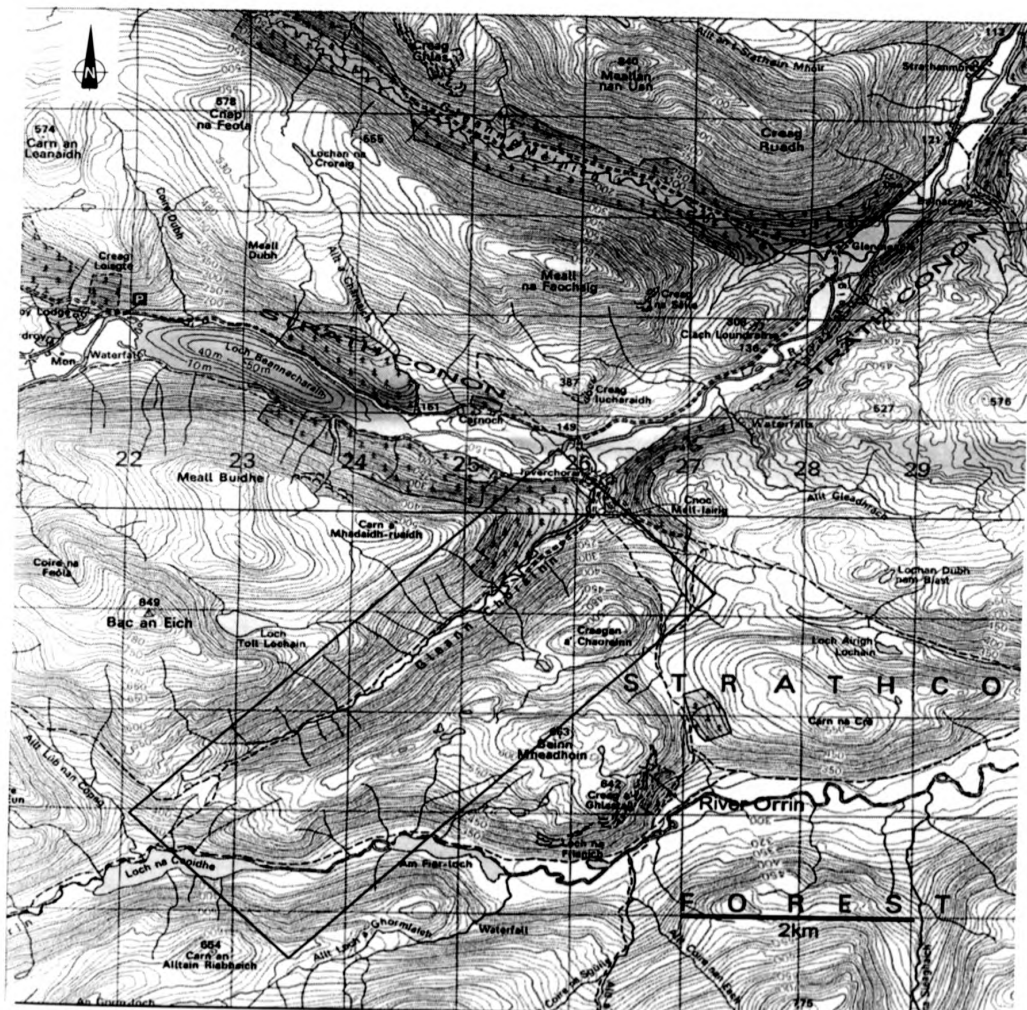
Based on the OS 1:50000 scale map © Crown copyright

FIGURE 5.1
Creagan à Chaorainn : location map
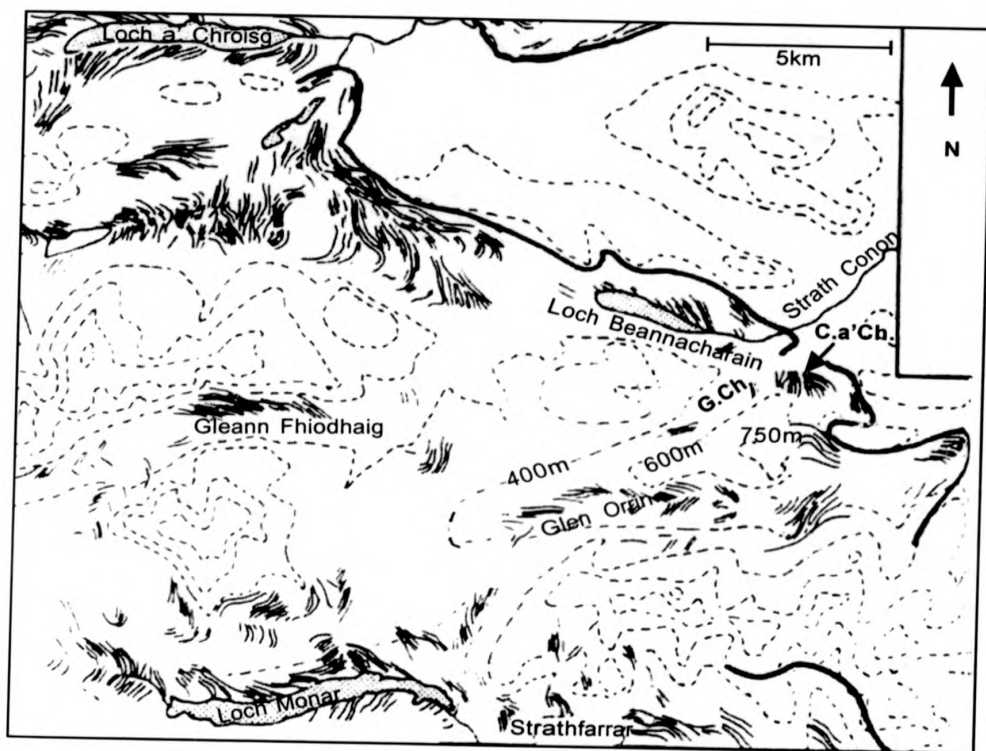
FIGURE 5.2
**The extent of Younger Dryas Stadial ice in Strath Conon, Gleann Chorainn
andsurrounding area**
(Adapted from Bennett 1994, Figure 7)

KEY

C.a' Ch. = Creagan a' Chaorainn and associated moraines

G. Ch. = Gleann Chorainn

〜 limits of Younger Dryas ice

≪ individual icefronts
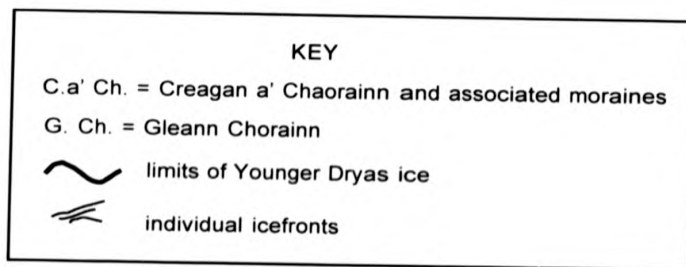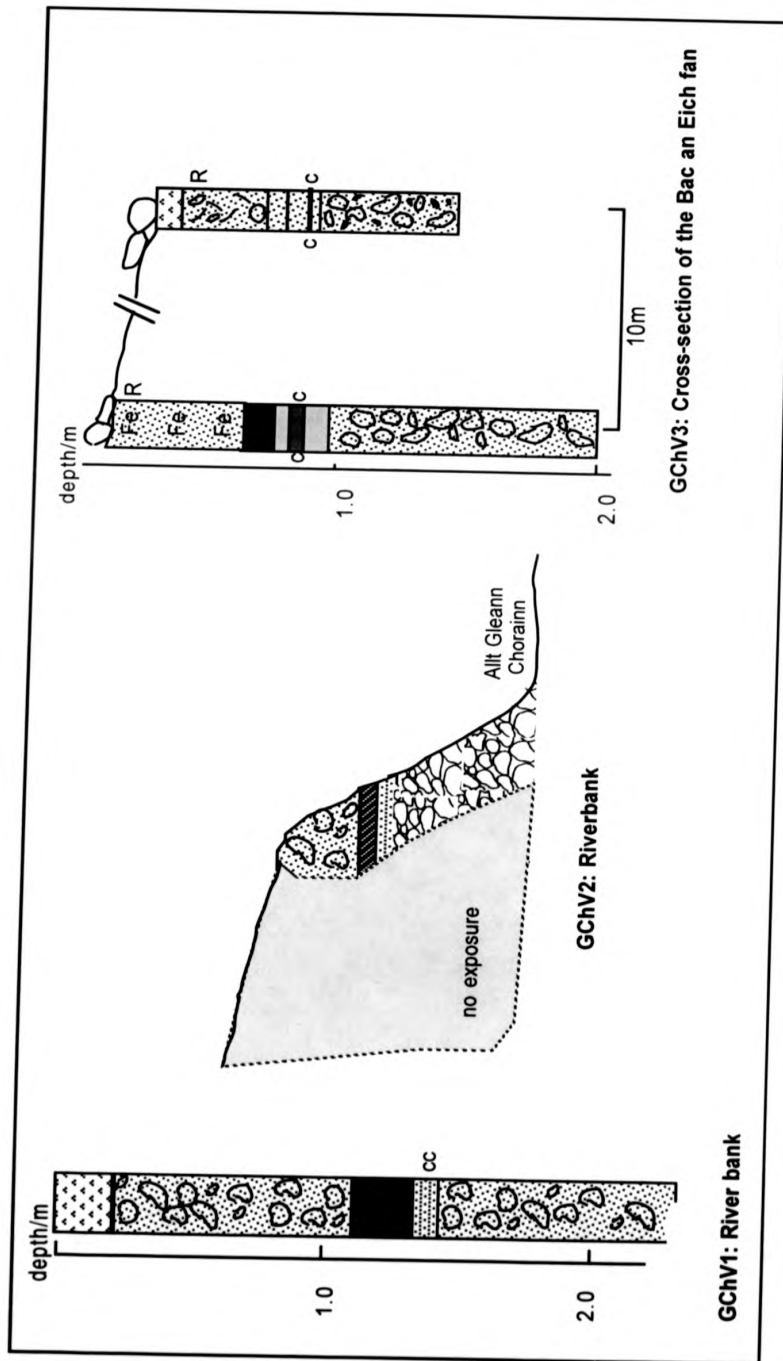
FIGURE 5.3
Gleann Chorainn Sections (undated):
Postglacial remodelling of lower slopes
Locations are marked on Figure 4.2
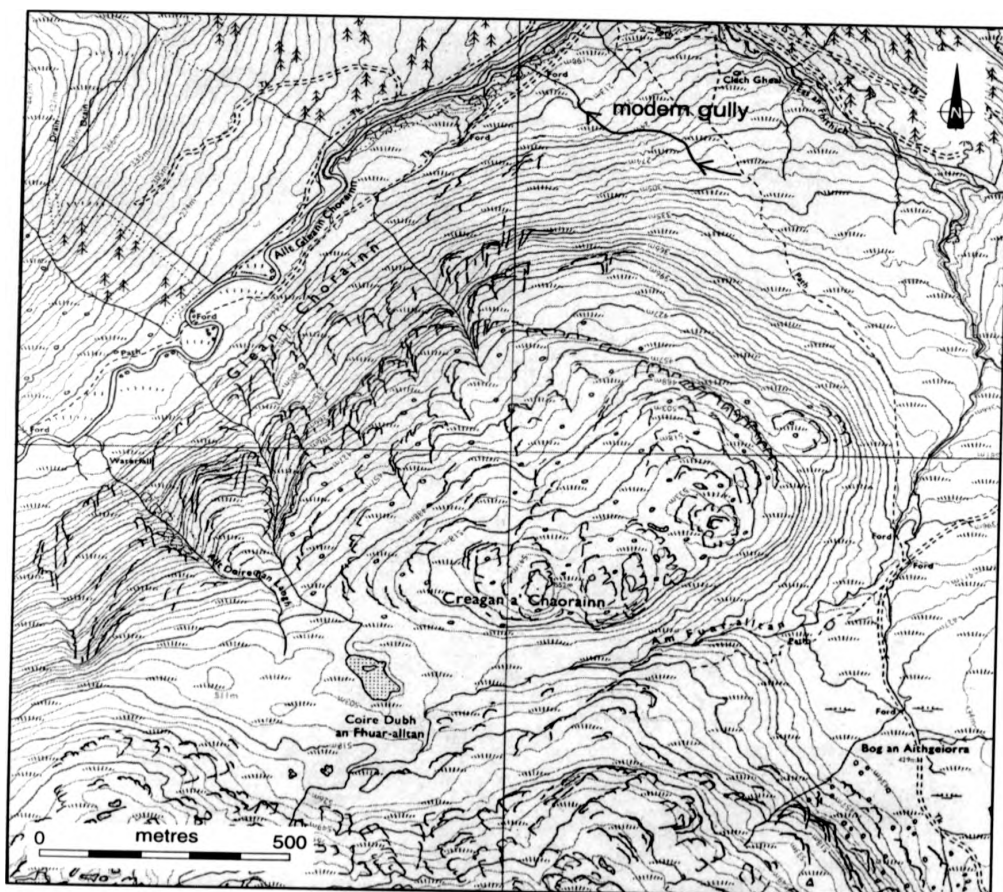
Based on the OS 1:10000 scale map © Crown copyright

FIGURE 5.4
Creagan à Chaorainn showing the post - 1970 gully
in which dated Sections are located.
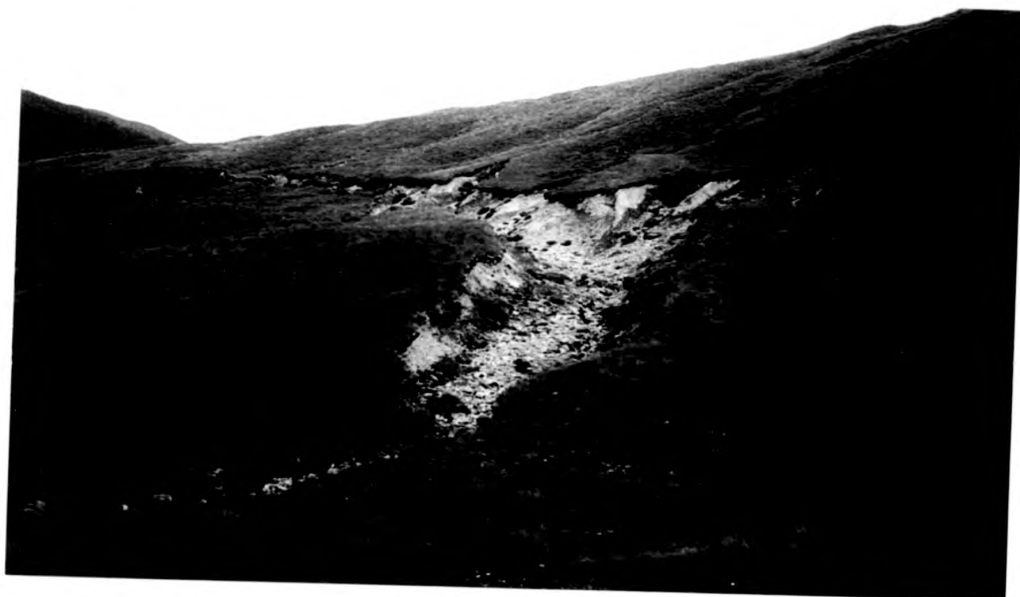
FIGURE 5.5a
Dissection of the palaeogully system by the modern gully, Creagan a' Chaorainn
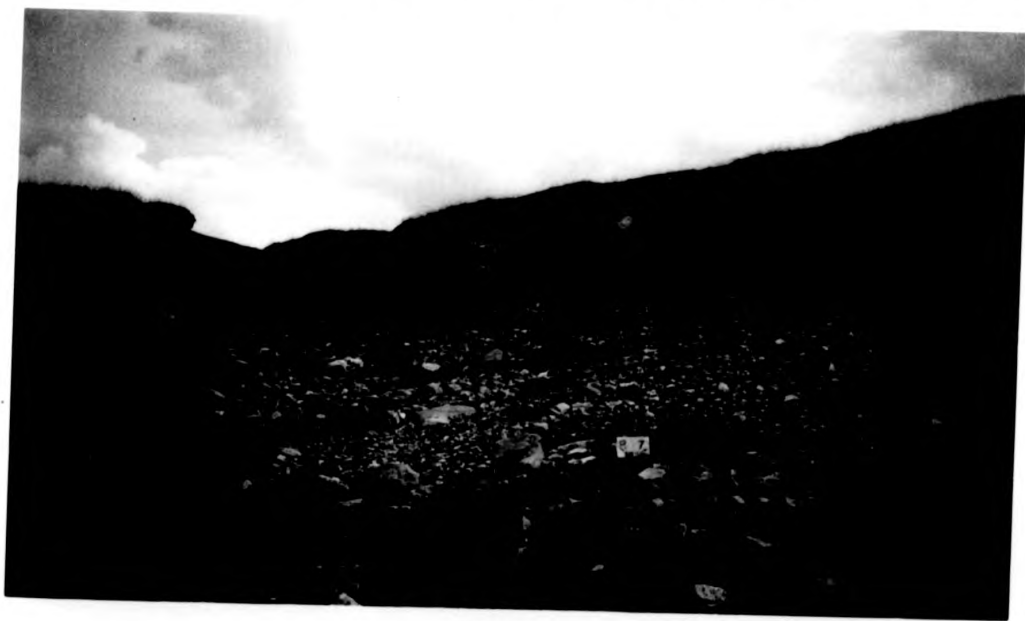
FIGURE 5.5b
The modern gully, Creagan a' Chaorainn

FIGURE 5.6
The palaeogully system, Creagan a' Chaorainn
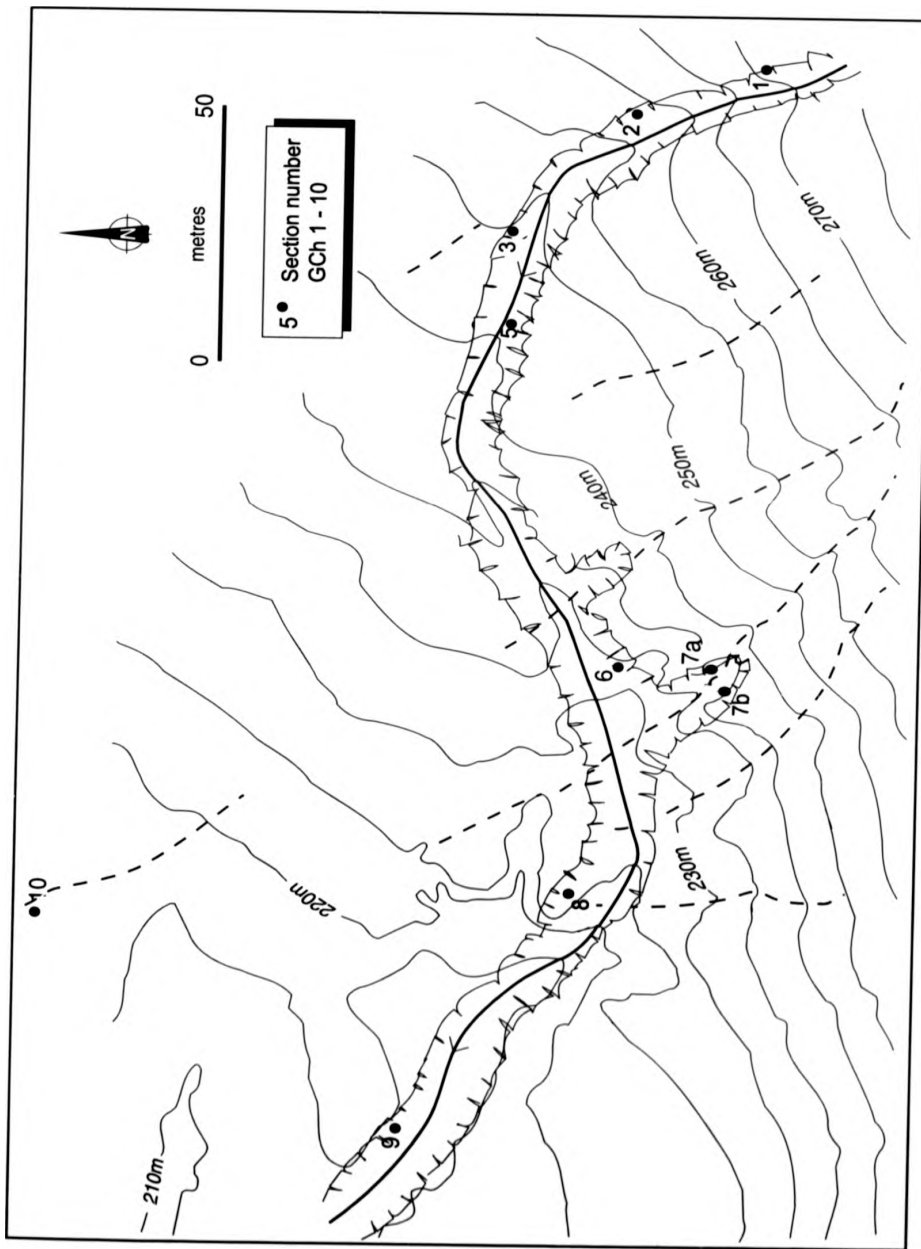
FIGURE 5.7

Creagan a Chaorainn: The modern gully and logged sections GCh 1-10. Palaeogullies (dashed lines) trend north-west.
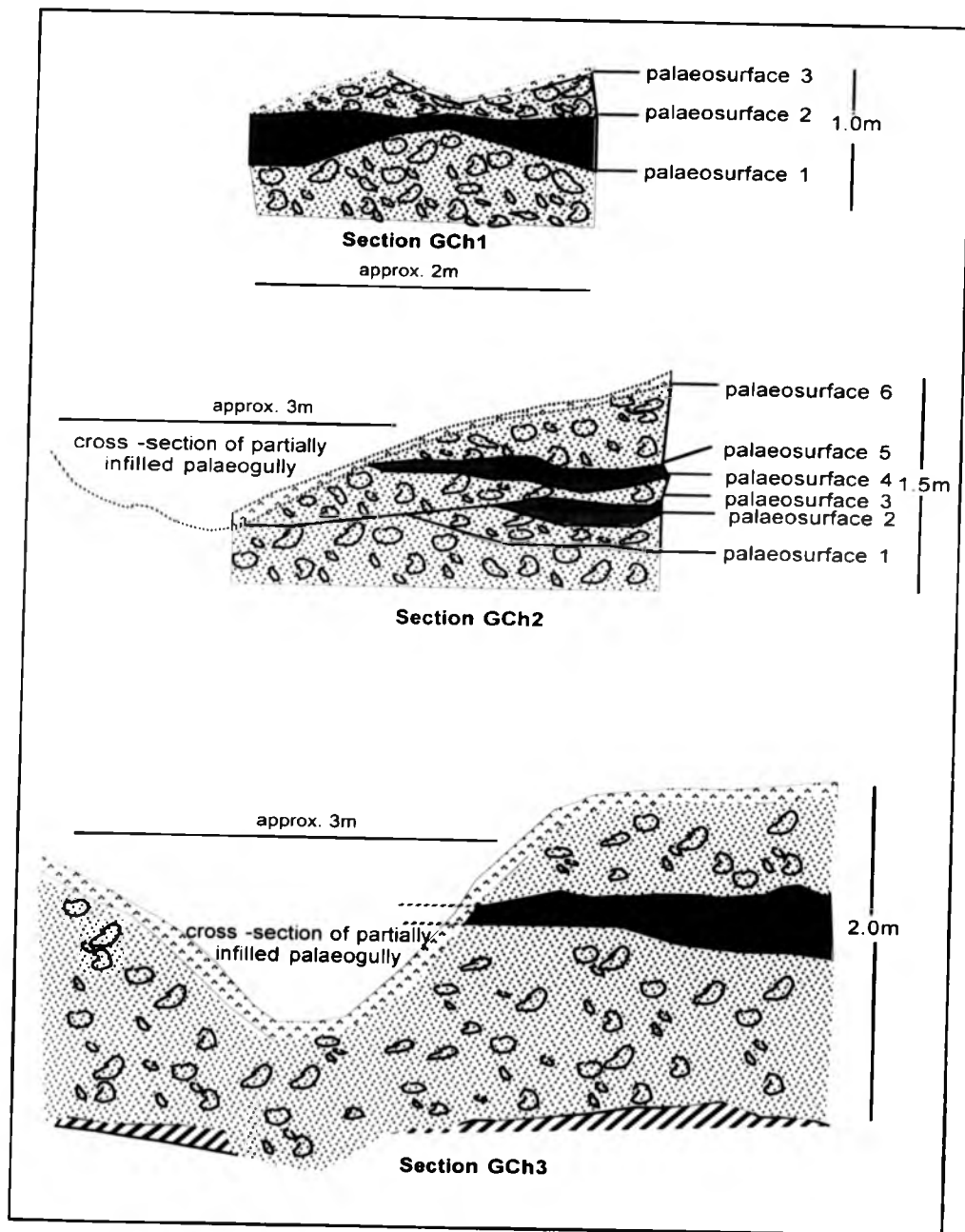
FIGURE 5.8
GCh Sections 1,2,&3 (Undated: locations marked on Fig. 5.7)
Field sketches of repeated, small scale, postglacial, mid slope, surface remodelling,
visible in the walls of the modern gully
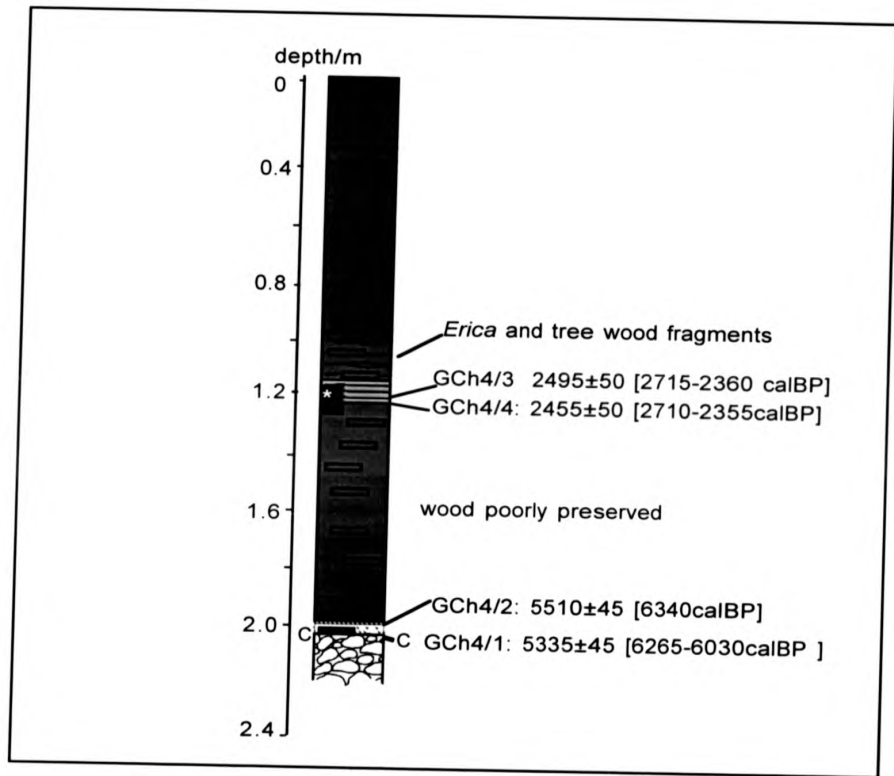
FIGURE 5.9
Creagan a' Chaorainn, Section GCh4
■ Kubiena tin sample & photomicrograph

depth/m labels on figure:
- 0
- 0.4
- 0.8
- 1.2
- 1.6
- 2.0
- 2.4

Erica and tree wood fragments

GCh4/3   2495±50 [2715-2360 calBP]
GCh4/4: 2455±50 [2710-2355calBP]

wood poorly preserved

GCh4/2: 5510±45 [6340calBP]
GCh4/1: 5335±45 [6265-6030calBP]
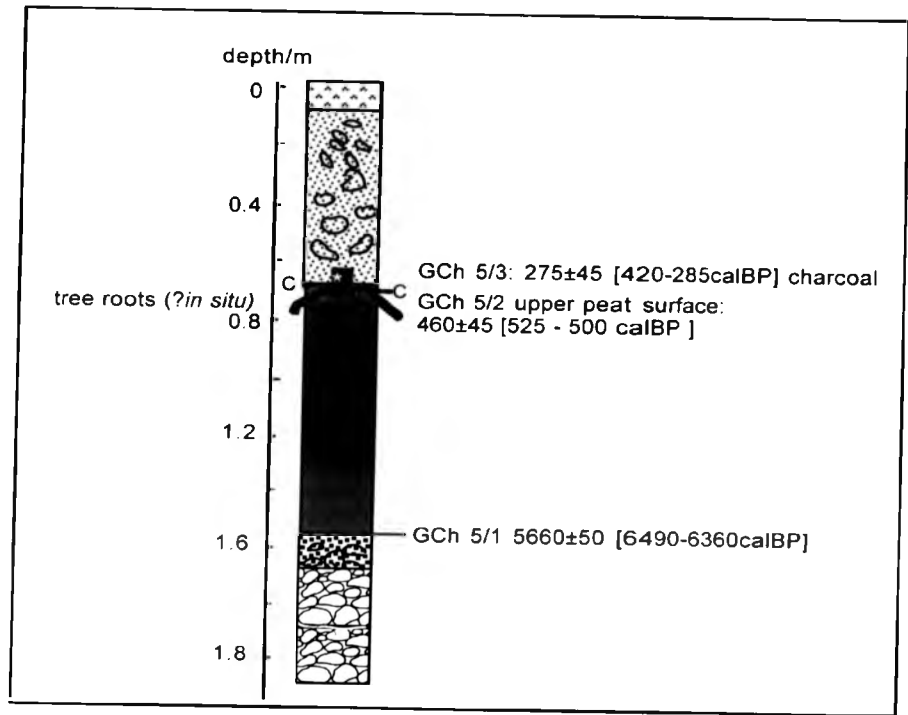
FIGURE 5.10
Creagan a' Chaorainn: Section GCh5

Kubiena tin sample & photomicrograph

Figure labels:

depth/m

0

0.4

GCh 5/3: 275±45 [420-285calBP] charcoal
GCh 5/2 upper peat surface:
460±45 [525 - 500 calBP]

tree roots (?*in situ*)
0.8

1.2

GCh 5/1 5660±50 [6490-6360calBP]
1.6

1.8
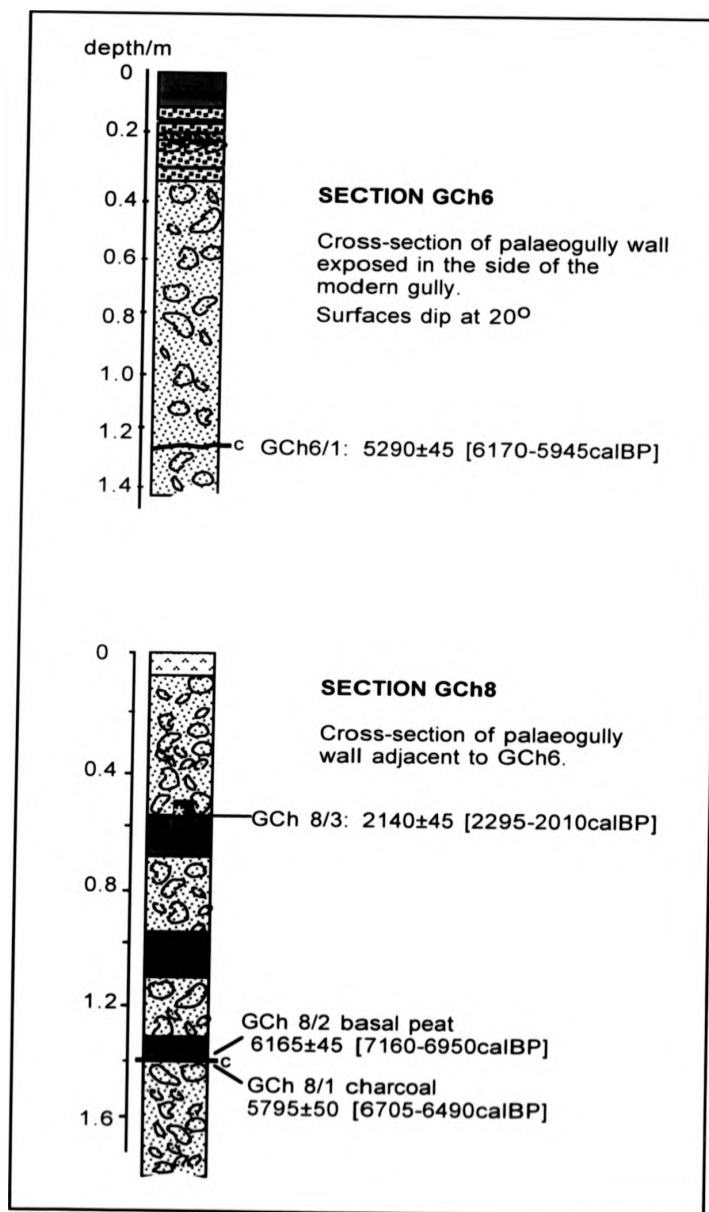
FIGURE 5.11
Creagan a' Chaorainn Sections GCh 6 & 8

■ Kubiena tin sample & photomicrograph

The diagram labels:

depth/m

SECTION GCh6

Cross-section of palaeogully wall exposed in the side of the modern gully.
Surfaces dip at 20°

GCh6/1: 5290±45 [6170-5945calBP]

SECTION GCh8

Cross-section of palaeogully wall adjacent to GCh6.

GCh 8/3: 2140±45 [2295-2010calBP]

GCh 8/2 basal peat
6165±45 [7160-6950calBP]

GCh 8/1 charcoal
5795±50 [6705-6490calBP]

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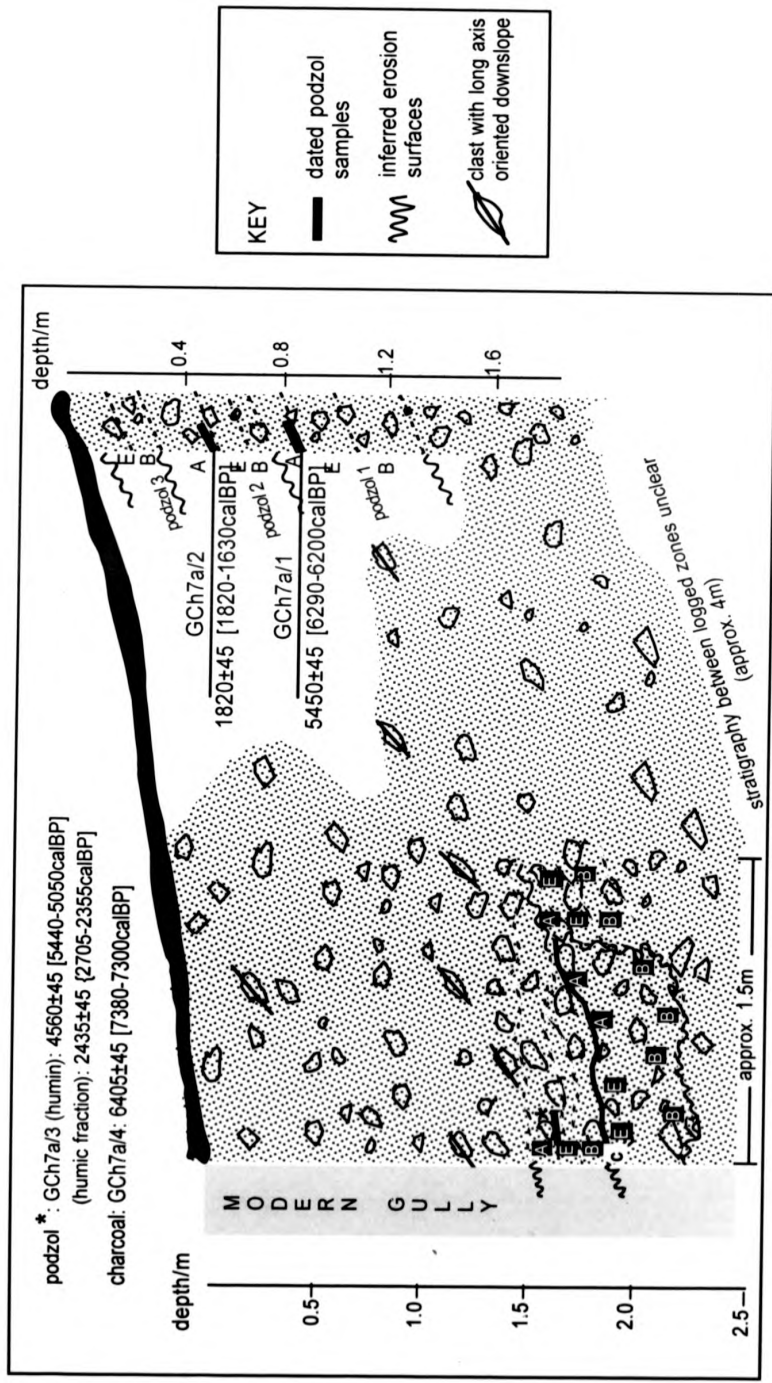
FIGURE 5.12

Section GCh7a Creagan a' Chaorainn (not to scale)

View of wall of short, headward-eroding gully nearly perpendicular to the modern gully and opposite Section GCh7b (see Fig. 5.7). Exposure is continuous, but stratigraphy between the zones represented is obscured by shearing.
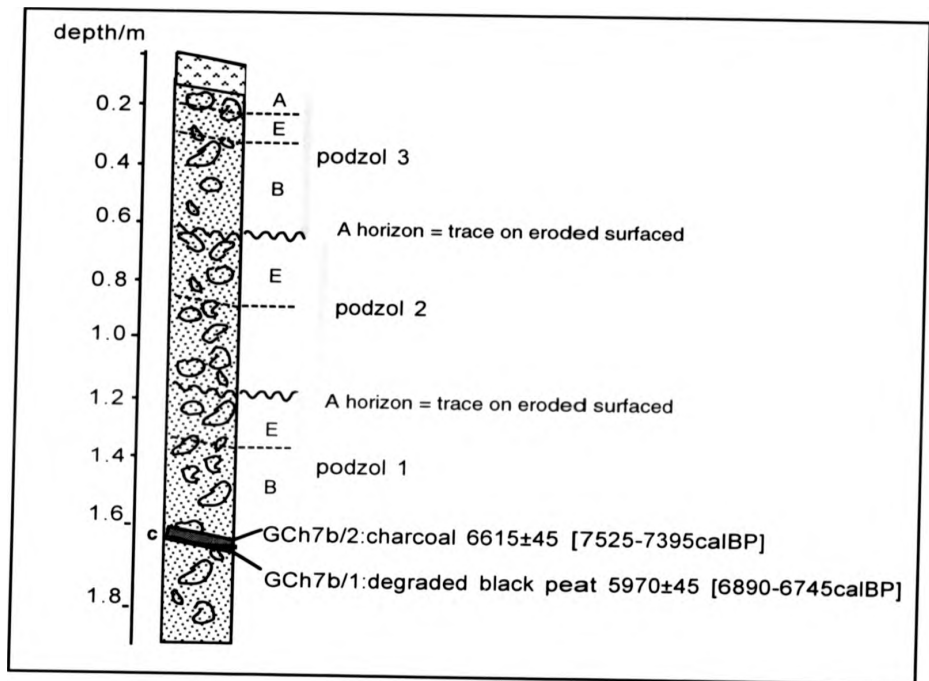
FIGURE 5.13

Creagan a' Chaorainn: Section GCh7b
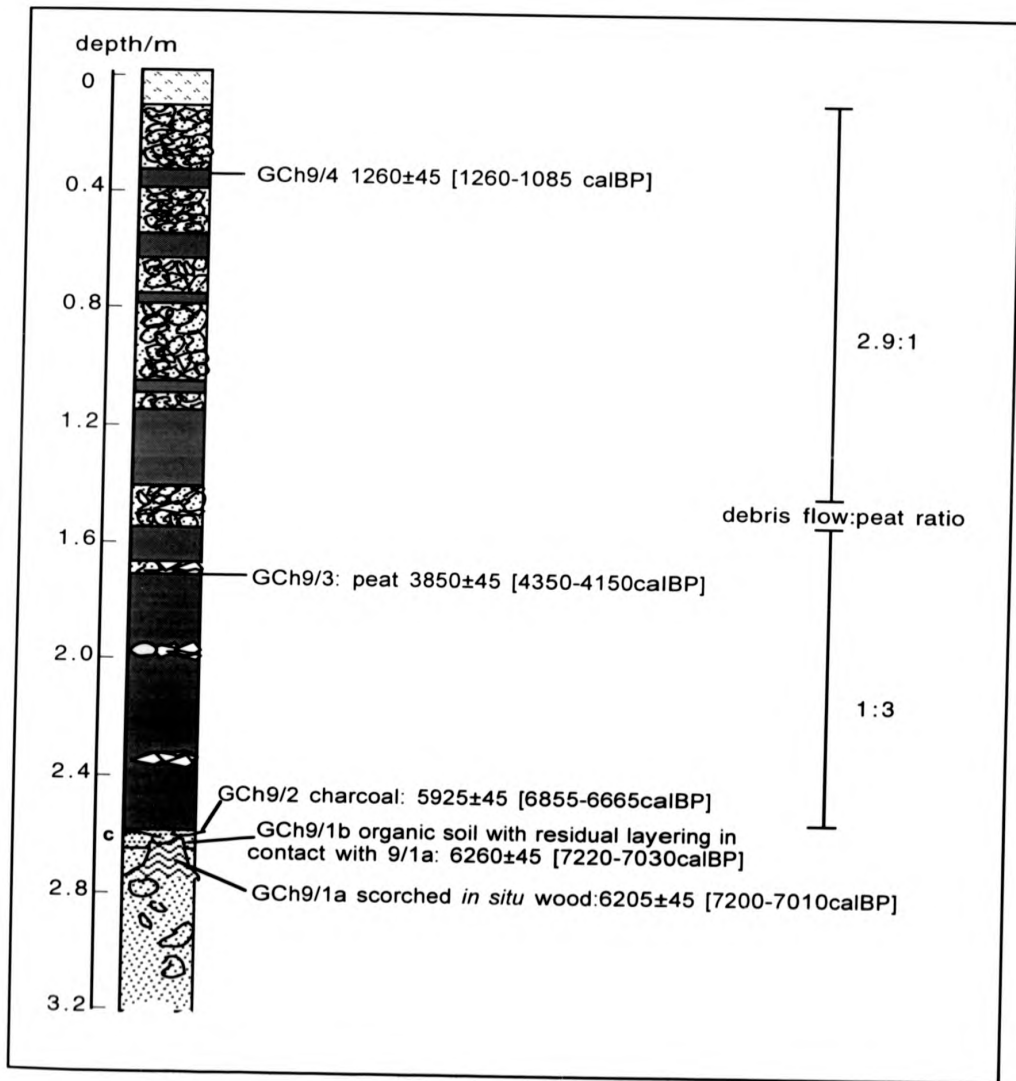
FIGURE 5.14

Creagan a' Chaorainn: Section GCh9
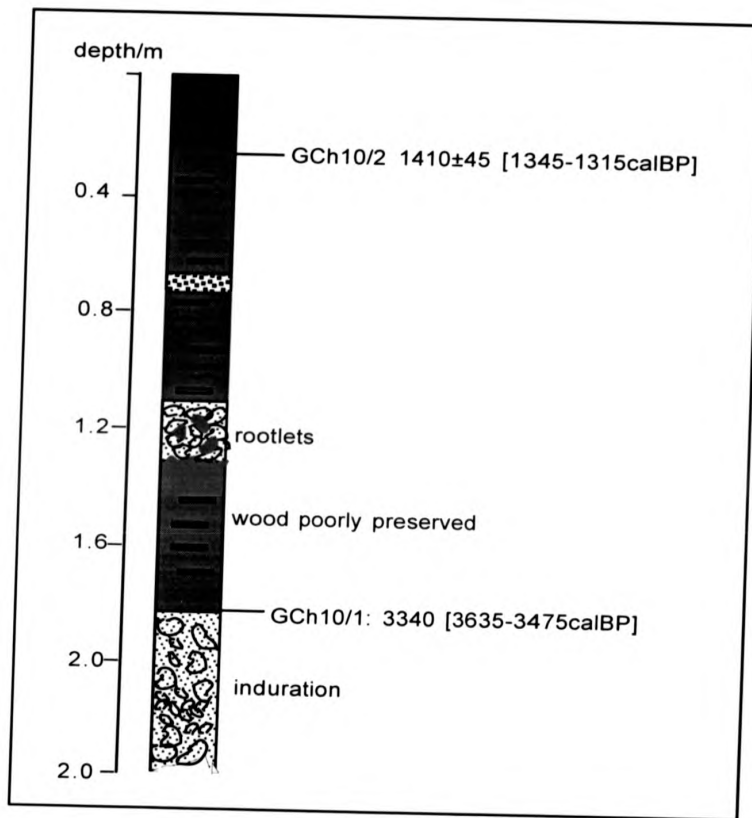
FIGURE 5.15

Creagan a' Chaorainn: Section GCh10

FIGURE 5.16
Creagan a' Chaorainn: Charcoal layer in till, Section GCh7a

FIGURE 5.18a (x 20 PPL) GCh4
Mid brown AOM and a woody cross-section in the lower peat
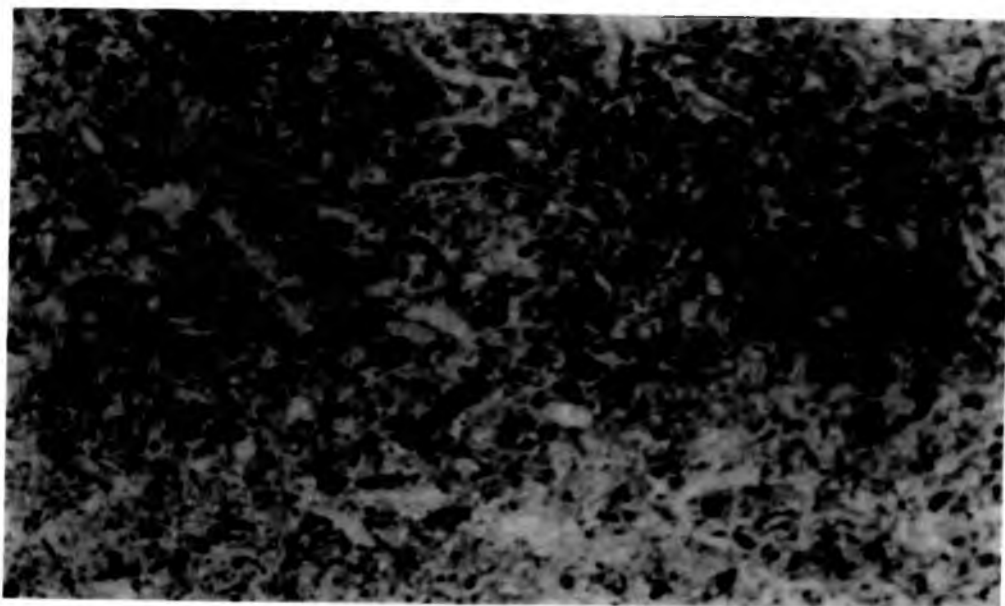
FIGURE 5.18b (x 20 PPL) GCh4
Poorly humified upper peat; scattered angular mineral grains & faecal pellets(?)

FIGURE 5.18c   (x 20 PPL) GCh4
Horizontal woody particles in mid brown OM at the upper surface
of the lower, well humified peat

FIGURE 5.18d (x 20 XPL) GCh4
The sharp boundary between upper and lower peats; numerous mineral grains and
well preserved plant stems above, well humified peat with woody fragments below.
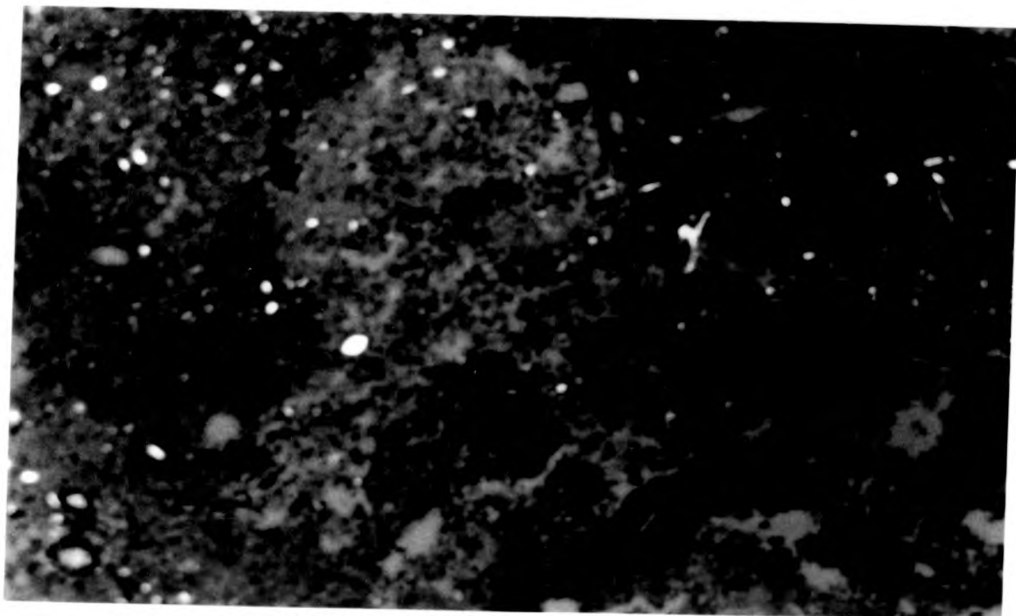
FIGURE 5.19a (x 20 P/XPL) GCh5
Well humified peat with mid brown AOM, partially decomposed woody particles,
and sparse silt and fine sand
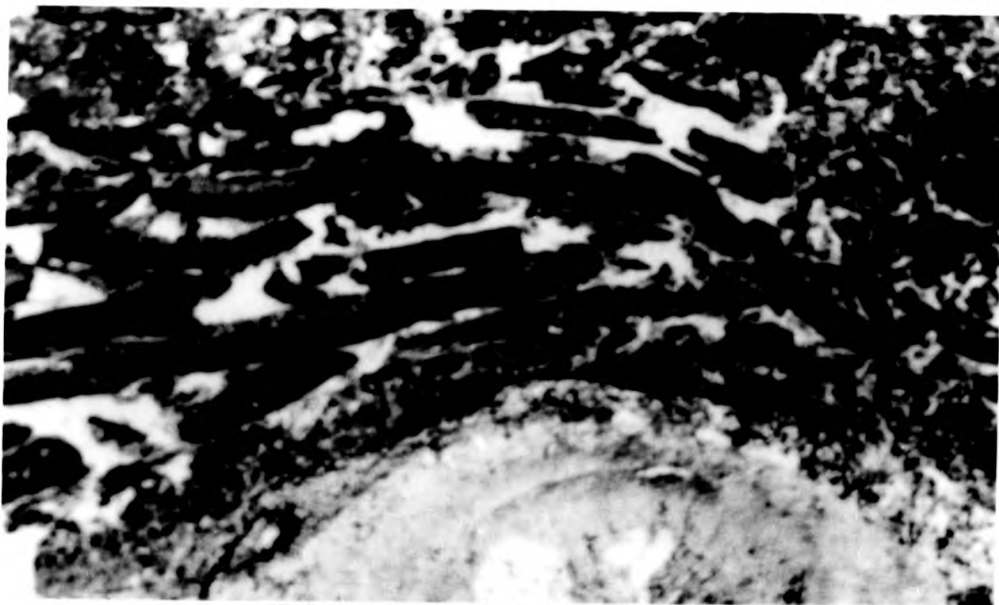
FIGURE 5.19b (x 20 PPL) GCh5
Charcoal and unburned woody tissue at the surface of the peat

FIGURE 5.19c (x 20 PPL) GCh5
Peat with amorphous woody tissue invaded by mineral grains
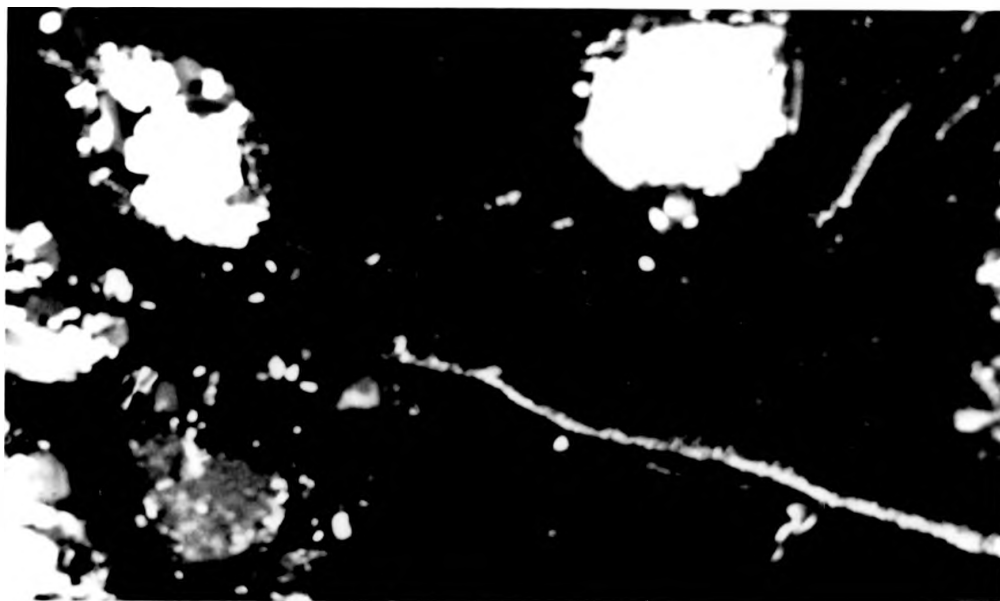
FIGURE 5.19d (x 100 PPL) GCh5
Close up view of mixed plant material and angular sand and silt
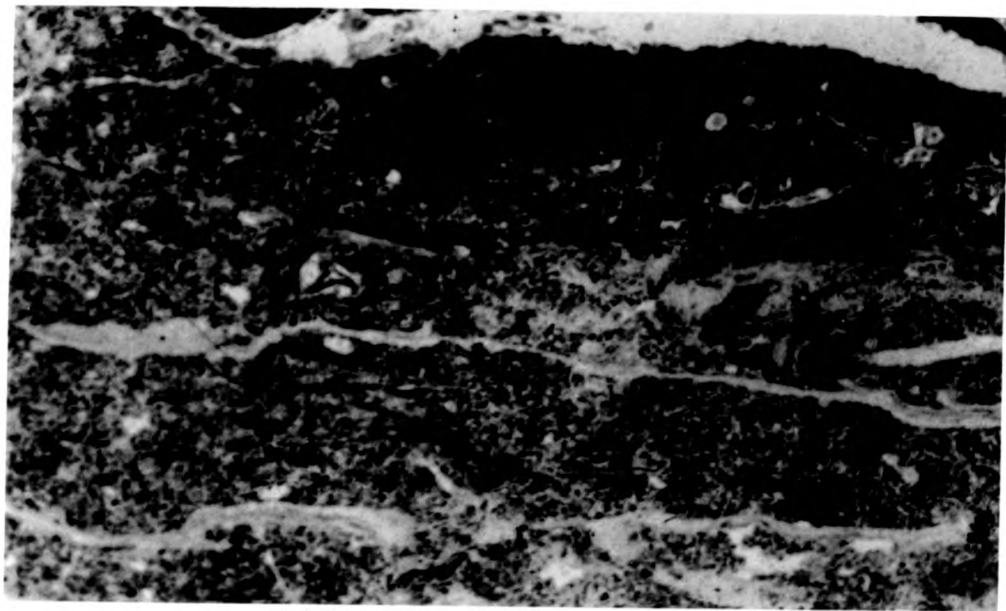
FIGURE 5.20a (x 20 PPL) GCh8
Peat consists of mid brown AOM with decomposed woody particles
and elongate horizontal vugs. Silt content is sparse.
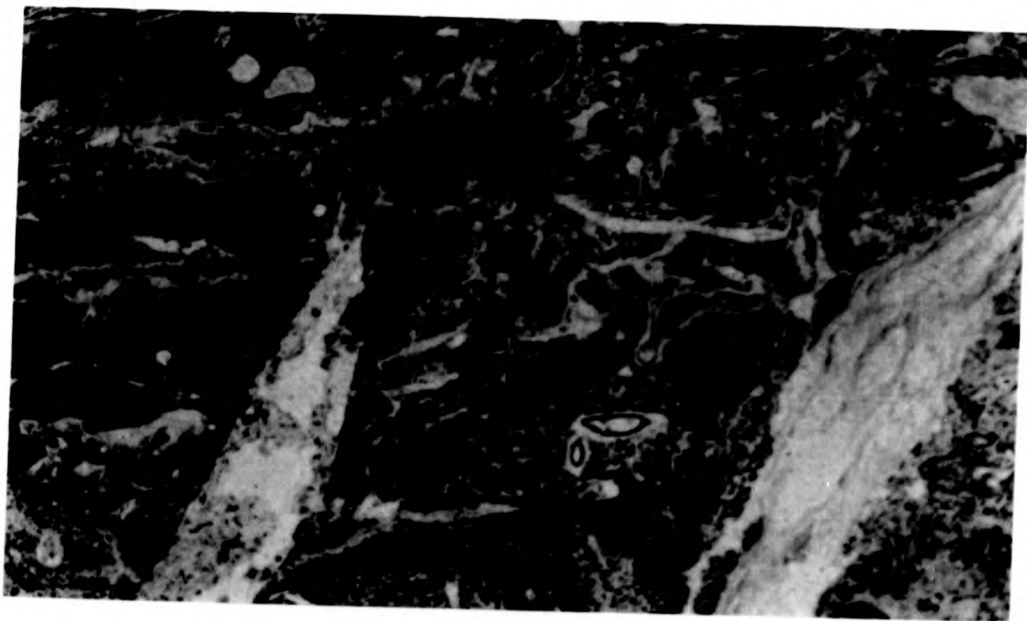
FIGURE 5.20b (x 20 PPL) GCh8
Peat with chaotically folded vugs and vertically rotated structures

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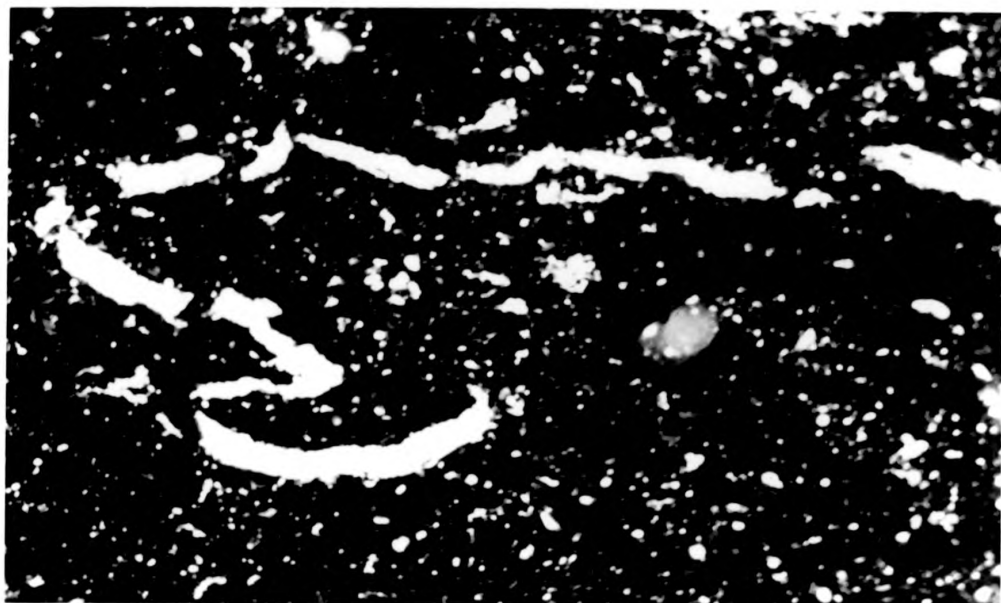
FIGURE 5.20c (x 20 XPL) GCh8
Light brown AOM permeated with angular quartz sand and silt.
A strand of plant material has been folded and pulled apart.
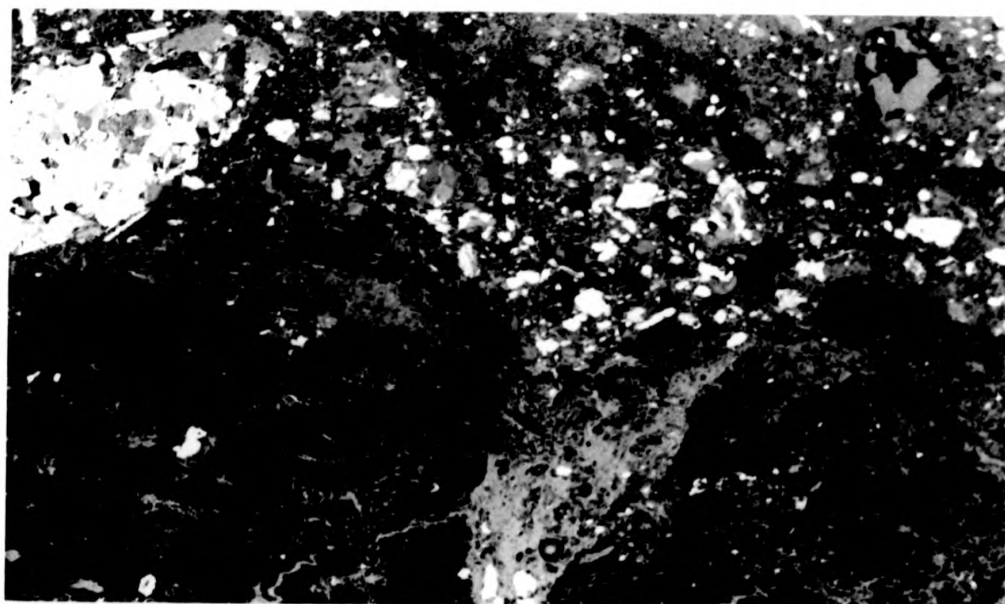
FIGURE 5.20d (x 20 P/XPL) GCh8
A wedge of sandy, light brown AOM has been driven down into the darker brown
peat, breaking the tip of a woody particle.

FIGURE 5.21
Palaeopodzols in Section GChS7a

FIGURE 5.22   DATED CHRONOSTRATIGRAPHY, CREAGAN A' CHAORAINN

CHAPTER 6: LANDSCAPE EVOLUTION IN UPPER GLEANN LICHD

6.0   INTRODUCTION: GLEANN LICHD

Gleann Lichd (the valley of the big rock slabs in Gaelic) has been glacially excavated in a NW-SE fault zone of late Caledonian age (Johnstone & Mykura 1989), terminating close to the north-west coast (Fig. 6.1). Both Gleann Lichd and its larger, parallel, sister valley, Glen Shiel, descend from the north-south watershed to sea level. Between them is a line of hills (The Five Sisters of Kintail) which rise abruptly from sea level to over 1000m. The rugged slopes falling steeply to sea level are typical of the west coast of Northern Scotland and generate rainfall ≥2000 mm per year at low levels, due to moisture-laden south westerlies from the Atlantic. On the summits, precipitation may be double this. The trunk section of Gleann Lichd is only 6.5 km long, and valley floor width is a maximum of 250m. The catchment has an area of 30 km^2. A high proportion therefore consists of long, steep slopes (Figs. 4.5, 6.1). The valley is drained by the River Croe which finds an exit to a sea loch, Loch Duich, round the margin of an ice-smoothed rock bar rising 20-30m above the aggraded valley floor.

6.1   HISTORY AND SETTING OF THE RESEARCH AREA

The dated sections (NH 005174) are near the head of the trunk valley, in the vicinity of Gleann Lichd House (Fig. 6.2). The elevation is only about 45m AOD, and the valley floor is less than 100m wide. A few hundred metres upstream the aggraded valley floor terminates against a rock barrier incised by a subglacial gorge.

During the Younger Dryas Stadial, ice filled the valley, while the uppermost slopes were affected by severe periglaciation (Bennett and Boulton 1993, Bennett 1994). Till fills pockets on the steep slopes between rock ribs and bosses, where bedrock is well exposed. Clusters of aligned moraines are preserved only at the head of the valley upstream of the dated sections, typically on raised ground at tributary confluences (Fig. 4.5). However, extensive fragments of kame terrace and, less frequently, glaciofluvial terrace, were seen downstream during fieldwork.

Bedrock in upper Gleann Lichd is predominantly steeply foliated acid gneiss with some flaggy psammitic, sometimes semi-pelitic, schists (pers. obs.). Major joint sets run both parallel to, and at high oblique angles to, the trend of the valley. Foliation, and one joint set, dip roughly parallel to the southern valley wall where fans are more common. On the north valley wall, the dip of these structures creates a series of minor benches which

form sediment stores. In Figure 4.5 joint sets appear as valley-parallel discontinuities. By contrast, on the southern side, vertical gullies incised in the foliation planes are far more common. The gullies occupy excavated joints, and have acted as conduits for debris which forms fan-shaped deposits at their bases. Large, high level, probably Lateglacial landslips of periglacial debris are marked by gullies, which coalesce at the base of the zone of denudation. On the mid and lower slopes, there is a scattering of much shorter deglacial gullies parallel to the steepest slope angle.

6.2   THE RESEARCH AREA

The main features investigated in detail were two adjacent, though not overlapping, fans (Fig. 6.2). A small fan with a maximum width of 70m lies immediately upstream of Gleann Lichd house. The 'large fan' immediately downstream has a width of 400m, and a long section of 440m. On the opposite side of the valley from the fans is a large slope failure which provided access to undisturbed till and recently exposed bedrock (Fig. 6.3).

The gullies feeding the two fans skirt the upstream and downstream margins of a boss of rock which acts as a shield for the house below, and separates their feeder gullies (Fig. 6.2). The slope into which the gullies are cut falls irregularly to the valley floor at angles of 25-50° from the >1000m ridge between Gleanns Lichd and Shiel (Fig. 6.1). It has a high proportion of outcrop, alternating with patchy, silty-sandy till, with a vegetation of low grasses and sedges. The source gully for the large fan originates at over 650m, has bedrock and till walls up to 7 - 8 metres high, and a width of 4 - 5 metres. The small fan is fed by two gullies which originate at a height of 400-450m above the river. The base of both fans is about 45m AOD, at the margin of the River Croe.

The surface of the large fan is grassy with scattered boulders, some over 1m in diameter (Figs. 6.4, 6.5). Contour spacing (Fig. 6.6) shows three vertical zones distinguished by gradient. The upper fan runs with a gradient of 28° for about 150m from the apex at 210m to the 120m contour. Below that, the mid fan gradient is 22° down to the 75m contour. The surface of the lower fan lies initially at 15° but lies mainly at an angle of 10°. From the apex, a number of debris flows paths with marginal levées lead to the mid and lower fan, which are indented by a few broad shallow, impersistent water channels. There is no incising stream. The structure terminates on the contemporary floodplain in

a low ramp of coarse debris with no sign of distal fining, displacing the line of flow of the River Croe (Fig. 6.2).

The small fan has a contrasting structure and morphology. Its apex is little above that of the adjacent glaciofluvial (?) terrace. Its grassy surface, which falls towards the river at an angle of <10°, lacks boulders, and is deeply incised by a small stream marked by a linear indentation on Figure 6.5.

The dated sections are in the south bank of a meander of the River Croe which has exposed a cross-section of the small fan (Fig. 6.5: GL1), and a distal, longitudinal, upstream section of the large fan (Figs. 6.4, 6.6: GL2, GL3, GL4). Section GL5 samples the fan/floodplain boundary immediately beyond the low, bouldery slope break formed by the toe-ramp of the large fan. In addition, two pits (Figure 6.6) were dug to sample materials below the unincised surface of the large fan. Radiocarbon dates from the logged sections are listed in Table 6.1 below. Issues relating to the precision of these dates are largely those reviewed in Chapter 5, and the discussion is restricted to additional considerations.

Lab Code	Sample	Material	^{14}C age	cal BP	^{13}C/‰
SRR-6040	GL1/1	charcoal (humin)	6590±70	7525-7390	-27.0
SRR-6041	GL1/2	charcoal (humin)	3675±50	4083-3890	-27.4
SRR-6042	GL2/1	peat (humin + humic acids)	4000±45	4515-4415	-28.7
SRR-6043	GL2/2	peat (humin + humic acids)	3500±45	3830-3645	-29.4
SRR-6044	GL2/3	peat (humin + humic acids)	830±45	785-675	-29.0
SRR-6045	GL3/1	peat (humin + humic acids)	3925±45	4420-4260	-30.2
SRR-6046	GL3/2	peat (humin + humic acids)	3140±45	3380-3270	-29.5
SRR-6047	GL3/3	peat (humin + humic acids)	1250±50	1260-1080	-29.2
SRR-6048	GL4/1	wood (humin)	4045±45	4570-4420	-29.8
AA-2723	GL4/2	peat (humin)	2435±60	2710-2350	-30.0
AA-27235	GL4/3	peat (humin)	1330±60	1230-1175	-29.6
SRR-6049	GL5/1	wood (humin)	230±45	300-0	-31.0
SRR-6050	GL5/2	peat (humin + humic acids)	515±45	545- 510	-27.9

TABLE 6.1
Radiocarbon and calibrated dates: Upper Gleann Lichd

6.3   RESULTS

The following sections describe the nature of mineral sediments found in both fans and in the till failure and river channel deposits (Section 6.3.1); the stratigraphy and morphology of the small and large fans and the floodplain (Sections 6.3.2, 6.3.3 and 6.3.4); and the micromorphology of peat-debris flow boundaries (Section 6.3.5).

6.3.1 Sediment analysis

The texture, grain size and mineralogy of contrasting sediment facies - fan debris, till, and matrix-rich river gravels - were analysed. Procedures were described in Chapter 2.

6.3.1.1 Till

There are no fresh exposures of till adjacent to the fans. The till sample analysed is from the large failure on the opposite bank of the river (Fig. 6.3) which appears to be still adjusting to present day conditions since its toe was seen to be reworked by the river after persistent heavy rain.

The till contains scattered clasts 200 (rarely 300) mm in diameter, markedly smaller than fragments of the disrupted bedrock it overlies. Its texture and composition are distinct from shattered bedrock only about 3m above rockhead. Since taking samples large enough to be representative of these well dispersed larger clasts was impractical on the steep (34-50º), unstable slope, only samples of the silt to granule fraction of the till were collected. Sampling was done near the top of the slip to avoid any mixing with fragmented bedrock and/or colluvium. In addition to till, the exposure contains jaggedly shattered rock fragments of a wide range of sizes up to an estimated two tonnes in weight. These, together with a range of smaller blocks, are *in situ* bedrock which had been shattered but not displaced. Interstices were filled with till, producing a massive, unlithified 'breccia'.

Particle size analysis is shown in Table 6.2, Column 1 overleaf. Figure 6.7 presents the analysis as a barchart and is included on the same page for convenience of reference. Nearly half the sediment volume is fine sand and silt devoid of organic matter. Grains with a diameter of >0.2 mm were predominantly rock fragments of gneiss, psammite and semi-pelite, matching local bedrock mineralogy. The sand/granule fraction contained fresh particles of chemically unstable mica and feldspar. Percentages of quartz grains derived from physical and chemical weathering of rock fragments approached 50% by volume only in the fine sand and silt fraction.

Above fine sand grade, particles were predominantly sub-angular. A probable explanation is the edge-rounding of clasts in traction at the former glacier bed, and/or in ice-contact meltwaters. A small accumulation of large (>300 mm) angular boulders was found at the toe of the slip. Given that the maximum size of clasts embedded in undisturbed till was <300 mm, their source may have been shattered bedrock.

size/ mm	mass/g (% of total mass in brackets)			
	till	DF[1] (ds) [2]	DF (us) [3]	fluvial
1.0-1.5	23.68(13.7)	4.03(6.1)	14.45(12.4)	18.79(28.2)
0.5-1.0	13.41(7.7)	3.67(5.6)	9.29 (8.0)	8.39(12.6)
0.2-0.5	53.74(31.0)	20.80(31.5)	50.48(43.3)	31.68(47.5)
<0.2	82.40(47.6)	37.46(56.8)	42.35(36.3)	7.85(11.8
Total	173.23(100)	65.96(100)	116.57(100)	66.71(100.1)

[1] = Debris Flow   [2] = downstream   [3] = upstream

TABLE 6.2
Clast size distribution: matrix of till, debris flows and fluvial sediment
from the large fan, upper Gleann Lichd
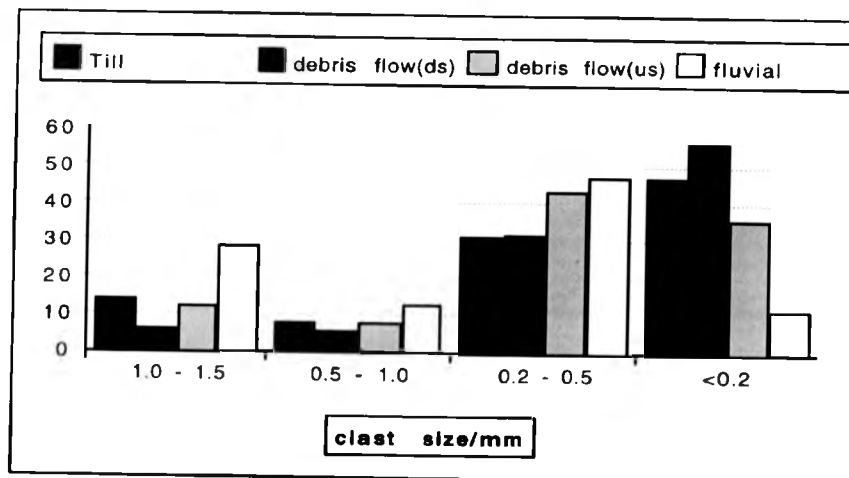
FIGURE 6.7
Comparison of the matrices of till, fluvial and large fan debris flows, Upper Gleann Lichd
Based on Table 6.2

6.3.1.2 The Large Fan

Debris on the surface of the large fan contained surface blocks up to 500 mm in diameter. Smaller boulders and cobbles were irregularly distributed, causing problems for representative sampling. As a result, only matrix of up to 1.5mm diameter was included in the analysis. Table 6.2 above shows sieved fraction data for two debris flows, and river sediment as well as till. The debris flows were at the extreme upstream margin (Section GL3), and downstream margin respectively, of the large fan apron. For comparison, a sample was taken from what appeared to be an unsorted, matrix-supported flood deposit in the river bank, just beyond the downstream margin of the fan. There was an element of uncertainty about the separation of fan and fluvial sediment at this point since they were superficially similar, and the fan toe ramp at this point was indistinct.

With reference to Table 6.2 and Figure 6.7, the debris flows cannot be wholly derived from till since they contain much coarser (>500 mm) fragments than either the till or fluvial sediment (< 300 mm). But the matrix of the downstream debris flow resembles the till matrix, showing a similar increase in the percentage of <0.2 mm as well as 0.2 - 0.5 mm particles. The upstream debris has features absent from the downstream sample: it contains a proportion of weathered clasts in the >1.5 mm fraction as well as a small amount of macroscopic organic matter. The downstream flow consists entirely of angular to very angular fresh rock and mineral fragments. It lacks both weathered particles and OM. It also contains a higher percentage of finer particles. However fluvial sediment is differentiated from both debris flows by a very low percentage of fines (<0.2 mm) which have been winnowed out, and consequently, a coarse skew.

6.3.1.3 The Small Fan

Clast size distribution in the fine-grained horizons associated with Samples GL/1 and GL/2, together with one of the overlying coarse debris flows are tabulated overleaf in Table 6.3 and Figure 6.8. The largest clasts of over 50 mm in the coarse horizon were infrequent and well dispersed and were not included in the analysis due to the difficulty of obtaining and transporting a sample large enough to be representative. Table 6.3 therefore over-emphasises the limited similarity between coarse and fine layers apparent from this analysis.

size/mm	mass/g (% of total mass in brackets)		
	lower fine layer	upper fine layer	coarse flow
30-50	nil	nil	95.73 (23.5)
10-30	nil	nil	87.43 (21.4)
1.5-10	nil	nil	119.39 (29.2)
1.0-1.5	2.28 (2.1)	5.48 (7.8)	27.10 (6.7)
0.5-1.0	2.48 (2.2)	6.04 (8.6)	11.70 (2.8)
0.2-0.5	29.20 (26.3)	17.63 (25.1)	31.66 (7.8)
<0.2	74.56 (67.2)	36.63 (52.0)	35.20 (8.6)
Total	110.89 (99.9)	70.37 (100)	408.21 (100)

TABLE 6.3
Clast size distribution fine layers and coarse flow matrix:
small fan, Upper Gleann Lichd
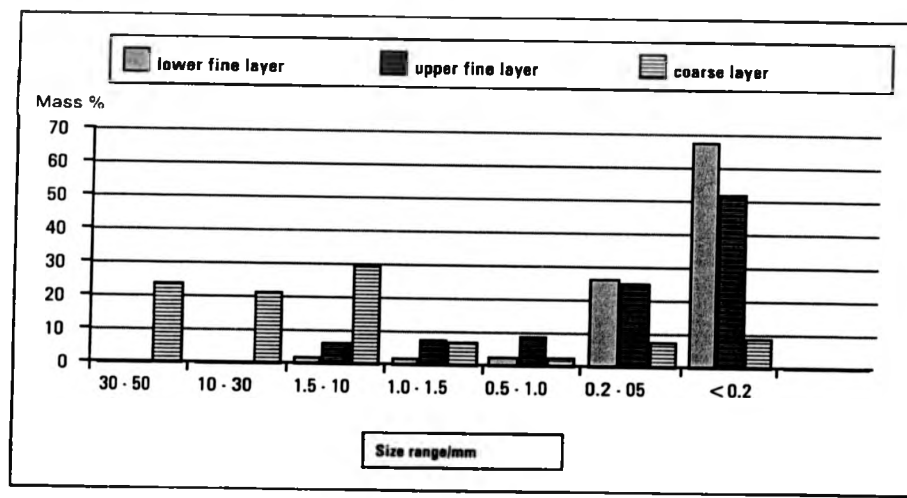
FIGURE 6.8
Debris flow clast size analysis, small fan, upper Gleann Lichd
Based on Table 6.3

Both fine layers are clearly differentiated from the coarse layer by degree of sorting and organic content, as well as by the range of clast diameters. Fine horizon sediment is angular to very angular, with unweathered mica and feldspar well represented. It also includes rootlets and numerous granules of uniform size in the >0.5 mm size range which were not found in the coarse horizon. These granules were resistant to vigorous sieving, but on gentle crushing proved to be soil aggregates composed mainly of silt and clay which were probably bound together by organic matter associated with soil development (Kemp 1985).

The two fine layers have similar patterns of grain size distribution although the upper one is slightly coarser, with 67% rather than 52% of grains in the <0.2 mm fraction. The finer layers entirely lack particles >1.5 mm in diameter, with 85% of the mass of the sample in the <0.5 mm range. In the coarse layer only 16% of grains are smaller than 0.5 mm and 29% of clasts are larger than 1.5 mm. Some overlap in size ranges between coarse and fine layers therefore occurs only in the 0.5 - 1.5mm range. Fig 6.8 above expresses these differences as a bar chart.

The weathering and sphericity of a proportion of 5- 10% of the >10 mm particles in the coarse layer of the small fan contrasts with that of finer fractions (Table 6.4 overleaf). The latter, as well as the bulk of the >10mm fraction, are characterised by a high degree of angularity and fresh mineral surfaces. However, some sub-rounded clasts show a degree of chemical alteration.

The angularity of fragile mineral grains such as biotite and muscovite, together with the low degree of chemical weathering, and high percentage of small rock particles (dominant even at fine sand grade), indicates little water transport and leaching. These are therefore mineralogically and texturally immature sediments transported in largely impact-free debris flows. Several features of the two fine layers confirm them as eroded mineral soils. These are:

- the restricted size range of mineral grains
- the presence of rootlets
- aggregates of organically bound silt and clay
- iron/manganese staining

		Till	Fluvial	Debris flows	Soil
Sphericity	Fine fraction	angular/ Sub-angular	sub-rounded + angular	angular, v. angular	angular, v. angular
	Coarse fraction	v. angular	sub-rounded + angular	angular, v. angular, (sub-rounded)	
Alteration	Fine fraction	fresh	fresh	fresh	fresh
	Coarse fraction	fresh	fresh +altered	fresh; (altered)	
Q:R*	Fine fraction	>1 in >0.2mm fraction	>1 in >0.2mm fraction	>1 in<0.2mm fraction	>1 in<0.2mm fraction
	Coarse fraction	<1	<1	<1	
OM**	Fine fraction	none	none	none	yes
	Coarse fraction	none	none	(in upstream flow, large fan only)	yes
Fe/Mn***	Fine fraction	none	none	none	yes
	Coarse fraction	n.a.	n.a.	n.a.	n.a.

*Q:R = ratio of quartz particles to rock fragments

**OM = macroscopic organic matter

***Fe/Mn = Iron (and manganese?) staining

( ) = minor component

TABLE 6.4

SEDIMENT CHARACTERISTICS, UPPER GLEANN LICHD

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The coarser layer is similarly mineralogically and texturally immature, but lacks all four of the indicators of soils listed above. It has a minor component of weathered clasts, no organic matter, and an unsorted range of clast sizes in the 1 - 250 mm size range. The slight rounding and weathering of some >10 mm clasts may, as on Creagan a' Chaorainn, be accounted for by incorporation in flows of disaggregated bedrock of small quantities of pre-existing, chemically altered surface particles. The extreme angularity of the mineral component of the soils open the question of whether they were derived from granular disaggregation of bedrock rather than washed from till, although the local variability of till has not been investigated.

6.3.1.4 Discussion of sediment analysis

The largest particles excepted, debris from the upstream margin of the large fan closely resembles the coarse sediment in the small fan, but differs from the downstream debris. These differences can be accounted for in more than one way. One possibility is that the downstream debris flow entrained some till. However, the characteristic subangularity of finer till particles was absent, so this explanation is poorly founded. An explanation based on debris generation, rather than to debris transport, is therefore preferred (albeit tentatively, since till variability has not been excluded). Although gully lithology was not mapped in detail, there are alternations of psammites and semi-pelites in the adjacent slopes (Section 6.1). The higher percentage of finer particles in the downstream debris flow may be the consequence of more pelitic rock producing more fines during granular disintegration. If this is correct, both upstream and downstream debris samples are largely the product of bedrock weathering rather than till failure.

In conclusion, sediment analysis tends to confirm that till and rock debris are genetically distinct types of sediment. Fluvial sediment is closely related to till, while slope podzols formed from fine-grained colluvium probably originated from granular disintegration of bedrock.

6.3.2   Stratigraphy of the small fan

The dated record in upper Gleann Lichd, as on Creagan a' Chaorainn, starts at 7525 - 7390 cal BP (Sample GL1/1, Figs. 6.6, 6.9). Charcoal of this age is dispersed in a small lens of sediment in angular, unsorted silt and sand, about 20cm above the base of the fan. Basal sediment was deposited on river cobbles exceeding 1m in thickness, which occupied a channel hugging the SW valley wall. The channel floor was therefore over-

run by the fan at around 7.5 cal ka BP. The fan toe has been removed by fluvial erosion, so the extent to which it over-ran the floodplain at different times is unknown.

The fan is built from two main sediment types which alternate up the section, and were analysed in Section 6.3.1. The first is coarse, matrix-supported, and very poorly sorted debris. Maximum clast diameter is generally in the range 80-150 mm, rarely 250 mm. The second type is relatively fine (coarse sand and finer), heavily iron-stained, and is found as shallow channel fills and occasional water-deposited small lenses as well as thin sheets.

A number of markers distinguish fan from fluvial sediment in the field and define the vertical fluvial/fan boundary. Diagnostic features are that:

- Coarse fan layers show some inverse grading not found in river sediment, but characteristic of debris flows.
- Long axes of elongate particles are aligned with the source gully and slope, and normal to river flow.
- Flow directions in the shallow channel and the charcoal lenses conform with the direction of fan, not fluvial, deposition.
- The extreme textural and mineralogical immaturity of the finer sediment is inconsistent with fluvial transport, which results in some winnowing of fines.

The next dated movement of slope sediment occurred at 4080 - 3890 cal BP, some three and a half millennia later than Sample G1/2. The long time gap was unexpected since the entire time interval was represented by only 0.65m of debris in two thin flows. Either there are undetected erosion surfaces between the two charcoal layers, or it represents a condensed sequence resulting from restricted sediment supply. There is little field evidence for erosion, although the sequence above GL1/2 is thicker than that below, and the shallow channel below sample GL1/2 indicates that small scale surface flow removed some sediment, but only to a depth of a few centimetres. Apart from water erosion, two other possible agents of fan surface erosion are the debris flows themselves, and river flooding. The lack of thick flows, their roughly planar flow boundaries, and preservation of a shallow channel and delicate charcoal lenses suggest that flows of debris did not incise the fan surface to any depth. Had flood water removed fan debris, the small lens of fragile charcoal in a low pile of unconsolidated sediment just above river level is unlikely to have survived. Moreover, no structures

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consistent with river flow are to be found within the fan. The thin sequence deposited between about 7.5 and 4.0 cal ka BP is therefore interpreted as reflecting contemporary sediment starvation, hence slope stability.

Above the second charcoal (GL1/2), that is, after about 4.0 cal ka BP, fan growth was more rapid as a result of slightly thicker flows. The coarse:fine (rock debris:soil) ratio above this horizon is 2.5 times greater, and the sequence terminates with two successive coarse flows.

6.3.3   Stratigraphy of the large fan

Organic material found beneath the surface of the large fan has allowed reconstruction of a history starting at about 4.0 cal ka BP and extending to some seven hundred years ago. The available dates from the two fans are therefore sequential.

Three logged sections GL2 - 5 (Figs. 6.10 - 6.13) sample a 30m cross-section of valley floor stratigraphy beneath, and adjacent to the upstream margin of the fan surface. As in the small fan, slope debris overlies palaeochannel deposits, but the relative volume is reversed, since the bulk of each of Sections GL2 - 4 is channel fill, with thin, marginal fan debris above. Gravels with clasts up to 150 mm in diameter thicken away from the valley wall. Gravel lenses, twigs, clast imbrication and pinch-outs trending towards the left and right margins of the deposits are aligned with the flow of the modern river, not with the fan (Fig. 6.10). Their strike therefore projects their continuation downstream, directly under the fan, which over-runs them at right angles from left to right.

Four main facies are represented. Basal deposits are poorly sorted, matrix-supported river gravels (flood deposits). Above this, is a vertically and laterally variable succession of peats and organic muds, interrupted by influxes of mineral sediment - palaeochannel fill - whose lateral variability restricts detailed correlation. This is overlain by the third facies, relatively thick peat. At the top is coarse fan debris overlain by 5-10 cm of immature, sandy soil.

Each of the facies is bracketed by radiocarbon dates as listed in Table 6.5 overleaf.

Facies	Description	Age/cal BP	GL Samples
1	River gravels	>4515-4415; >4420-4260; >4570-4420	2/1; 3/1; 4/1
2a	Palaeochannel fill	between 4515-4415 and 3830-3645	2/1 - 2/2
	(organic + mineral)	between 4420-4260 and ? c. 2000	3/1-undated peat base
		between 4570-4420 and 2710-2350	4/1 - 4/2
2b	Upper palaeo-channel peat	between 3830-3645 and 785-675 between ? 2000 and 1260-1080	2/2 - 2/3 undated - 3/3
		between 2710-2350 and 1230-1175	4/2 - 4/3
3	Fan debris	after 785-675; 1260-1080; 1230-1175	2/3; 3/3; 4/3
4	Modern soil	c. 10 cm silt + sand post-dates fan debris	

TABLE 6. 5

Calibrated radiocarbon ages of bounding surfaces below the surface of the large fan, Upper Gleann Lichd

The river gravels found at the base of the palaeochannel were deposited in a high energy fluvial environment. The organic materials, which began to accumulate on top of them between about 4.5 and 4.2 cal ka BP, mark a replacement of the fluvial system by subaerial, floodplain conditions, since the thin basal peats in Sections GL2 and GL3 can only have formed in the absence of flowing water and erosive floods. The 70 cm-thick organic deposit in GL4 at the same stratigraphic level appears to have been deposited in a single episode. The event is interpreted as a sediment-laden flow which entrained a section of peaty river bank as a randomly oriented slurry of peat, wood fragments, irregular lumps of peat, and peaty sand. The more central position of Section GL4 in the palaeochannel is reflected in thicker strata than are found in GL3 and GL2. Although the accumulation rate of the lowest peats is not known, their thickness of 10-20 cm probably marks a period of the order of hundreds of years rather than decades. During this time, flooding was absent, and vegetation was able to colonise the former palaeochannel. Abundant wood in these basal peat and muds makes it clear that, unlike today, trees and shrubs clothed the valley walls. Only birch was unambiguously identified from well-preserved bark, but downstream, frequent pine remains (a species absent from the valley today) can be seen buried 1 metre or more beneath floodplain

peats. Four hundred metres upstream, the rock gorge through which the river descends to the trunk valley contains a relict fragment of the formerly rich flora, including native tree species.

A further episode of bank erosion followed the period of basal peat formation. It is recorded in the homogenised peat, sand and pebble layers found in Sections GL2 and 3. Nearer the centre of the channel in GL4, the corresponding(?) deposit is fluvial gravel. If these strata can correctly be correlated, they record the passage of a flood, which occurred after a period of several hundred years of low water flow and/or channel migration. However, the possibility cannot be excluded that the gravel in GL4 was deposited either earlier or later than the bank erosion deposits, in a flood which was not large enough to affect more marginal areas of the channel. The frequency of floods is difficult to determine precisely on the basis of this evidence, although they appear to have been rare. Above these horizons, correlation between Sections GL2 and GL3 is more difficult, perhaps because of the combined effects of erosion and deposition. However, both Sections have a second thin peat which indicates a second episode of sustained vegetation growth and a continuing rise of the palaeochannel floor before a further incursion of flood gravels.

In summary, Sections GL2 and GL3 close to the valley wall contain thin peats, twice interrupted by flood deposits, before the thicker, sediment-poor peat, found in all three Sections above this stratigraphic level, formed in the absence of large turbid floods. Overall, the stratigraphy records replacement of fluvial processes in an active river channel by subaerial processes. Peat accumulation was interrupted on two (or possibly more) occasions by large floods in the three millennia between 4.5 - 4.2 cal ka BP and about 1.2 cal ka BP (Table 6.5). But floods started only after a period of uninterrupted peat growth, probably at about 3.3 cal ka BP (GL 3/2, Fig. 6.12), and ceased to reach the palaeochannel after about 2.5 cal ka BP (Sample GL 4/2). Thereafter, peat growth resumed for more than a millennium, until fan debris reached the valley floor at about 1.1 - 1.2 cal ka BP (Samples GL3/3 and 4/3). The uppermost peat surface in Section GL2 (sample GL 2/3) at some 700 cal BP, is several hundred years younger than that in Sections GL3 and GL4. Surface contamination of this shallow boundary by younger carbon is a possibility. Alternatively, at the very margin of the fan, flows of different ages are possible. The evidence is ambiguous, but large flows of debris clearly reached the floodplain at the upstream fan margin for the first time at 1.2 - 1.1 cal ka BP and

possibly even later. Substantial parts of the massive lower fan are therefore very late Holocene in age.

6.3.3.1 Peat accumulation rates

As on Creagan a' Chaorainn, only a selection of organic horizons could be dated. Two samples (GL3/1 and 3/2, Fig. 6.12) allow maximum and minimum average accumulation rates for the thin (23 cm) peat at the base of Section GL3 to be calculated. At 95% confidence levels, and averaged over a period of about 1000 years (between 4420 - 4260 and 3380 - 3270 cal BP), the range is 2.6 - 2.0 cm/100 yrs. For comparison, the growth rate of the uppermost peat in Section GL4 (Fig. 6.13) can be calculated between samples GL4/2 and 4/3. Over 1535 (max.) to 1120 (min.) years, 60 cm of peat formed. The average accumulation rate over a period of between 1.5 and 1.1 calendar years was therefore 5.4 - 3.9 cm/100yrs. Although the uncertainty appears larger than in Section GL3, proportionately, it is not dissimilar: ± 14% as opposed to ± 12.5%. Peat filling the upper part of this section of the palaeochannel in Section GL4 therefore grew, on average, on a timescale of $> 10^3$ yrs, approximately twice as fast as that in Section GL3. Since the two sections are only 6m apart, lateral variability, as well as the vertical variability discussed in Chapters 1 and 5, appears to be large. One possibility is that conditions for peat growth were enhanced after the 4th millennium BP, and that the increase is real. Another is that local variability is high. Because of the size of the error bars and the unknown variability within the peat column, extrapolation of these rates to timescales of less than 10^3 years, or to other peats in the logged sections, potentially introduces large errors. In this type of setting, the usefulness of peat accumulation rates except in very high resolution studies is therefore questionable.

6.3.3.2 Stratigraphy beneath the undissected surface of the large fan

Sediment and peat lenses, clast orientation and occasional imbrication in Sections GL2, 3 and 4 unambiguously show flow down valley at right angles to the long axis of the large fan (Fig. 6.10). Two pits were therefore dug on the fan surface (Figs. 6.6, 6.14) in an attempt to intersect palaeochannel organic horizons under the body of the fan, and to assess the thickness of the lower fan apron.

The fan surface consists of a thickly grassed peaty soil, littered with boulders. Pit 2 was dug 3m in from the river bank sections to a depth of 0.5m in a shallow, longitudinally-impersistent depression. It revealed nothing more than crude water sorting of reversely

graded fan debris, but confirmed the broad shallow depressions shown in Figure 6.6 as drainage channels associated with debris deposition. Pit 1 was 8m in from the bank sections, and in the centre of the inferred extension of the palaeochannel. It intercepted two stacked debris flows, the upper of which was >1m thick. No organic horizons or sign of the base of a debris flow were found even at 1.45m below the surface, some 45 - 50 cm below the projected level of the peat surface and 20 cm below the water table where further excavation was impracticable.

6.3.4   Floodplain stratigraphy

Section GL5 (Fig. 6.12) is beyond the fan toe. Peat 0.5m thick, whose upper surface is dated at 545-510 cal BP, rests on river gravels. However, wood immediately below its *base* has a radiocarbon age indistinguishable from that of mature contemporary trees (300 - 0 cal BP). The reason for the age reversal was sampling error.

Small diameter wood, rather than basal peat, was chosen for dating in the interests of precision. The sample was taken from an apparent flood deposit of organic detritus lying horizontally between the base of the peat and the river gravel. However *Alnus glutinosa* (alder) on the riverbank downstream, was found to have a tangled root system which often turns to run horizontally on meeting buried gravels. The sample age makes it probable that the dated root belongs to now vanished alder less than 300 years old. Based on the very wide range of rates of peat growth calculated above (5.4 - 2.0 cm/100 yrs, - Section 6.3.3.1), the real age of the basal peat, and therefore of the timing of initial peat growth on this section of floodplain, which continued with only very minor incursions of sediment until after 545 cal BP, is probably between about 1.5 and 3.0 cal ka BP. Unfortunately, this date could not be more closely constrained.

After 545 cal BP, peat gave way abruptly to repeated layers of sand grade overbank deposits 1-2 cm thick, each probably representing a separate flood, interlayered with approximately centimetre-thick peaty stringers representing perhaps a few years to a few decades of floodplain vegetation. The overbank sequence is in turn overlain by a thickly grassed peaty soil recording a cessation of the regular overbank deposition which permitted a continuous soil and sward to develop up to the present day.

6.3.5   Sediment micromorphology

Four thin sections were made for micromorphological investigation. Photomicrographs are reproduced in Figures 6.15 to 6.18. Additional sampling was planned, but was

abandoned because poor sorting of even the finer sediment introduced coarse particles unsuitable for inclusion in large thin sections. Each of the four samples is matched to a radiocarbon dated sample whose location is starred on the corresponding graphic log. One purpose of the exercise was to examine peat-debris flow boundaries with a view to detecting possible hiatuses in the stratigraphy. A second aim was to establish whether microscopic features of mineral and peat were consistent with interpretation based on morphology and stratigraphy.

Sediment containing charcoal sample GL1/1 (*Fig. 6.9)

The thin section samples the sediment containing the charcoal lens dated at about 7.5 cal ka BP. It is texturally and mineralogically immature and matches bedrock mineralogy (Tables 6.3, 6.4). It is composed mainly of fine to medium grained poorly sorted angular quartz with fresh muscovite and biotite laths with no preferred orientation, and with accessory chlorite and feldspar (Fig. 6.15a). The matrix is light brown amorphous organic matter (AOM) (Fig. 6.15b). Occasional aggregates of mineral grains are set in distinct masses of AOM, sometimes with plant residues. These are soil particles (Fig. 6.15c).

Scattered carbonised plant fragments have three forms (Fig. 6.15b); cross sections of stems showing well preserved tissues; partially decomposed fragments with residual cell structures; and decomposed fragments lacking any clear structures. A single, long section, about 5 mm in length, lies in a sub-vertical plane partially filling an elongate vug. The groundmass contains vugs up to two or three millimetres in diameter. The vertical plant fragment in this view illustrates the random arrangement of particles.

Some vugs have black coatings of what are probably Fe/Mn precipitates (Fig. 6.15d) or rarely, coatings of light brown AOM. The coatings are less common on mineral grains such as quartz, and never appear on mica (Fig. 6.15e). Coatings vary from complete, to partially detached and partially formed. Partial coatings exhibit no systematic way up. Encircling coatings of AOM probably formed in the same way as encircling silt cappings (Elliot 1996). The latter are created by grain rotation when excess pore water pressure causes small-scale viscous flow. This may be associated with freeze-thaw conditions in winter and early spring. The oceanicity and variability of the Scottish mountain climate (Chapter 3) produces 'stop-start' winter conditions. It may therefore create more frequent near surface freeze-thaw episodes - and hence more opportunities for grain rotation in soils - than in the alpine conditions in which these effects have been

studied (Caseldine & Matthews 1985, Matthews 1984, 1993). A low incidence of complete grain coatings suggests minor slope instability sufficiently frequent in some years to severely hamper the lengthy process of soil profile development. This in turn suggests the possibility of two causes of soil immaturity: insufficient length of time for profile development, or frequent small-scale instability.

The mineralogy, texture and organic content of this deposit are those of an immature leached mineral soil eroded by run-off and transported as a debris flow as shown by:

- co-deposition of mineral and organic materials
- lack of sorting and rounding of mineral particles
- vertical orientation of elongate plant tissue and vugs
- random orientation of unweathered mica laths
- grain coatings developed as a result of pre-depositional leaching
- no consistent way up of grain coatings and vug lining
- the presence of soil aggregates.

Plant remains do not include woody tissue, suggesting that the source of the flow was a thinly vegetated slope largely lacking a shrub or tree layer, but supporting a thin developing podzol. The freshness of the micas suggests rapid erosion and deposition. However intervals of stability, although unquantifiable, must have been long enough for processes of soil leaching and re-precipitation of iron complexes to have got under way.

The micromorphology of this sample confirms interpretation based on stratigraphy and field observations. The earliest recorded slope invasion of the riverbed in Upper Gleann Lichd, was by an eroded soil at the base of the small fan. Slope denudation was infrequent, occurring at intervals long enough for vegetation to become established and processes leading to podzolisation to occur. The charcoal did not retain any tissue structure and disintegrated into a fine paste under gentle finger pressure, which is perhaps confirmation of a lignin-poor source. The soil developed in equilibrium with outcrop-enhanced run-off, so its erosion may be linked to the fact that its stability thresholds were lowered as a result of destruction by burning of a shallow root mat.

Sediment immediately below charcoal sample GL1/2 (Fig. 6.9)*

The sample is topped by the charcoal wash dated at about 4.0 cal ka BP (GL1/2). It is separated from GL1/1 by 65 cm of debris incised by a shallow (up to 20 cm) scour filled with fines. Mineralogically, the two strata are similar. Microscopically, this

sediment differs texturally in having several sub-horizontal layers and streaks of silt embedded in light brown AOM. Plant material is scarce (Fig. 6.16a), but rounded soil aggregates are present (Fig. 6.16b). The percentage of mica of both sand and silt size is higher than in the sediment containing sample GL1/1. Poor sorting and random orientation of mica laths are characteristic. Dark areas in the interstices between mineral grains are AOM.

Similarities between the two sediments are greater than their differences, although there is evidence in the sediment adjacent to sample GL1/2 of minor water sorting. In addition, the slightly coarser texture and lesser content of organic matter and grain coatings suggest a less mature soil. Charcoal is again associated with soil erosion.

Peat /debris flow boundary Sample GL4/3 (*Fig. 6.12)

This thin section samples the boundary between the topmost peat in the palaeochannel fill below the large fan and the base of the overlying reversely graded debris flow. Angular, poorly sorted quartz, mica and accessory mafic minerals are dispersed throughout the mid-brown AOM of the peat in variable (15-30%) concentrations for the whole 60 mm length of the thin section (Figs. 6.17a & b). In one 15-20 mm thick layer with indistinct boundaries, the mineral content rises to 60-80%. The AOM here is darker in colour than the characteristic light brown soil AOM seen in samples 1 and 2. At the base of the thin section (Fig. 6.17b), sand-sized rock fragments are found adjacent to variably preserved woody tissues, some >5 mm in length, in a matrix of peat (dark coloured AOM). Some plant remains are sub-vertical, though most lie horizontally.

The peat contains a scattering of black carbonised flakes of plant material. In some of these, residual cell wall structures form a black lacey pattern. Non-woody stem cross sections and (?)seed cases (Fig. 6.17a) are also present. Apart from one large (c.10 mm) vug which appears to pseudomorph a fragment of wood which has been slow to decompose (Fig. 6.17c), vugs are small and rare. Black Fe/Mn grain and vug coatings are absent. Mica laths are disposed in a roughly horizontal plane in some poorly defined millimetre scale zones, but are randomly oriented elsewhere.

A distinguishing feature of this sample is the transitional boundary between the organic and mineral units. There are two possible explanations: either frictional mixing as the debris flow overwhelmed the peat, or inwashing of silt and coarse sand as the peat

accumulated. The process of peat formation with loss of a high percentage of the original vegetation through oxidation and bacterial decomposition allied to compaction, must mean that very minor additions of mineral grains become volumetrically exaggerated during peat formation (Chapter 5). For instance, if 15% of plant matter survives decomposition, a 5% scattering of sand and silt, whose volume, unlike organic matter, is unchanged by the transformation, would be exaggerated to about 30% by volume of the resultant peat. This matches the percentage noted above which permeates the whole thin section.

There is evidence to support both hypotheses. Frictional mixing is suggested by the intimate mix of mineral clasts with peat, the randomly oriented mica flakes, the sub-vertical woody tissue, the unusually small number of vugs, and indistinct layering. Incorporation of a steady sediment rain as peat accumulated (with a thin mineral flush at one point) - hence maintenance of peat surface integrity beneath the debris flow - is supported by the presence of mineral sediment throughout the peat, the lack of differentiation into distinct layers, the indistinct zones of roughly horizontal mica flakes sedimented from suspension, and an absence of features attributable to shearing or sudden compaction. In summary, the evidence from this sample is ambiguous, but retention of some depositional textures suggests that there was no major peat surface disturbance below the margin of the debris flow.

Peat/overbank sediment boundary Sample GL5/2 (Fig. 6.12)*

This sample spans the peat/overbank sequence boundary on the floodplain just beyond the toe of the large fan. It was selected to test the hypothesis that floodplain sediment younger than 545 cal BP was deposited from suspension during moderate overbank floods. If confirmed, the nature of the boundary forms a useful benchmark against which to gauge the extent of friction and disturbance between peat and debris flows in adjacent Sections.

The boundary between the peat and overlying mineral deposit is distinct and near-horizontal (Fig. 6.18). The peat contains poorly preserved plant fragments in a matrix of AOM. The mineral layer consists of moderately well sorted angular to subangular quartz and mica with sub-horizontal orientation of muscovite laths. It also incorporates occasional millimetre-scale masses of peaty AOM with sharp boundaries. The boundary between the peat and the mineral layer is non-erosional and the surface undulation apparent in the photomicrograph is on a scale of c. 750μ. Crude alignment of elongate

grains in the mineral layer indicate suspension deposits from shallow floods when water velocity abruptly declined in contact with floodplain vegetation. The moderate sorting is characteristic of upland river transport. Small peaty rip-up clasts originate from upstream banks. The texture and sorting of the mineral sediment together with the non-erosional boundary on the peat surface confirms field interpretation of passive deposition from a waning flow as the river overtopped its banks.

In conclusion, there is a degree of ambiguity about some of the evidence, but microscopic features of the peat/debris flow boundary in the large fan provide no reason for believing that the debris flow dates based on upper peat surface ages are significantly in error due to peat erosion. Some small-scale surface disturbance is apparent, in contrast to suspension deposits on floodplain peats. Low friction, reversely graded debris flows at the fan margin contrast markedly with evidence for erosion of the substrate only eight metres away (Fig. 6.14), where thick, defluidised debris appears to have displaced 1-2m of channel fill. In the small fan, sediment micromorphology has confirmed that finer grained strata are eroded soils, and that soil erosion alternated with transport and deposition of rock debris.

6.4 DISCUSSION

Three main sources of mineral sediment of Holocene age have been distinguished in upper Gleann Lichd. These are: till, slope podzols, and eroded bedrock from gullies. Small volumes of relatively fine grained debris were deposited on the former floodplain when podzols were transported as dense fluidised slurries which preserved some delicate soil features and mineral grains. Large volume, reversely graded, coarse flows were sourced mainly in bedrock, transported via rock-cut, joint-controlled gullies, and had greater momentum and longer travel paths. One flow included a till component, probably eroded from gully walls together with minor amounts of slope surface material.

Organic deposits fall into four categories, including: small amounts of charcoal, which may or may not be anthropogenic; well humified peat with numerous remains of woody vegetation; and chaotic, homogenised bank erosion deposits. Debris flows composed of slope podzols bridge the mineral/organic groupings.

Floodplain and slope history are considered together in the following sections along with triggers for landscape evolution and the robustness of the chronostratigraphy.

6.4.1 Slope development

Slope evolution can be reconstructed by examining the processes and events which formed the fans and caused till failure.

6.4.1.1 The small fan

As a landscape feature, the small fan is not striking, but it preserves evidence of some key events. The earliest known Holocene event occurred at about 7.5 cal ka BP, when sediment with a charcoal lens over-ran the former high energy river cobbles. The conservation of a few decimetres of easily erodable sediment close to the early fan toe suggests that the river had, by then, moved away from the base of the slope on which the fan was deposited. Given the narrowowness of the floodplain, and the presence of a flow-accelerating rock gorge upstream (Section 6.3, Fig. 6.9), it seems likely that the river had, by this date, occupied (?incised or re-incised) its present, rock-walled channel close to the opposing valley wall.

Two separate processes appear to account for aggradation of the small fan: erosion of immature sandy soils developed in drift deposits, and mobilisation of coarse debris whose characteristics suggest derivation from bedrock. There is no evidence that shallow erosion of soil organic layers was the precursor of, or synchronous with, deeper, more massive flows. Deposits on the fan surface were superimposed during separate events, and are separated by unknown time gaps. Since soil forms mainly in intergully areas where run-off is normally insufficient to disturb the surface, its erosion after a period of development long enough to initiate podzolisation suggests unusually intense rainfall or snowmelt. Flows which entrained soil, show some evidence of water sorting, whereas coarser debris was deposited in more viscous flows. This confirms a greater role for surface runoff when soil was eroded.

The association of charcoal exclusively with the finer layers is a result of depositional processes. The vertical sequence of alternating fine/coarse couplets in Section GL1 could have more than one cause. One possibility is superposition of debris of different origins. Alternatively, fine/coarse couplets could have been formed by superposition coarse flows, and small terminal flows during water-dominated transport. The latter possibility can be excluded because both types of horizon exhibit very poor sorting. In addition, all the fine material is stained by iron (and ?manganese) while the coarser

deposits are not. The differences between the layers are therefore attributable to source and timing, not transport and deposition. At the top of the sequence, two successive coarse layers testify to the predominance of rock debris erosion towards the end of the period of fan aggradation. Coarse material is exclusively inorganic, while finer horizons have the characteristics of soils, which have been transported as saturated debris. The fires which produced the charcoal preserved in two of the fine layers, may have been implicated in destabilisation of slope soils.

The fan stream has cut down through the entire depth of the fan, and incision must have occurred when the water:sediment ratio was high. A high ratio of bare rock to soil and sediment such as is found above the fan, inhibits rapid infiltration and enhances surface run-off (Chapter 2). In the absence of mobile sediment, vigorous run-off would routinely have been channelled to the fan apex by the two feeder gullies. Incision seems likely, therefore, to have been established at an early stage on the low fan surface between 7.5 and 4.0 cal ka BP when the condensed stratigraphy suggests sediment starvation, and to have continued between each fan-building event.

Given the geomorphological setting, the failure of the river to recycle the low, unconsolidated fan deposits over a period of about three and a half thousand years, is worth noting. Both charcoal layers are found in easily eroded, sandy sediment less than 0.65m above the palaeoriver floor. At this point, the rock-walled valley floor is less than 100m across, and debris from modern floods reaches close to the base of the fan (personal observation). A few hundred metres upstream, a rock gorge accelerates the flow from the two head stream tributaries (Fig. 6.2). In consequence, a low fan palaeosurface would have been poorly protected from large or exceptional floods. Its preservation is therefore interpreted as evidence of limited overbank flooding during the interval of time represented. At the same time, the condensed sequence suggests that the slope above the fan remained generally stable.

Four conclusions can be drawn from available evidence: first, that before basal horizons were deposited, slope failures (if any) were reworked by the river; secondly, that throughout the period of time between samples GL1/1 and GL1/2 (7.5 - 4.0 cal ka BP), the river did not reoccupy its former channel; thirdly that the stability of the steep, debris-mantled slope was, during the next three millennia, little disturbed; and fourthly that slope mass movement and the bulk of fan aggradation at this site occurred mainly after 4.0 cal ka BP. The fan therefore preserves evidence of several events

whose possible triggers and threshold conditions are discussed in Section 6.4.3 below.

6.4.1.2 The large fan

The absence of organic horizons beneath the body of the fan, accompanied by a tripling of debris flow thicknesses compared to the fan margin 8m away, illustrates both the difficulty of finding preserved evidence to date fans of this type, and the fortuitous nature of preservation and exposure of the extreme marginal deposits in this instance. Without minor erosion of the fan margin at the point where some organic horizons were preserved, this fan would resemble many other coarse, undissected fans in Highland valleys which appear to be wholly mineral, relict (presumed early Holocene) landforms.

Figure 6.19 suggests a model for replacement of soft palaeochannel deposits by fan debris, based on information about the material and processes of deposition found in Pit 1. The forces involved in replacing palaeochannel fill with fan debris have not been modelled, but are suggested by both field observation and experimental conditions, where poorly sorted, water-saturated sediment is known to move as an unsteady surge (Reading 1978, Iverson 1997). Coarse surge-front debris, in which little or no pore pressure has been established, forms levées on the margins of flow paths as found on the upper and mid fan. At the base of the slope, coarse surface debris has travelled over 400m from the fan apex, and up to 1000m from the upper gully (Fig. 6.6). The same process could explain the nature of the debris emplaced on peat in the logged, marginal exposures (Sections GL2 - 4). It is both coarser, and much less compacted, than material sampled in Pit 1. The fan toe ramp - a bouldery deposit - is the third element of fan morphology that matches the proposed separation of rigid and fluidised elements of flows. The finer-grained, saturated debris, appears in the field to have been enveloped, marginally, and above, by non-fluidised coarse debris, as described in laboratory experiments by Iverson (1997). As momentum was lost, and the flow came to a halt, undergoing a phase transformation from fluid to near rigid, abrupt dewatering would have led to compaction.

These two physically and morphologically distinct elements of flows - fluidised and surge front deposits - seem likely to have had contrasting impacts on the substrate. As a large debris flow comes to rest, water is expelled, reducing the buoyancy of the deposit.

It follows that the force exerted by a given volume of water-saturated debris flow on a volume of water-saturated peat will increase as it comes to rest. The erosivity of a large, fluidised debris flow is therefore likely to be greatly increased as it decelerates.

If this model is valid, peats and muds in the abandoned channel in upper Gleann Lichd were vulnerable to erosion by an approximately 2m thick mass of rapidly decelerating debris (Fig. 6.19). At the margins of the flow, where non-fluidised debris accumulated, the peat would not have been significantly eroded. Field evidence in the form of both the absence of organics from Pit 1, and micromorphology of the upper peat surface beneath the marginal deposits (see 6.3.5 below) is consistent with this prediction. From the evidence of Pit 1, debris thickens rapidly away from the fan margin, while thin, non-fluidised debris overlies peat preserved at the very margin of the fan, but is absent a few metres into the body of the fan.

Most of the history of the large fan (before 1.2 cal ka BP) remains undated. However, the fan morphology and setting allow some inferences to be made about its earlier development. Changes in boundary conditions are likely to result in the formation of a 'nested' cone-fan sequence such as that created in Strathfarrar by changes in boundary conditions (Chapter 4, Section 4.4.1) (Ballantyne 2001, *pers. comm.*) In the absence of any such features here, the three zones (Fig. 6.6) may represent either three separate phases of growth, or a long history of episodic aggradation and progradation. The steep upper zone which terminates about 60m above the Holocene valley floor is consistent with its development as an early debris cone (as defined by Brazier 1987), which formed in periglacial conditions above a thinning valley glacier, during the retreat of Younger Dryas ice. Cones of this type are common in the northern Highlands (Chapter 4). There is no evidence that sediment in the middle fan was derived from upper fan erosion. It may therefore have formed (or begun to form) during rapid, early Holocene slope adjustment, although a younger age cannot be excluded. The low angle distal apron is unambiguously late Holocene in age with the earliest dated phase of aggradation occurring within the last 1200 years.

However, the dated deposits present an incomplete picture of lower fan aggradation. Debris flows at the upstream and downstream fan margins have probably tapped different sources of more psammitic and more pelitic bedrock respectively (6.3.1.2 above). Different source rocks, and travel paths to the valley floor suggest that these flows originated at different times. Complex fan surface morphology of debris flow

paths and lobes, and the erosional boundary between the superimposed flows in Pit 1, indicate that the dating has provided only a partial picture of aggradation.

Some indications of the nature, frequency and timing of additional events are available. Mapping shows three broad, shallow, temporary channels that die out in the middle section of the fan (Fig. 6.6). A small number of other channels extend from the mid fan to the low fan. The distal fan therefore appears to have been built by debris flowing down several separate channels to build lobes some of which, including the lower one in Pit 1, have been partially or wholly buried. The preservation of the channel which deposited debris above the dated peats suggests it formed late in the history of fan aggradation. It can be seen at the centre of a lobe in Figure 6.4. On the basis of the number of channels visible in Figure 6.6, the lower fan was built by about seven separate events which may or may not have been contemporaneous. This is a minimum number of flows, because each channel may have been the site of more than one event, and other channels may have been buried by later flows. Two (possibly three) flows were logged at the fan margin. The number of flows which built the lower fan could therefore be in the range of twelve to fifteen.

From Figure 6.6 the fan surface area is approximately 125,000 m^2. Assuming an average depth of 2m for each debris flow, or stack of debris from repeated flows, the volume of debris is about 250,000m^3. If the distal fan formed in the course of 12-15 events, each flow mobilised about 16 - 21 x 10^3 m^3 of sediment. At 2m thick, each flow, if it had a similar length/breadth ratio as the fan as a whole, would have had a surface area of at least 100 x 80m. This approximation is given some credence by empirical data: the lobe above the logged sections is of those dimensions and in excess of 1.5m thick. The morphology of the lower fan is therefore consistent with aggradation in a number of separate flows, which coalesced and/or over-ran each other. Most of these flows must have preceded the available date of 1.2 cal ka BP.

6.4.1.3 Fan incision

The small fan stream may have evolved in parallel with the fan itself, incising the surface when the run-off:sediment ratio was high, with fan incision and aggradation representing alternating system states. There is no evidence that incision occurred after completion of aggradation. Past changes in the run-off:sediment ratio throughout the fan's history are a more plausible trigger. Currently the fan surface is vegetated and

stable, and the fan stream has dissected it to base level. This suggests that the contemporary run-off:sediment ratio is high, and that debris is not being generated fast enough to form accumulations susceptible to destabilisation, but is instead being removed by frequent small scale attrition in balance with the energy of run-off.

If the controlling factor for large fan incision was also the run-off:sediment ratio operating throughout its history, the high permeability of coarse fan surface materials and the relative complexity of drainage may explain the absence of an incising stream. Large volume flows were enveloped by coarse, non-fluidised debris (Section 6.3.3 above), creating a rapidly draining surface. The large fan is built of stacked and overlapping flows, possibly resulting in additional free-draining flow paths below the surface. The large-volume, hydraulically complex, sink area, probably ensured that the run-off:sediment ratio was consistently lower than in the small fan. As a result, much more intense precipitation, and/or a much larger run-off zone would have been needed to initiate incision.

This hypothesis and its application to other coarse fans in the region can be considered with reference to two fans in nearby Glen Shiel (Fig. 6.1). A small fan with some buried organic horizons was found at NG954165 at the foot of Allt a' Mhuing (Fig. 6.20). Like the small fan in Gleann Lichd it originates in a joint controlled gully, has a low surface angle, and is incised by a stream. A large fan on the opposing valley wall at NG 960166 (the Allt Carnach fan, Fig. 6.21) is also incised, although its size and coarse surface are similar to the large fan in Gleann Lichd. However, its source, instead of being a long, joint-controlled gully, is a narrow, steep-sided, structurally controlled valley, with a surface area greater than $1km^2$, and with a correspondingly greater potential volume of run-off, relative to the Gleann Lichd gullies. Based on a diameter of about 500m, and a measured thickness of up to 2m near the base, the volume of debris in this fan is similar to that of the large fan in Gleann Lichd, while the source area is much larger. The run-off:sediment ratio is theoretically, and, as seen in the field, therefore biased in favour of incision. Co-evolutionary fan incision alternating with aggradation, provides an alternative to the anthropogenic fan incision proposed by Brazier *et al* (1988).

6.4.1.4 Till Failure

The till failure on the slope opposite the fans was found to be associated with a spring line 80m above the valley floor. Three vigorous flows arise from the bedrock at the same altitude as the base of the oversteepened 50° backwall of the slip, which contrasts with the adjacent slope angle of 34°. Massive (>1.0m) blocks of barely attached, fractured bedrock are exposed within the zone of failure. Other than in gullies, loose fractured bedrock on stabilised, till-covered slopes is unusual. Together with the contrast in slope angle, it indicates continuing slope disequilibrium. The unlithified, *in situ* 'breccia' appears to have been held in place by a covering of till, becoming susceptible to erosion only when the till was removed.

The freshness and continuing disequilibrium of the till failure opposite Gleann Lichd House suggests that it is fairly recent. In fact, it is not visible in aerial photographs from the 1950s, and may postdate 1978 (Ballantyne 2001 *pers. comm.*). The springs have been identified as marking the outlet of water draining the fracture zone of a major paraglacial rock-slope deformation (Holmes & Jarvis 1985), but the stability of the till and fractured bedrock overlying them until, apparently, the late 1970s, suggests a sudden flow enhancement which rainstorms in previous millennia failed to trigger - hence Late Holocene evolution of slope boundary conditions.

The nature of the till suggests an explanation for the large discrepancy in the size of clasts found in the river and those scattered on slopes. Upland floodplains in the Scottish Highlands characteristically contain glacifluvial deposits of till-derived boulder lag, and at first sight, the litter of large boulders in Gleann Lichd on intergully zones of slopes and on high terraces, also suggests the presence of bouldery till or moraines. However, only cobble size lag deposits are present in the riverbed in the main valley. Their maximum diameter of 200 mm matches the size of the coarsest till fraction (Section 6.3.1.1 above). Down valley, fill terraces of medium sand and silt resemble the products of winnowing of finer material in large energetic events which separated traction and suspension deposits carried in dense turbid flows (Postma 1986, Todd 1989). Table 6.2 shows the Gleann Lichd till as having a high percentage of fines, and the clay to cobble range of river particles matches the composition of the till.

The boulders which litter the surfaces of slopes and high terraces have a size range well in excess of the majority of till-derived clasts. They must, therefore originate in falls of unstable bedrock, some of which postdate glaciation. The obvious loci of this activity

are rock cut gullies, and an explanation is needed to account for substantial rock falls in intergully zones. The till failure opposite Gleann Licht House suggests a mechanism, although the timing remains uncertain. These rocks falls may belong to the period of postglacial relaxation, when till slippage exposed cracked and unstable bedrock. Alternatively, they may have occurred much later in the Holocene, as at the foot of the current failure.

Two possible causes of *in situ* brecciation are neoseismicity and periglacial frost shattering. Exposure of unlithified breccia by a till failure, together with the widespread occurrence of large, non-till-derived boulders below intergully zones, therefore opens the possibility that unstable cracked bedrock may underlie other apparently stable slopes. If this is the case, there may be a potential for further Holocene slope failure, whenever environmental conditions allow the integrity of the till cover to be breached.

6.4.2 Floodplain development

Dated floodplain evolution in upper Gleann Lichd covers a longer time-span than that for either of the fans: 7.5 - 0.3 cal ka BP, starting with abandonment of a high energy channel. This channel probably moved towards the opposite side of the narrow floodplain where it re(?)-excavated a rock-cut glaciofluvial (?) channel which is still visible today (Fig. 6.3). Deep incision of glaciofluvial deposits, and/or infrequent large floods, would account for the river's subsequent failure to reoccupy the abandoned channel except on very few occasions between 3.3 and 2.5 cal ka BP (the upper and lower surfaces respectively of the basal and uppermost peats beneath the large fan). A few hundred years before the soil failure at c. 4.0 cal ka BP which produced the second charcoal deposit in the small fan, the river channel below what is now the large fan was also abandoned to vegetation. No water-lain structures are found in the small fan sequence either before or after this time. Slope debris derived from the gully systems did not reach the floodplain, and neither were the abandoned channel peats invaded by silt and sand persistently washed from the adjacent slope.

Possibly at the same time, (although poorly defined by an estimated date between 3.0 and 1.5 cal ka BP, section 6.3.4 above), floodplain peats replaced winnowed river gravels in the middle of the valley floor. The earlier date is consistent with the above evidence for the river now being confined to a deep, rock-cut channel. Floodplain peat continued to accumulate undisturbed until shortly before 500 cal BP, when the sediment:water budget in upper Gleann Lichd underwent a transformation. Mineral-free

peat accumulation on the floodplain was replaced by sandy overbank deposits with thin peaty horizons. This marks the start of a period of moderate, sediment-laden floods repeated on a timescale of decades. The abrupt change in gradual floodplain aggradation from organic to largely mineral, indicates renewed erosion. An obvious source of the sand and silt is till. Although the extent of till failure upstream of the study site has not been investigated, further till failures were seen in the field and in air photos adjacent to headwater streams. This style of overbank deposition persisted until an unknown, recent date, when 10 cm of silty sand accumulated to form the matrix for the contemporary soil.

In summary, after early Holocene aggradation by mineral sediments, the floodplain record is one of incision, followed by a build up of organics, with only occasional incursions of mineral sediment. That gave way after about 2.5 cal ka BP to organic-only aggradation until 500 BP, when a supply of fine sediment again became available. The sediment:water balance may have been more important than changes in groundwater flow *per se* in controlling these changes. When erodable sediment was scarce, the channel (which today has a high depth:width ratio) incised its channel, and any overbank floods were sediment-poor, contributing little mineral content to floodplain peats. Energetic floods after 3.3, but before 2.5 cal ka BP recycled coarse sediment already in the fluvial system. Fresh sediment reached the river only when made available by probably non-anthropogenic till failures some 500 years ago.

6.4.3   Environmental triggers

This section considers evidence for the causes of observed slope and floodplain evolution in Upper Gleann Lichd.

The elevation, rockiness, and steepness of the recharge and runoff zones, make anthropogenic triggers for most of these changes unlikely, although decreased slope sediment shear strength arising from anthropogenic fire cannot be excluded in the case of the two dated soil failures logged in Section GL1. While it is possible that early inhabitants of the valley caused those fires, the soils were thin and immature, and apparently carried no woody vegetation, so the purpose of any such burning is obscure. The evolution of the landscape through debris flows, soil erosion, channel avulsion, and floodplain aggradation, may therefore mainly be attributable to relative changes in past sediment:water budgets of slope and river systems, all of which reflect intrinsic, local, and/or climatically triggered processes.

Currently available evidence suggests that there was little activity on the steep, gullied slopes above Gleann Lichd House between about 7.5 and 4.0 cal ka BP. Alternatively, the apparent absence activity at this time may reflect burial of evidence by later deposits, on the large, though probably not the small, fan (Section 6.4.1.1). The unknown depth to bedrock beneath the lower large fan, prevents assessment of the probable number of stacked flows, hence the strength of the argument for burial of evidence.

During the same time interval, archaeological records (e.g. Edwards & Rowntree 1980) show increasing human impact on landscapes. Early Holocene adjustments to deglaciation may therefore be decoupled from later Holocene landscape remodelling by a period of slope and river channel stability which lasted for more than three thousand years. Yet in the oceanic, high rainfall and winter snow area of north-west Scotland, there is no reason to suppose that there was an absence, during these millennia, of occasional, intense, random and seasonal run-off, via long, rock-floored gullies. If the absence of mass movement is real, the relationship between precipitation, run-off and slope instability must therefore have been strongly mediated by another (or other) environmental factor(s).

Frequent freeze-thaw arising from highly variable snow cover and run-off from the long, steep, rock-floored gullies above the Gleann Lichd fans may account for the apparent dominance of rock-sourced, rather than till-sourced, fan material (Section 6.3.1 above). In these conditions, frequent, small scale transport of till from gully walls to the valley floor is likely. Large fan-building flows, in which fresh rock fragments predominated, may only have occurred when long-term weathering of bedrock produced substantial unstable masses of debris. If the system has indeed been weathering-limited, the timing of both the accumulation of such debris and its erosion can be explained in terms of interactions between weathering rates, local conditions, the presence of excess shallow groundwater, and the passage of time. If sediment availability was not a constraint - this is, if the system was transport-limited, flows are likely more closely to reflect the timing of rare, extreme rainstorms.

Sustained high water tables, run-off, and intensified shallow subsurface flow would tend to increase the rate of hydrolysis, and hence outcrop disaggregation. During periods of relatively high humidity, the frequency of intense storms capable of

destabilising the slopes might also increase. But intense storms cannot generate mass movement if they coincide with a time when there is little unstable debris on the slope to be moved, or if it is securely contained in interim storage systems. Sediment starvation is, therefore, one possible explanation for the long period of slope stability. On the other hand, a lengthy period of enhanced weathering may ensure that the capacity of slope sediment stores is gradually exceeded even though groundwater conditions remain unchanged. Late Holocene activity could, if this explanation applies, reflect the slow weathering rate of bedrock rather than being a direct response to changing levels and patterns of precipitation.

The presence of large debris flows and late Holocene fan systems in Upper Gleann Lichd demonstrates that slope system thresholds have been exceeded. But without corroborative data it is not possible to say which factor or factors were dominant at the time of the slope failures. Nor can it be assumed that the parameters of failure were identical for each event.

Corroborative evidence might be sought in coeval processes affecting each of the adjacent fans, and the floodplain. Unfortunately, the dated record for the small and large fans is sequential, not parallel in time, and there is limited overlap between the fan and floodplain dates. However, most of the small fan aggraded after 4.0 cal ka BP (6.4.1.1 above). A match between timing of these debris flows and the floods which carried cobbles to the already vegetated palaeochannel is possible, but cannot be confirmed from available data. Similarly, although there is no direct evidence that the large fan was active before 1.2 cal ka BP, its architecture strongly suggests that possibility (6.4.1.2 above). A tentative conclusion therefore, is that, in contrast to very infrequent disturbances in previous millennia, a number of intense meteorological events between about 4.0 and 1.2 cal ka BP reshaped both the lower slopes and the floodplain. Differences in the storage capacity of the gullies feeding the two fans, together with differences in the morphology, mineralogy and shear strength of higher level outcrop, probably meant that aggradation of the two fans was asynchronous within this broad period of time. The timing of these events reflects the availability of masses of unstable bedrock which had formed as a result of the weathering of bedrock. Both the weathering products and the frequency of destabilising precipitation may be climate related, but that cannot be confirmed from this site alone.

From the dated debris flows on the large fan after 1.2 cal ka BP, and the stratigraphy of

Section GL5 (Fig. 6.12), there was an absence of sediment deposition on the floodplain during the final (dated) stages of aggradation of the distal large fan. This can be inferred from the fact that thick peat abruptly replaced rounded, clast-supported river gravels, possibly as early as 3.0 cal ka BP. Between this date and the start of sandy/silty overbank deposition at after 545 cal BP, sediment appears to have been contained within the river channel. If the source of river sediment has been correctly inferred as eroded till (6.4.2 above), this implies stable till slopes and/or a lack of overbank floods for at least the preceding 1000 years, and probably much longer. It also implies, as on Creagan a' Chaorainn, a difference in the environmental conditions which produced rock debris, and those which released fines from till. The fresh till failure opposite the fans possibly dates from the time of deposition of silts and sands in overbank floods. No change in its morphology was observed over four field seasons, indicating that it is not responding rapidly to short-term triggers. However, its cause - a spring line - is known. The suggested co-incidence of frequent, moderate, overbank floods with a new spring line, suggests elevated precipitation, although this circular argument is open to reassessment. The exposure of fractured, over-steepened bedrock beneath the till demonstrates a link between elements of the slope system, which is capable of triggering further change.

6.5   SUMMARY AND CONCLUSIONS

The Holocene history of Upper Gleann Lichd remains to be fully reconstructed. However, it is clearly not a 'relict' landscape, in the sense of preserving the morphology imposed by early postglacial equilibria. A chronostratigraphy based on the known and inferred evolution of the floodplain and the two fans is shown in Figure 6.22. The principal uncertainty which it embodies is the extent to which the dates and other evidence have permitted realistic interpretation of the full sequence of events. Because of this lack of precision, the term 'event stratigraphy' is avoided.

If the provisional interpretations set out in Section 6.4 are correct, the landscape adjusted to postglacial conditions before 7.5 cal ka BP. Only after this date (as on Creagan a' Chaorainn) has preserved organic material been found. For the next three thousand years, the steep slopes and valley floor may have been largely stable. The first of a sequence of disturbances began at about 4.5 cal ka BP when the river channel switched to the far side of the valley, incising or re-excavating a bedrock channel to a depth which prevented reoccupation of its former course except in rare major floods.

Although no dates after c. 3.9 cal ka BP are available from the small fan, aggradation did not cease then, since two-thirds of the sequence lies above this level, with the rock debris:soil ratio (hence rate of aggradation) increasing markedly in the final (?late Holocene) stages (Section GL1 Fig. 6.9). The size, and complex architecture, of the large fan, suggests that it too was actively aggrading well before the earliest dated deposit at 1.2 cal ka BP. On the floodplain, thick peats devoid of sand and silt replaced cobble lag, possibly as early as 3.0 cal ka BP. Mineral sediment only began to build up the floodplain between 545 and 510 cal BP, probably due to a combination of frequent, moderate, sediment-laden floods with the release of sediment from till slope failures.

If these interpretations are reliable, judging by the time required for slope podzol evolution and erosion (>300 years, Chapter 5), the recovery time between till slope disturbance and establishment of new equilibria is hundreds of years. In contrast, the weathering cycle, which drives bedrock failure and fan aggradation, may operate on a time scale of 10^3 years.

In conclusion, Holocene landscape evolution in Upper Gleann Lichd can be understood in terms of a combination of in-phase and out-of-phase, climate-influenced sediment:water budgets, which have interacted in complex ways with local controls and bedrock weathering rates. The coupling of slope and fluvial systems is best understood in terms of changing sediment:water ratios. In exposing over-steepened sections of fractured bedrock, climate-influenced till failure, together with long-term weathering, may have the capacity to drive continuing slope and fluvial system modifications.

FIGURE 6.1 Gleann Lichd and Glen Shiel location map
1 Upper Gleann Lichd; 2 Allt a' Mhuing fan; 3 Allt na Carnach fan
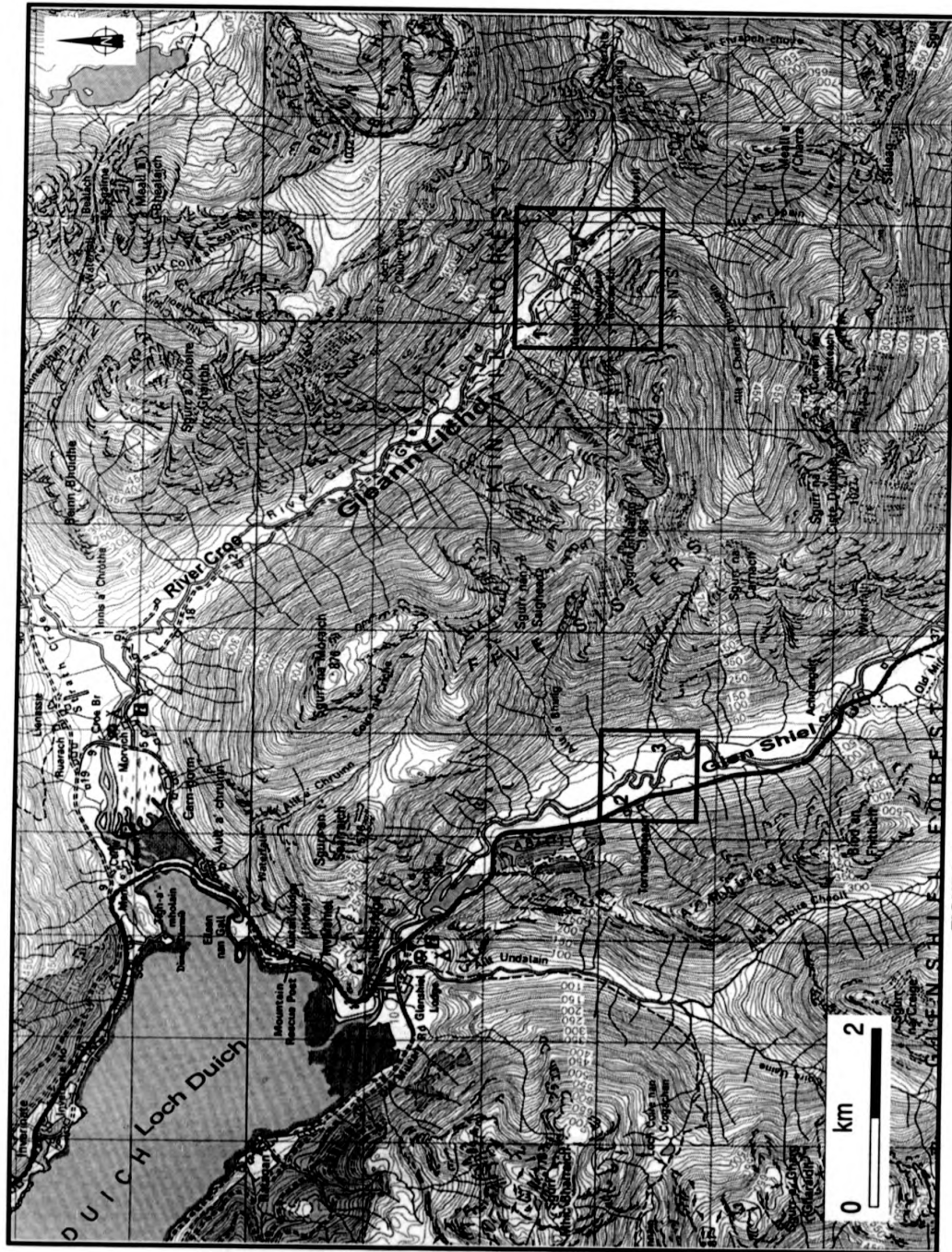
Based on the OS 1:50000 scale map Sheet 33 © Crown copyright
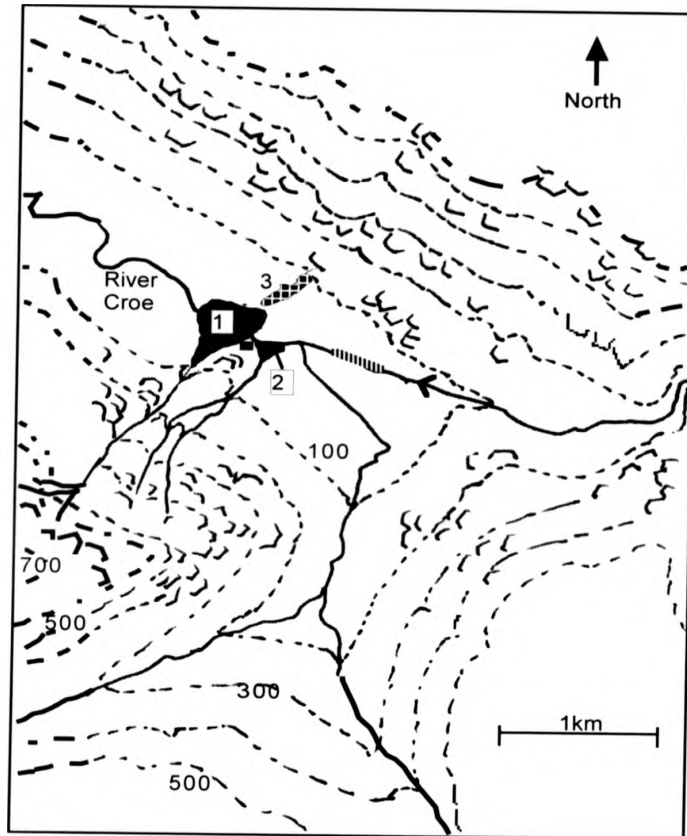
FIGURE 6.2
Upper Gleann Lichd
(Based on OS 1:25,000 map Pathfinder 221 (NH 01/11). Crown copyright 1990)

KEY

			outcrop
1	The large fan	■	Gleann Licht House
2	The small fan		Rock gorge
3	The till failure		100m contours
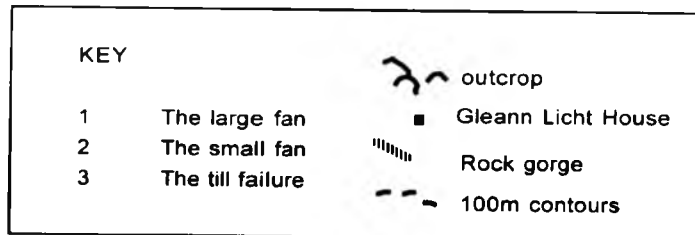
FIGURE 6.3
The till failure, upper Gleann Lichd
View from Gleann Licht House across the toe of the large fan and the River Croe
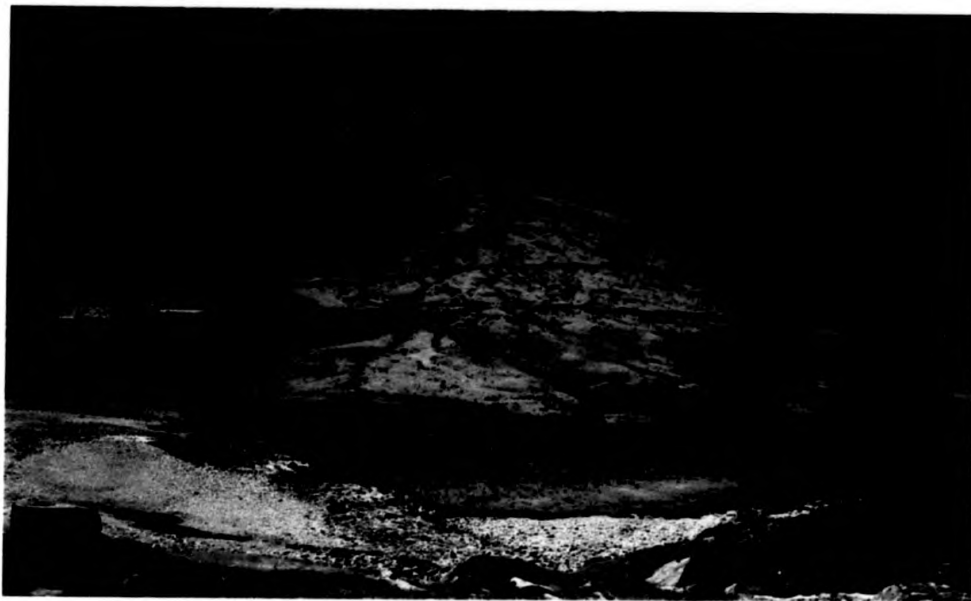
FIGURE 6.4
View across the R. Croe to the large fan and Gleann Licht House.
Sections are exposed in the meander bend below the house

FIGURE 6.5
Looking NW down Gleann Lichd across the toe of the small fan (S)
and the lower large fan (L)
Logged Sections GL1 is adjacent to 'S'. GL2- 5 are to the right of 'L'
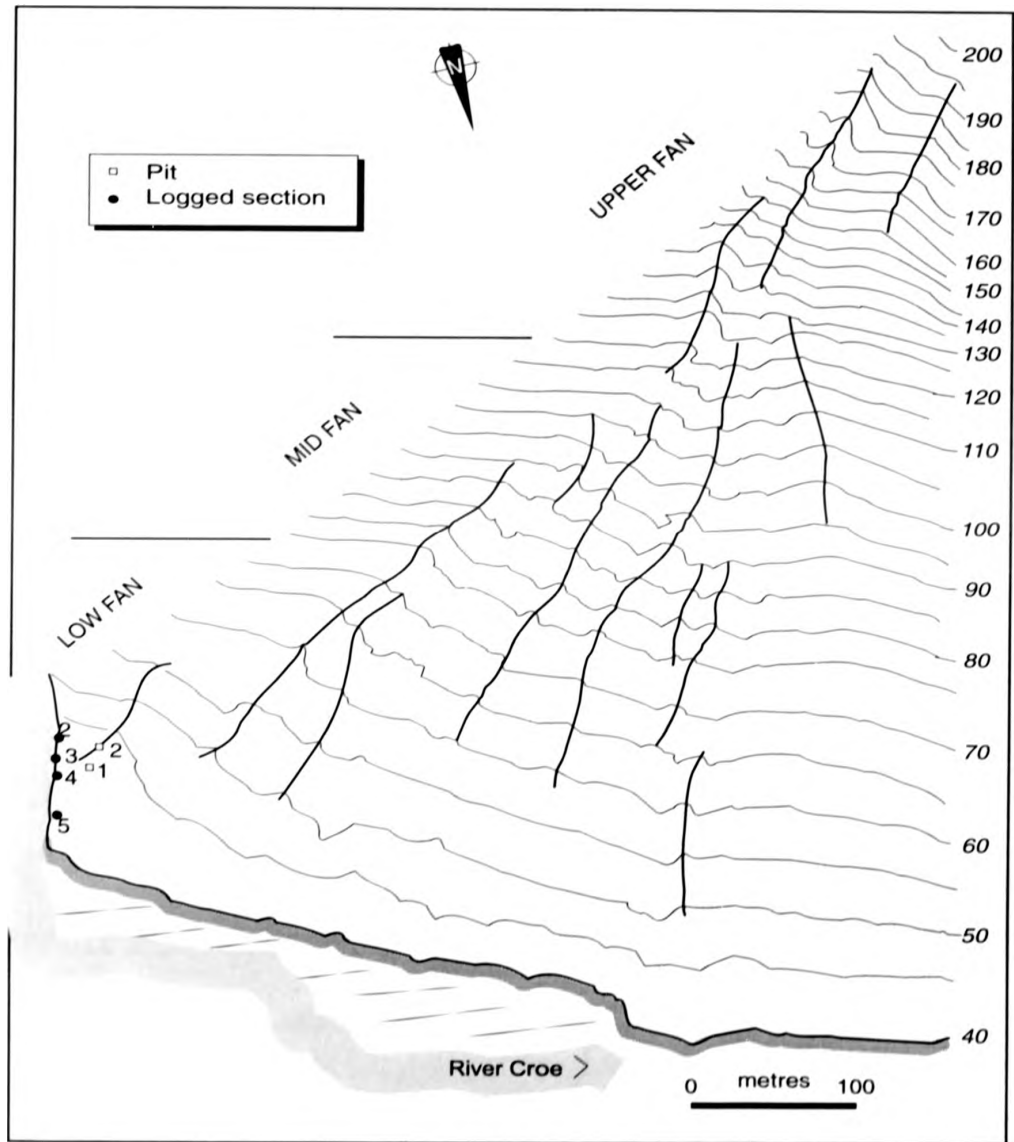
FIGURE 6.6
Survey map of the central area of the large fan

| shallow drainage channels | fan toe ramp | contours\ mAOD 50 | floodplain |
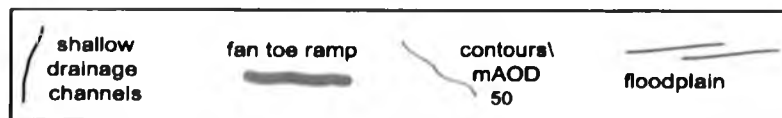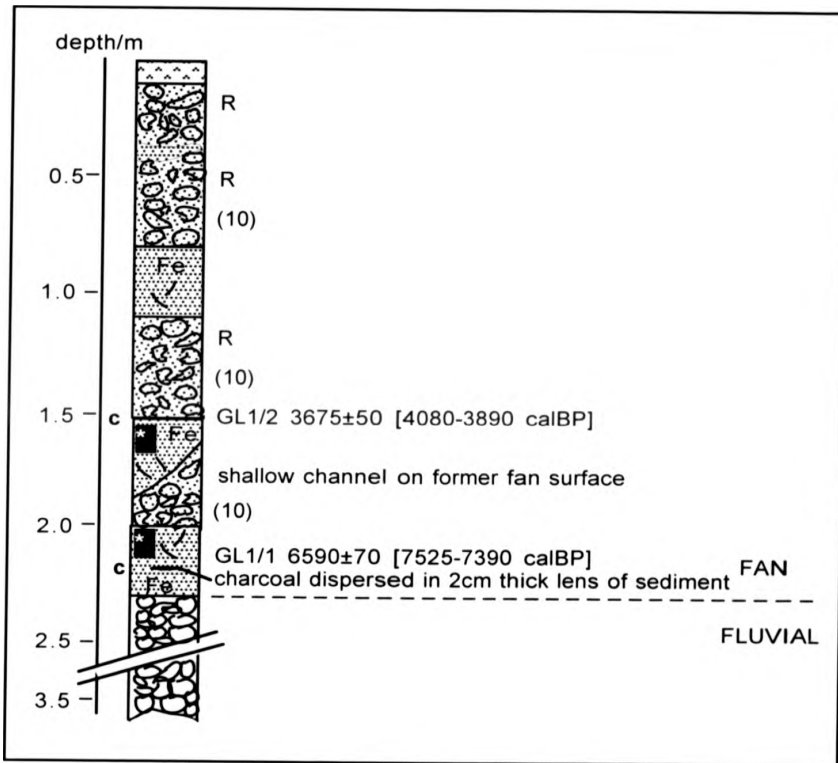
FIGURE 6.9
Section GL1: The small fan, Upper Gleann Lichd

⬛ Kubiena tin samples & photomicrograph
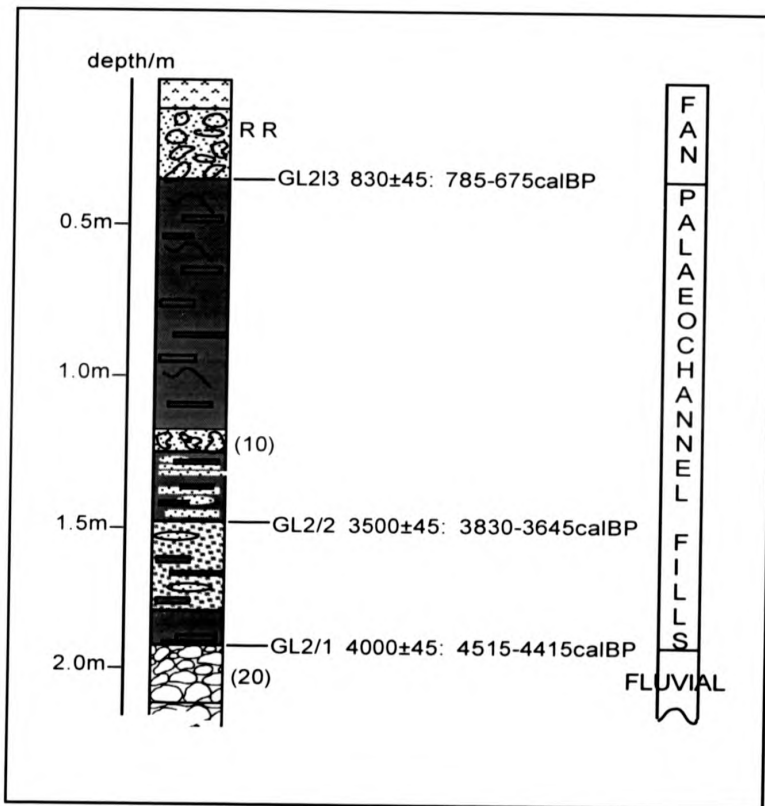
FIGURE 6.10
Section GL2, The large fan, upper Gleann Lichd
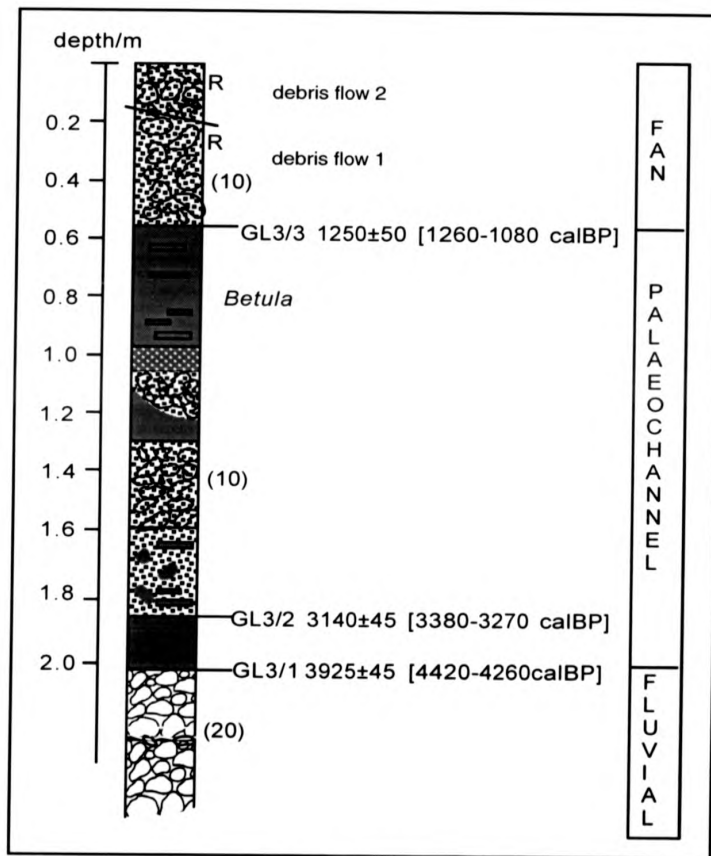
FIGURE 6.11
Section GL3: The large fan, upperGleann Lichd
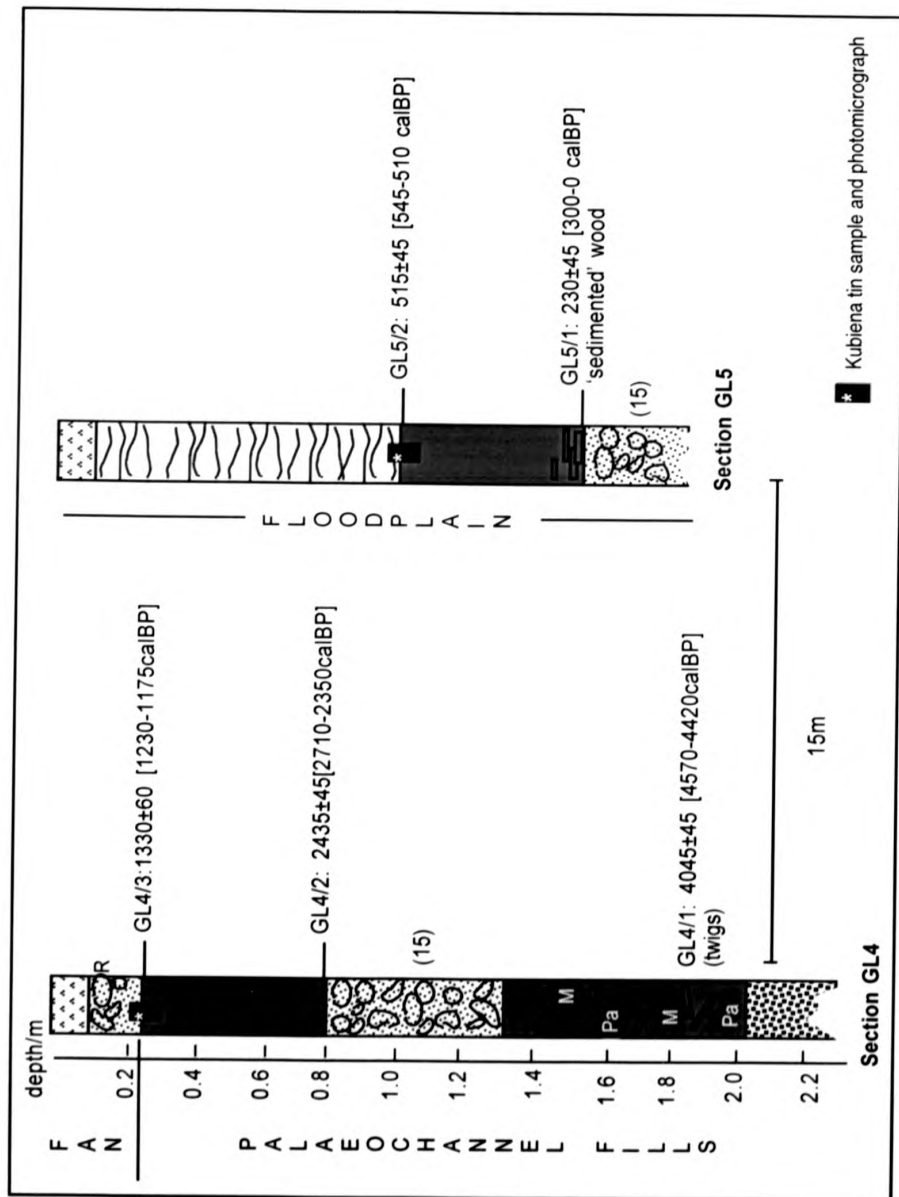
FIGURE 6.12

Sections GL4 & GL5: The large fan, upper Gleann Lichd
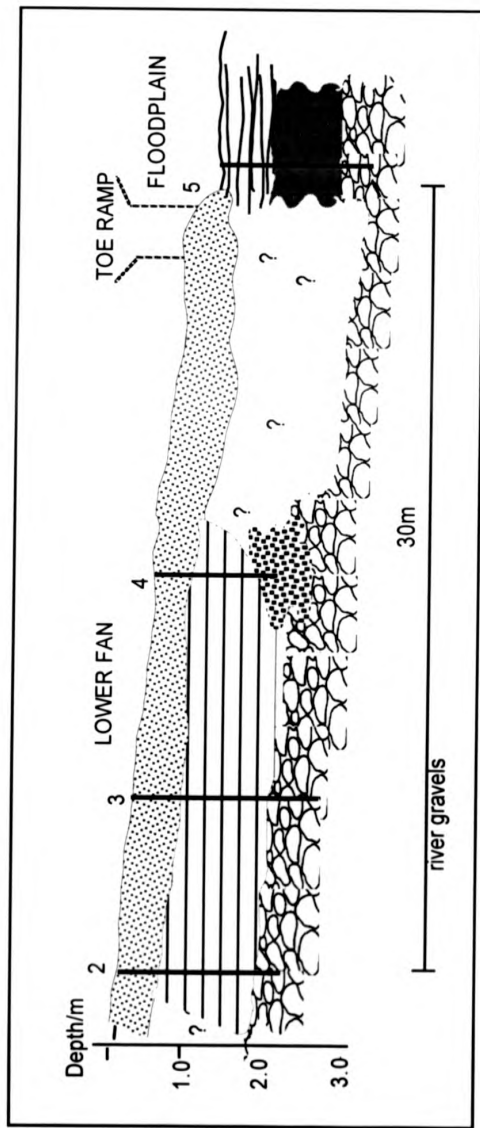
FIGURE 6.13

Schematic interpretation of the long-section of the up-valley margin of the Gleann Lichd large fan toe.
The view is down valley, to the north-west. Logged section numbers as in Fig. 6.6

coarse fan debris, reversely graded

palaeochannel fill of peat, muds & gravels (See Figs. 6.11, 6.12)

coarse pebbly sand

overbank deposits

peat

winnowed (indurated) river gravels

? no exposure; inferred boundary

2-5 : logged sections GL2 -5
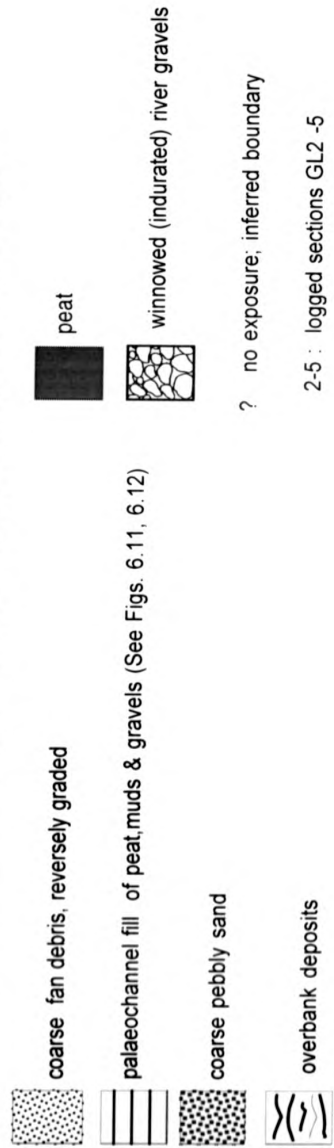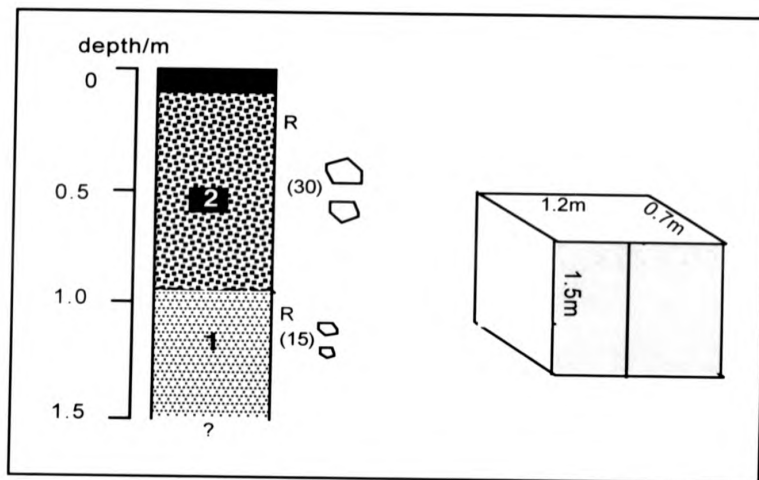
FIGURE 6.14
Gleann Lichd large fan: Pit 1

1   debris flow 1

2   debris flow 2

All clasts very angular; abundant silt in both horizons
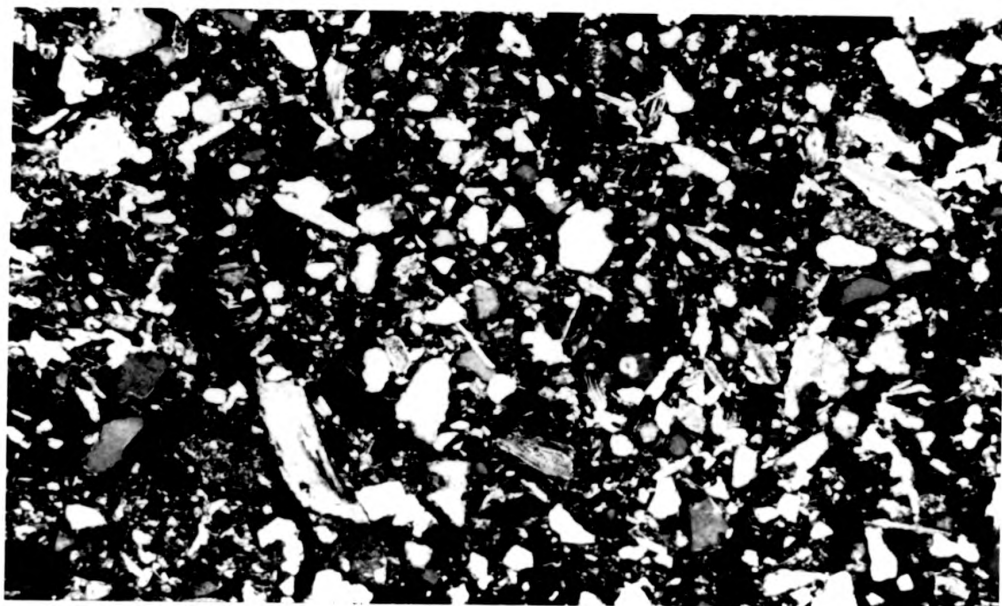
FIGURE 6.15a (x 40 XPL) GL1/1
Poorly sorted angular quartz & mica laths in a matrix of OM
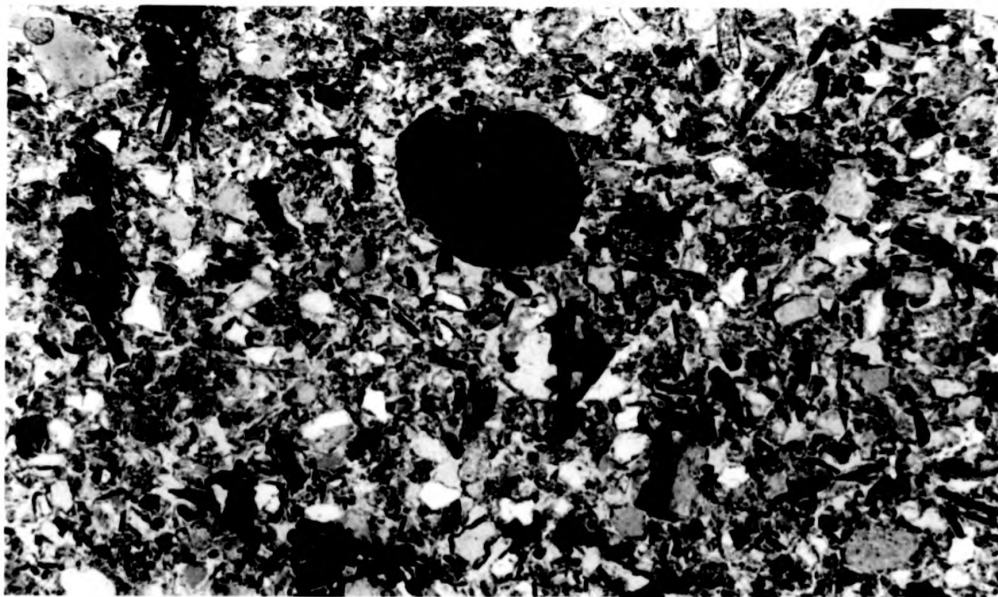
FIGURE 6.15b (x 40 PPL) GL1/1
Adjacent to Photo 1: OM matrix packed with mineral grains and scattered plant remains
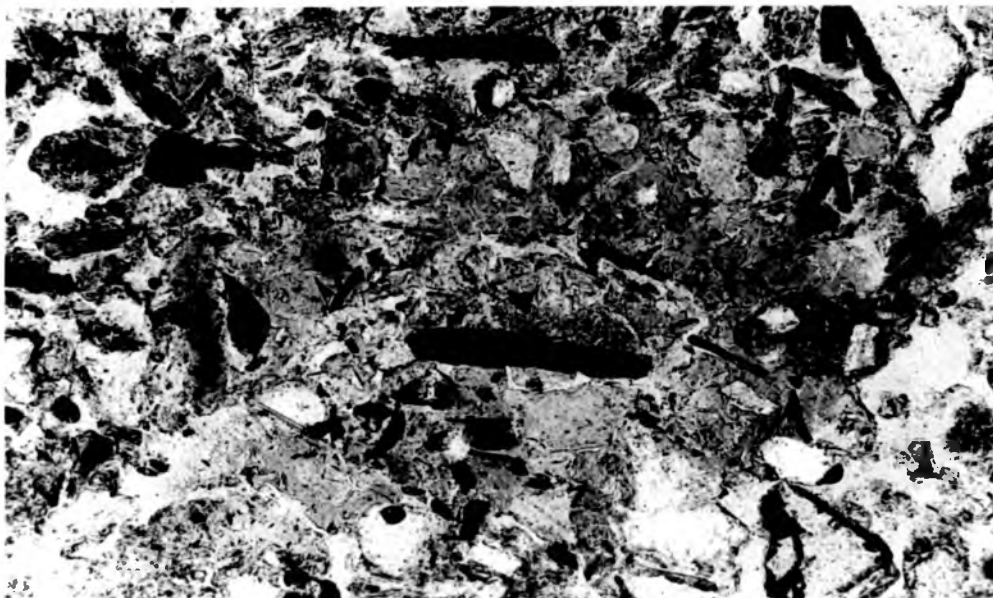
FIGURE 6.15c (x 40 PPL) GL1/1
A coherent soil particle c. 100µ in diameter; silt, plant fragments and
faecal pellets (?) in light brown OM; note muscovite laths RHS
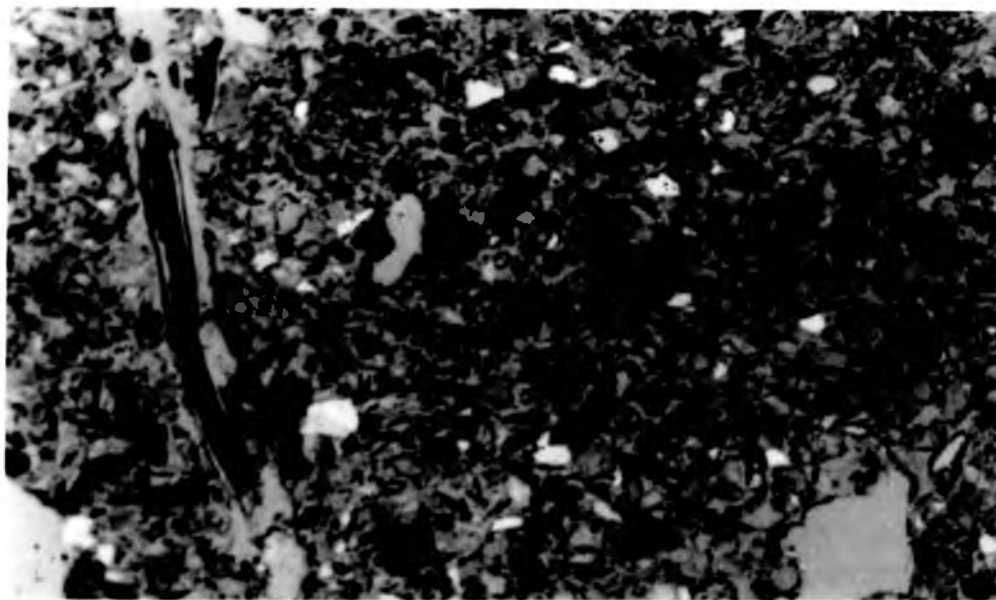
FIGURE 6.15d (x 20 PPL) GL1/1
Vertical plant fragment partially filling elongate vug; other more equant vugs have a dark
Fe/Mn(?) coating
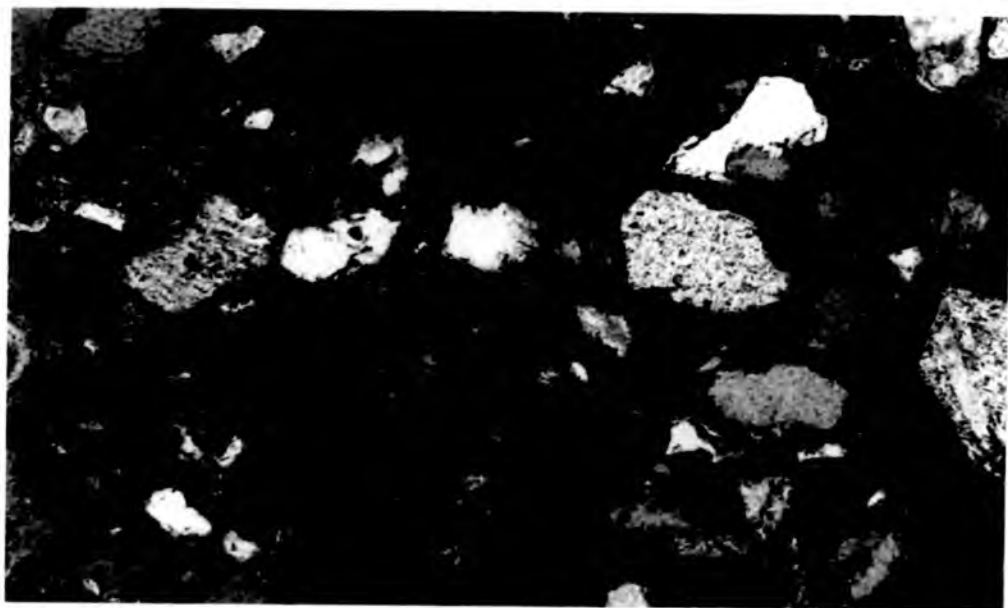
FIGURE 6.15e (x 100 XPL) GL1/1
Grain and vug coatings on silt are typical of soil leaching; incomplete coatings have no
consistent way-up, so formed before deposition of the sediment
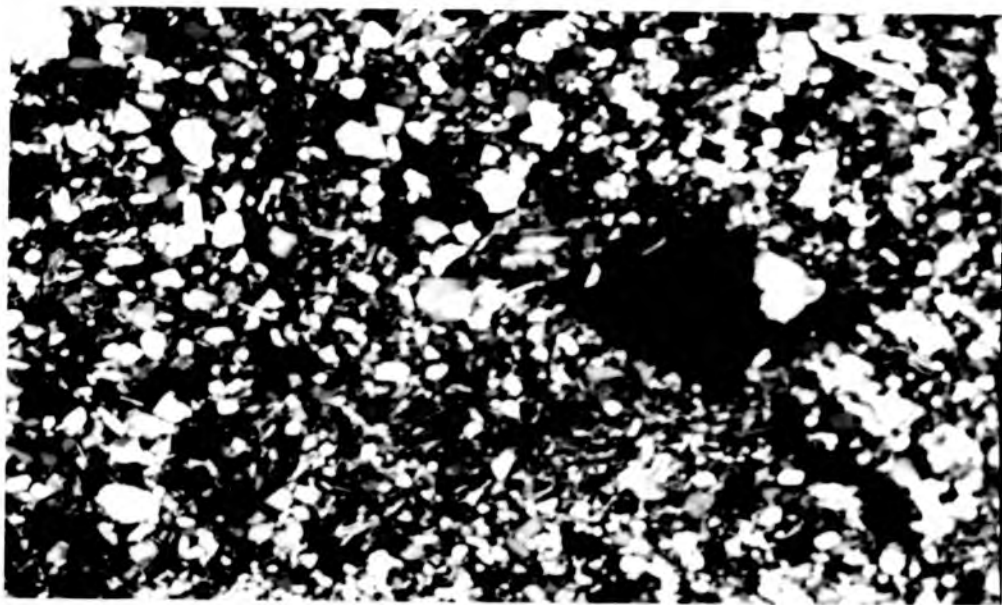
FIGURE 6.16a (x 20 XPL) GL1/2
A 2mm strained quartz particle is mixed with silt and sand containing randomly oriented mica
particles. Plant fragments are rare
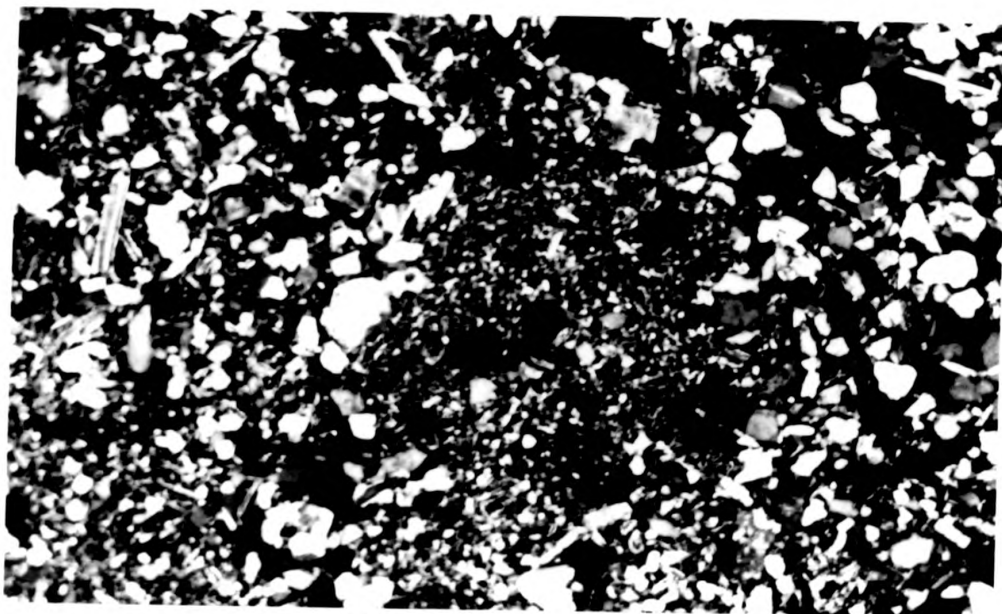
FIGURE 6.16b (x 20 XPL) GL1/2
Silty soil particle c. 2mm in diameter in fine to medium angular, poorly sorted sand
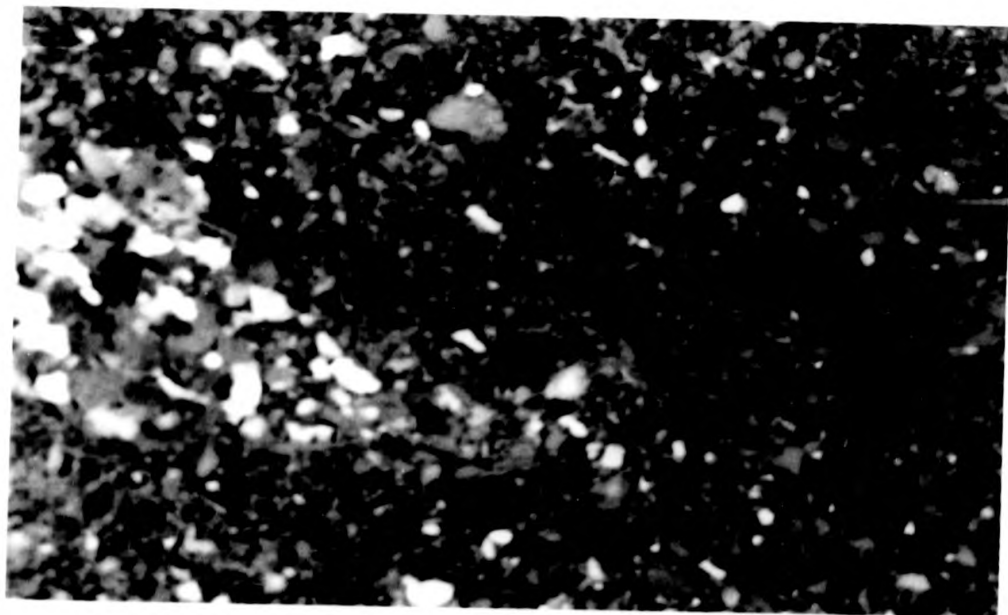
FIGURE 6.17a (x 20 XPL) GL4/3
Angular, unsorted mineral grains mix with peat containing woody fragments
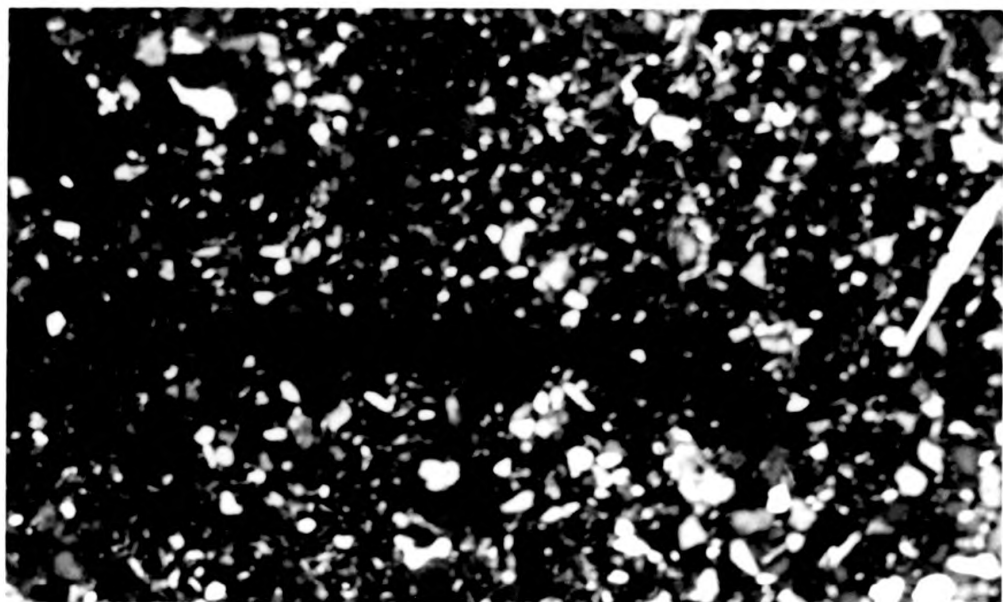
FIGURE 6.17b (x 20 P/XPL) GL4/3
A randomly oriented wood fragment, mica laths and angular quartz mixed with peat at the base of the debris flow
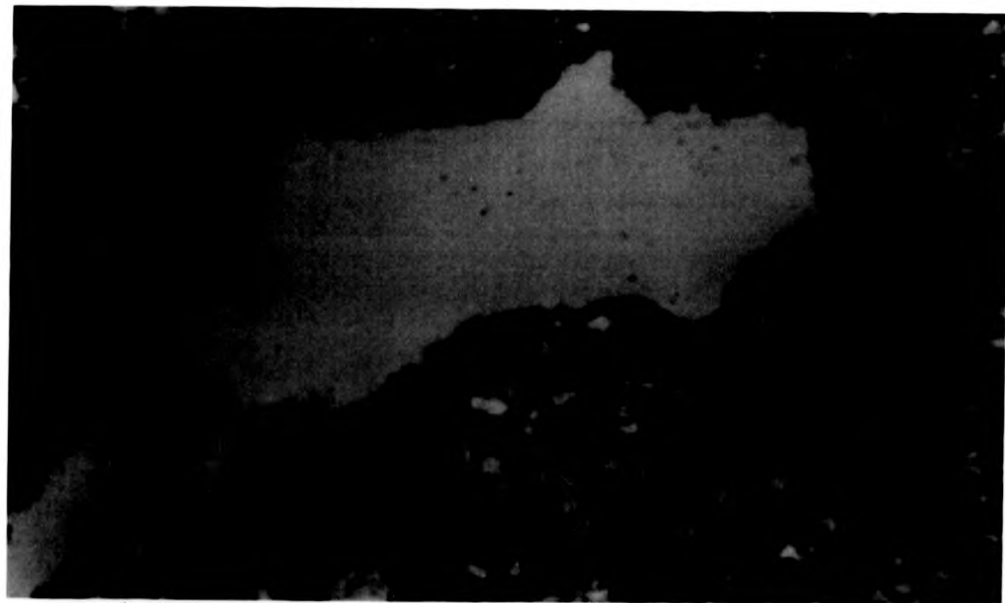
FIGURE 6.17c (x 20 PPL) GL4/3
Vugs pseudomorph woody fragments in peat

257
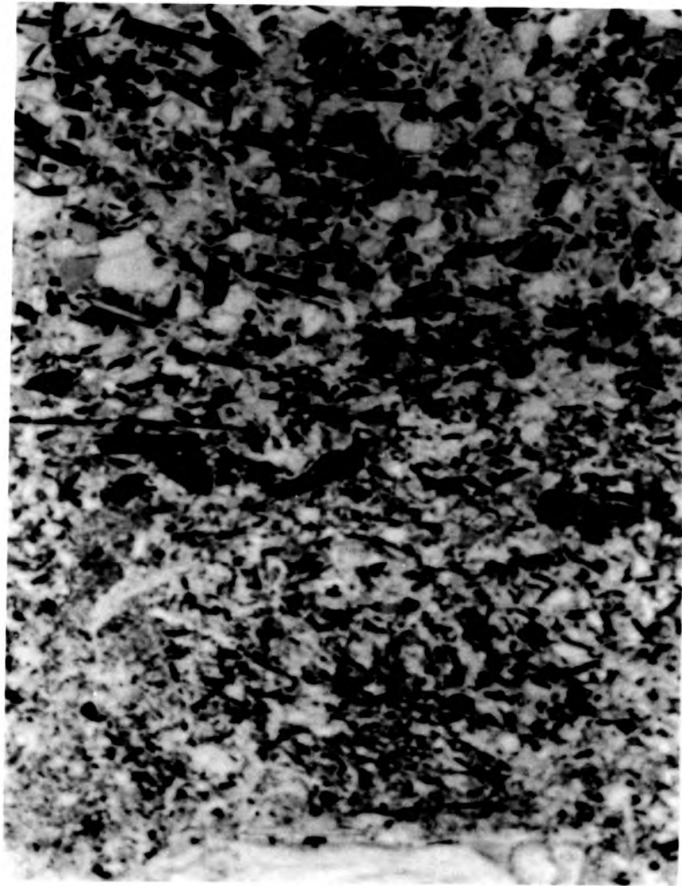
FIGURE 6.18 (x 20 P/XPL) GL5/2
Contrast with Fig. 6.17b: a sharp peat/mineral boundary and
sub-horizontally layered mica laths in the mineral layer
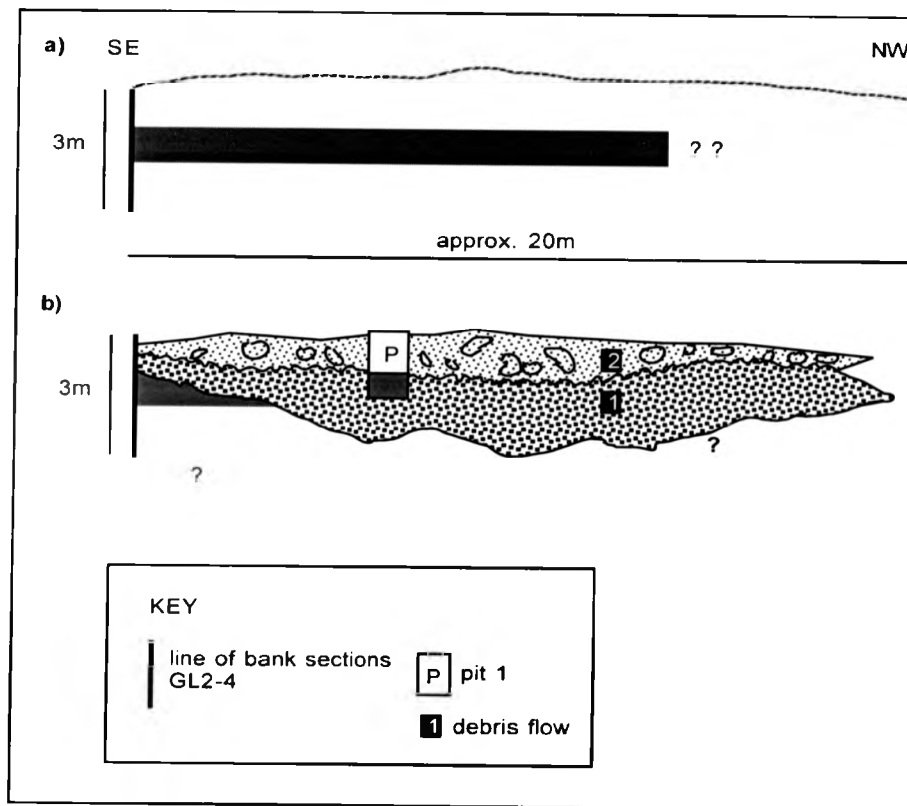
FIGURE 6.19
Inferred cross-section of the toe of the large fan, Upper Gleann Lichd
The view is up fan.

a) Inferred original extent of palaeochannel peat

b) Debris flows replace palaeochannel fills
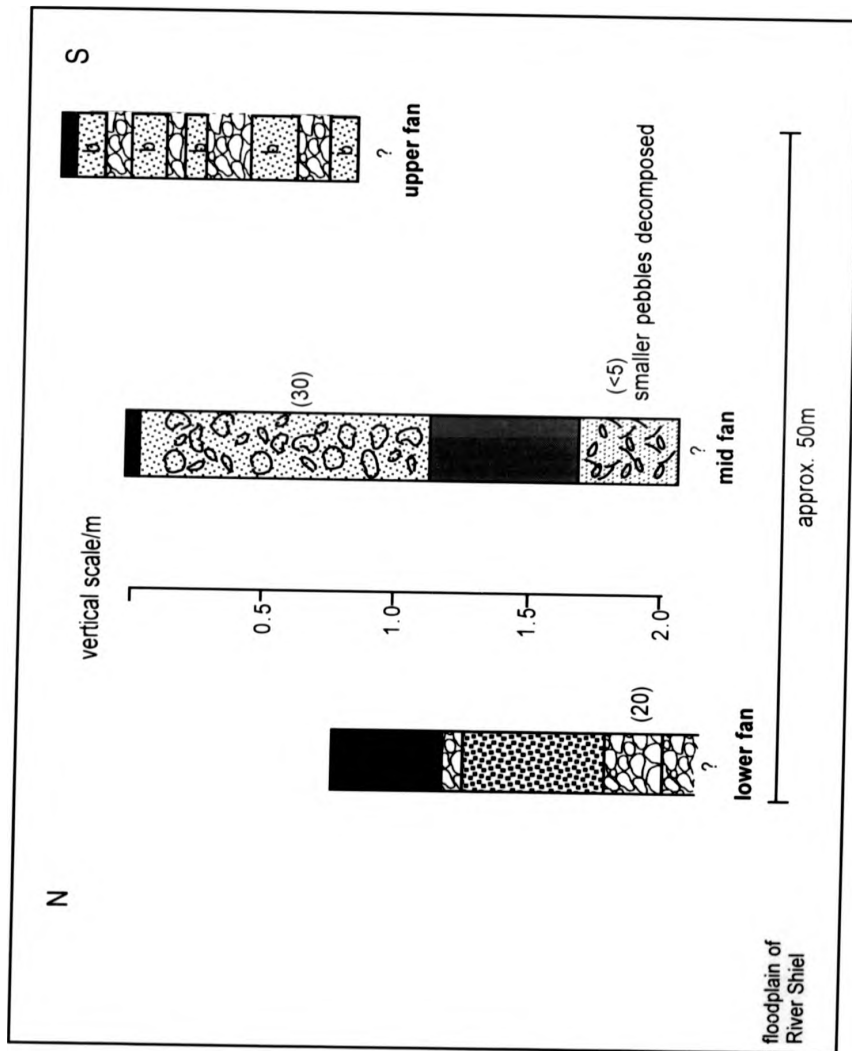
FIGURE 6.20

Three successive vertical sections through the Allt a' Mhuing fan, lower Glen Shiel.
The fan surface slopes at 10 - 15°, decreasing downslope, to the left of the diagram.

(There is no, or poor, exposure between sections.)

FIGURE 6.21
Allt Carnach fan, Glen Shiel
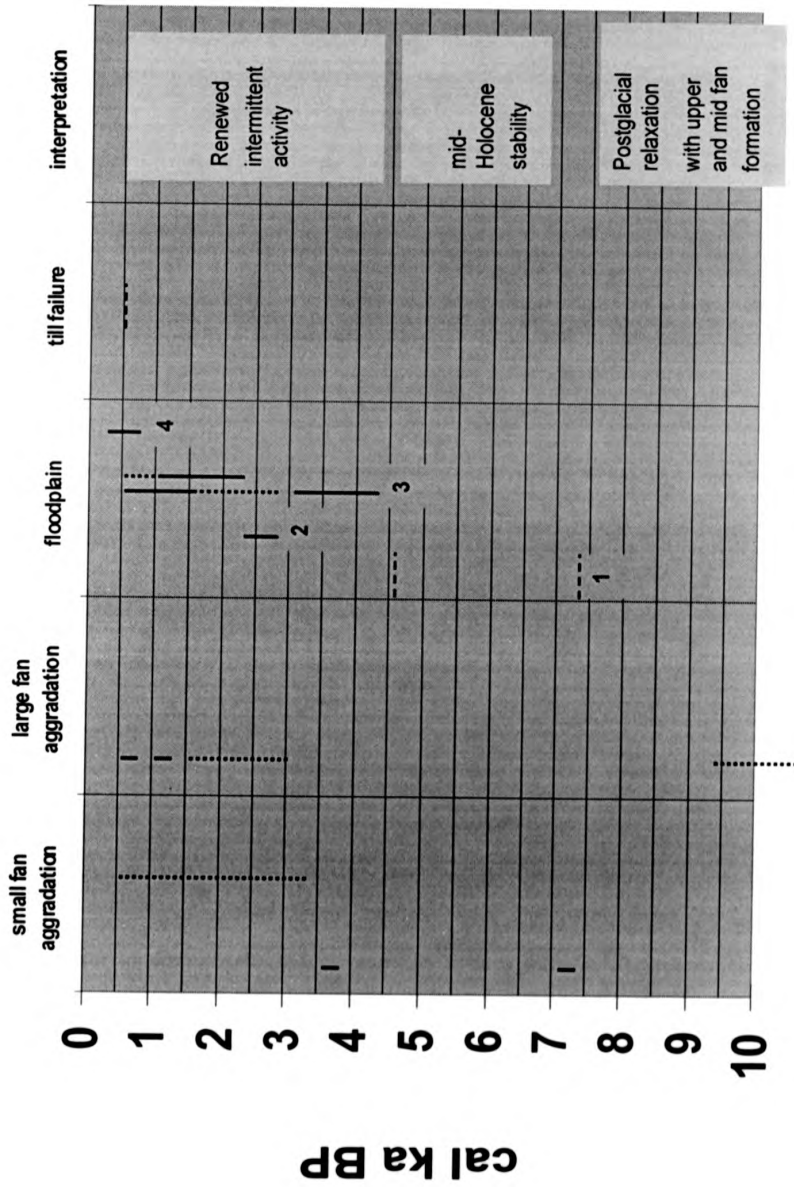
FIGURE 6.22 HOLOCENE CHRONOSTRATIGRAHY OF UPPER GLEANN LICHD

Dashed lines represent the inferred timing of events (see text for details). Peat growth excepted, time lines indicate a period of time within which discrete events, not continuous processes, occurred.

1 channel avulsion; 2 rare, large floods; 3 peat growth; 4 overbank deposition

CHAPTER 7 DISCUSSION: EVIDENCE FOR MID AND LATE HOLOCENE LANDSCAPE EVOLUTION AND ITS ENVIRONMENTAL TRIGGERS IN NORTHERN SCOTLAND

7.0   INTRODUCTION

This chapter integrates data from Chapters 4 - 6. The nature and validity of the chronostratigraphy based on these data is discussed in Section 7.1, and critically examined in the light of published data from Scotland (Table 2.1) and data from the wider region of N. W. Europe. The evidence for triggering mechanisms in Northern Scotland is reviewed in Section 7.2. Sections 7.3 and 7.4 consider slope sensitivity and mechanisms of mid and late Holocene landscape evolution in Northern Scotland.

7.1   CHRONOSTRATIGRAPHY

Evidence of slope evolution preserved in upper Gleann Lichd and on Creagan a' Chaorainn dates from the mid and late Holocene. Sites in Strathfarrar, Glen Cannich, Strathdearn (Chapter 4) plus Gleann Chorainn (Chapter 4, Section 4.3, Chapter 5, Section 5.1) reveal the stratigraphy of a period of Holocene time whose extent is partly undefined, but which, in every case, post-dates sustained growth of pine woodland and blanket peat. The record is of slope failure, river channel avulsion, peat growth and degradation, soil erosion and fluvial activity, accompanied by a rise in the water table which may have caused sections of river to expand temporarily to form ribbon lakes. The youngest horizons, deposited by high frequency moderate floods, immediately underlie immature modern soils. One date from Gleann Lichd indicates that the frequent flood deposits immediately precede modern soils in age.

The following discussion draws on Figure 5.22 and Figure 6.22 as well as the dates of small fan aggradation in upper Gleann Lichd at about 7.5 and 4.0 cal ka BP. It takes into account the fact that the radiocarbon dated record is incomplete, but that the number and frequency of debris flows increased towards the end of the mid Holocene and intensified in the late Holocene, on Creagan a' Chaorainn (Section 5.4.4), and in upper Gleann Lichd (Sections 6.4.4.1, 6.5).

Coincidence cannot be excluded, but the earliest preservation of organic matter beneath debris flows at both dated locations occurred at about 7.5 cal ka BP. On Creagan a' Chaorainn this date plausibly marks a boundary between early, frequent slope instability with poor preservation of organic horizons, and, more speculatively, an extended period

of general equilibrium which lasted for several millennia. However, a marked change in system state could have arisen as a result of quite subtle changes in threshold conditions (Chapter 5, Section 5.4.4). At both sites, the available evidence points to the mid-Holocene prior to about 4.5 cal ka BP (GL) and 4.3 cal ka BP (GCh) as a period when slope stability was seldom interrupted.

By about 7.1 cal ka BP, trees and peat-forming vegetation had already blanketed the stabilised till slopes of Creagan a' Chaorainn, although gullies were still active. Between 7.1 and 6.1 cal ka BP gullies stabilised and filled with vegetation, undergoing a switch from net erosion to net deposition. This was probably most readily accomplished in the presence of a volume of groundwater, which was much reduced in comparison to that which previously maintained erosional conditions. Yet during the same period of time, podzol and peat development testify to persistent, at least seasonally damp, leaching conditions, with soil and till erosion occurring only at long intervals (hundreds of years as a minimum). At this stage, just as a few hundred years earlier, quite subtle alterations in the balance between interacting system thresholds, may have resulted in a marked change of state.

Slope mass movement continued to be rare until about 4.3 cal ka BP (Fig. 5.22). Thereafter, although individual dates are few, the frequency and volume of flows of rock-sourced debris increased on average by a factor of nearly nine compared to the previous two and a half thousand years, with flow frequency increasing up the section (Fig. 5.14). From about 2.3 cal ka BP onwards, bedrock and till failures, accompanied by slopewash, lasted for some 1200 years. However, 2.3 cal ka BP is a spurious 'boundary' which the number of recorded, but undated debris flows suggests might be extended backwards by up to a thousand years with more extensive dating.

Due to the combination of erosional boundaries, a limited number of dated horizons, and the uncertainties inherent in dating palaeopodzols, the error bars on the boundaries between system states on Creagan a' Chaorainn are too large to attempt a precise event stratigraphy. Nevertheless, a trend is discernible. It can be summarised as mid Holocene spread of tree and peat cover and gully stabilisation, with occasional debris flows, followed by full peat cover and slope stability, before a period of high water table and much more frequent and intense slope destabilisation in the late Holocene.

In upper Gleann Lichd, where slopes are steeper and longer, with more high altitude outcrop, the 'early mid-Holocene transition period' on Creagan a' Chaorainn (Fig. 5.22)

is not represented. Instead, there is an absence of evidence for any activity for three millennia after 7.5 cal ka BP (Fig. 6.22). Abrupt renewal of sediment cycling started a little earlier than recorded on Creagan a' Chaorainn, at about 4.5 cal ka BP, when the River Croe cut (?re-excavated) a new channel in bedrock. Over the next few hundred years, rare, major floods carried coarse debris into the abandoned channel, and slopes supporting developed podzols failed from time to time. Large fan morphology and stratigraphy make it improbable that the mid and lower fan began to accumulate only at the time of the earliest available date of 1.2 cal ka BP, which is taken from the distal margin (Chapter 6, Section 6.4.1.2). Like the small fan, it probably aggraded in a number of discrete events, mainly after 4.0 cal ka BP, in parallel with the start of increased debris transport on Creagan a' Chaorainn. As on Creagan a' Chaorainn, gaps in the record in the Gleann Lichd fans mean there is a poor basis for a centennial scale event stratigraphy. But here too slope activity increased from the mid to late Holocene, with an intensification in the latter period of time. The start of this period of activity, in the middle of the 5th millennium BP seems to have been characterised by infrequent, but severe, system perturbations.

Tentatively, a similar trend is discernible in the undated catchments. In Gleann Chorainn, up valley of Creagan a' Chaorainn, peat, eroded soil, charcoal and peaty soils are buried beneath flows of sand and coarse debris in the Bac an Eich fan and the banks of the incised floodplain (Fig. 5.3). These sequences record repeated fan and floodplain aggradation, which interrupted periods of stability during which organic horizons were formed. They predate the deposition of the sandy matrix of modern soils.

In Strathfarrar, the known history of the Allt Innis na Làrach fan (Table 4.1) starts with a major disturbance, in the course of which the River Farrar abandoned its ancestral course and, like the River Croe in upper Gleann Lichd, incised a new, deep, partially rock-walled channel. Pine woodland then colonised the drier ground in the former river bed. On damper ground, peat grew, full of woody fragments, and invaded by flushes of sand carried by Allt Innis na Làrach from the Lateglacial cone which had been twice incised and extended, to form a coarse medial fan. The pine trees matured, but were overwhelmed by spreading peat. Evidence for changes in slope hydrology is compatible with a conclusion that the rise in water table is not attributable simply to the growth of the peat itself: during this period, sand from the eroding slope above was increasingly spread across the peaty fan terminus by sheet floods; the slope erosion which was the

source of that sand, culminated in a massive grain flow which incorporated eroded soil and charcoal; above the grain flow deposit, overbank deposits indicate the abrupt onset of frequent, decadal(?), floods. The whole sequence records sheet flooding followed by extreme flooding, then increasingly frequent, though more moderate flooding and slope destabilisation. A few kilometres further up Strathfarrar, pine beneath the Braulen fan (Fig. 4.21) was similarly overwhelmed by peat, whose own growth was later interrupted by incursions of coarse fan debris consisting of eroded till. Rising water tables after the spread of pine to low lying areas were therefore not restricted to a single site, although dating of pine remains is required to test whether the pine stands are of similar age.

The Liatrie fan in Glen Cannich (Fig. 4.23a) contains layers of cement, probably derived from disintegrating slope peats, and formed at a time when the surface of the river Cannich was at least 5m above its current level. Contour lines suggest formation of a ribbon lake if the water surface was raised to this level. A similar cement layer in the Allt Crumaidh fan a few kilometres up-valley from Liatrie, is overlain by a massive soil failure suggestive of wet, destabilising conditions (Figs. 4.26a and b). Although the nearby hydro-electric dam has caused changes to the river which have obscured any evidence of alluvium onlapping the Allt Crumaidh fan toe at a time of raised water levels, cement there also occupies a thin, horizontal horizon in coarse, permeable fan debris, now being incised by the fan stream. As at Liatrie, it cannot be explained by percolation of iron-rich fluids down from the surface, and appears to conform to the model of deposition suggested for the Liatrie fan. That model requires a high water surface abutting the fan toe, and depositing a barrier of relatively impermeable fine sediment (Fig. 4.23b). If the source of the cement is correctly inferred to have been eroding slope peat (Chapter 4, Section 4.2.1.1), the maximum age of these fans is post early Holocene. While peat is known to have formed in some localities at a very early date, its degradation and erosion to release iron-rich water, is more likely to have occurred during a period of intensified humidity, than during the relatively warm dry climate of the early Holocene (Chapter 2).

Interpretation of the history of the Allt Coire Eòghainn fan in Glen Cannich and the deposits exposed on the margin of the Loch Mullardoch reservoir (Ch. 4, sections 4.5.2, 4.5.3), is complicated by neoseismicity. However, the stratigraphic sequence in the Allt Coire Eòghainn fan is similar to that at other localities in two main respects. A sequence of floodplain sands and muds, interpreted as an upward-fining fluvial sequence, onlaps

basal slope peats and abuts the toe of a coarse fan several metres above the present wide, marshy floodplain (Fig. 4.25c). As at Liatrie and probably, Allt Crumaidh, the floodplain sequence could only have been deposited in this location if the river had risen to a level at which contours suggest formation of a ribbon lake. On the fan surface, peat with abundant wood was overwhelmed by coarse slope debris.

In summary, water table rise and slope movement appears to have been not uncommon in Northern Highland catchments, wherever long sequence Holocene stratigraphy has been investigated. Pending further dating, these events may have occurred across all the five catchments identified for this thesis, from the mid Holocene onwards, intensifying in the late Holocene. After about 2.3 cal ka BP, but probably at least several hundred years earlier, sandy flushes from degrading slopes reached peat in areas normally devoid of mineral sediment (Strathfarrar, Creagan a' Chaorainn); thick floodplain peats accumulated in conditions of persistently high water table (Gleann Lichd); peat aggrading low angle fans (Strathfarrar, Glen Cannich) and slopes (Creagan a' Chaorainn); accelerated erosion of widespread peat produced intense concentrates of iron and manganese in solution which precipitated as crude cements in permeable fans (Strathfarrar, Glen Cannich).

Overbank deposits which formed after the features just described, were found as the uppermost buried horizon of every stream bank sequence examined, from Strathdearn in the east, to Glen Cannich and Strathfarrar in central Ross-shire, and Gleann Lichd in the west. They invariably have a sharp boundary with underlying horizons, and an equally sharp boundary with overlying immature, sandy, modern soils which they immediately precede. In upper Gleann Lichd, they began to form at about 500 cal BP. Their recent termination appears to have been equally abrupt. They are interpreted here as reflecting a specific combination of environmental conditions. Till and thin blanket peat on steep, formerly stable slopes was disturbed, releasing sand and silt which aggraded distal fans. At the same time, rivers began to exceed bankfull discharge during moderate floods, which deposited fines from suspension on peaty bank vegetation. Flood frequency of the order of 10^1 - 10^2 years is a likely explanation for centimetre-scale, sometimes impersistent, stringers of peat forming between floods. It remains to be determined whether the conditions which produced these strata were not only regional in scope, but synchronous. Persistent high humidity associated with these floods is likely also to have affected subsurface flow. In Gleann Lichd, a spring line which destabilised till and

underlying bedrock fractured it may be a recent event, speculatively, coeval with the inception of overbank flooding.

Modern slope and floodplain soils in the locations overlying overbank sediments (Strathdearn, Gleann Lichd, Strathfarrar) and slope debris (Creagan a' Chaorainn, Strathdearn, Glen Cannich, Gleann Lichd, Strathfarrar) are invariably immature or peaty, showing no sign of differentiation into layers. This may reflect the fact that podzols take ≥350 years to become established (Chapter 5, Section, 5.3.6). In Gleann Lichd, roots of river bank alders less than (probably much less than) 300 years old remain *in situ* on the narrow floodplain, although the trees themselves have disappeared. So there has been limited floodplain reworking in the course of the valley's most recent development.

On the basis of the above evidence, a testable chronostratigraphy of landscape evolution across northern Scotland in the mid and late Holocene can be outlined.

1. Slope stability and low water tables evident at about 7.5 cal ka BP. Sediment starvation in gullies and thin soil development on steep, rocky slopes; spread of slope peats, with only rare system threshold breaches on gentler slopes until about 4.5 cal ka BP.

2. Large, rare geomorphic events in the few hundred years after 4.5 cal ka BP.

3. Many more slope failures after about 2.7 cal ka BP (but probably earlier) with a rise in water tables. Tentatively, formation of ribbon lakes as rivers rose metres above their present levels at this time (although base-level controls may have varied between catchments). Large soil failures, slope peat erosion and debris flows.

4. Persistent decadal flooding starting at about 500 cal BP accompanied by erosion of fines from till exposed by peat degradation.

This sequence does not constitute an event stratigraphy *sensu* Ager (1973) and Bjørck (1998) (Chapter 2, Section 2.3.1), since the multi-proxy nature of any such record, time gaps, and a general scarcity of data means it lacks precise boundaries. Unfortunately, on periodically active slopes, which generate accumulations of sediment with datable horizons, hiatuses may be the rule. However, the principle of combining diverse processes and events to form a chronostratigraphy has yielded useful data.

7.2   ENVIRONMENTAL TRIGGERS

The purpose of this section is to discuss the basis for proposing or excluding environmental triggers for the pattern of slope evolution summarised above. The extent to which the model of paraglacial relaxation and intrinsic change, overprinted by anthropogenic impact and random events, (Chapter 2, section 2.3.4), can account for the empirical data, is also reviewed.

7.2.1   Landscape 'memory' and slope sensitivity

The 'memory' that Holocene landscapes and landforms retain of past Holocene events is superimposed on a geological and glacial memory, so that system thresholds are not constants, but themselves evolve through time, on different timescales, both gradually and abruptly. Externally generated system perturbations may be superimposed on intrinsic local system inputs and feedbacks so as to initiate change leading to threshold conditions (e.g. Phillips 1999, White *et al* 1994). A modest external perturbation might be sufficient to create threshold conditions and trigger irreversible change in some cases, while only large inputs would be able to change the nature of systems which were intrinsically more robust - at that time. Systems that are intrinsically robust at Time A, may also, without any external input, progressively evolve into a more sensitive state. An identical external input will therefore have different effects, depending on whether it occurs at Time A or Time B in the landscape's history. Similarly, large perturbations will have an effect whose magnitude and duration is proportional to system sensitivity.

In consequence, an event stratigraphy of these landscapes is a composite, multi-proxy record of events and inferred processes, which are variable in time and spatially heterogeneous. One consequence is that, in the absence of overwhelming system perturbations, reconstruction of forcing factors is difficult. However, regional drivers of slope weathering, erosion and floodplain reorganisation, which persist over long time scales, may have a lasting expression in contrasting records of event frequency.

In principle therefore, slope evolution in this setting has limitations as a proxy for past environmental triggers. However, this principally affects the precision of timing and discrimination of variables. It does not invalidate the proposition that lower slope sequences of Holocene age, in conjunction with fluvial stratigraphy, preserve a record of long term sensitivity to environmental forcing factors, including precipitation.

7.2.2 Paraglacial relaxation

Rapid redistribution of paraglacial sediment has been described in the early Holocene (Chapter 1), with, probably, a reduction in both the magnitude and frequency of formative events through time. A small number of slope failures and their aftermath in Northern Scotland have been interpreted as continuing mid-Holocene adjustment, with subsequent stability due to paraglacial sediment exhaustion (Chapter 2). Until recently, rates of sediment cycling which exceed those expected from Holocene weathering, and interpreted as the consequence of paraglacial sediment redistribution, have only been confirmed in young, active mountain belts such as the Canadian Rockies (Church & Slaymaker 1989, and New Zealand (Church & Ryder 1972). These are very different environments from the old mountain landscapes of the less recently glaciated Scottish Highlands, where relatively rapid Holocene weathering has been identified (Chapter 1, Section 1.5.5). Confirmation of high weathering rates is found in Loch Etive, SW Highlands (Howe *et al* 2001). Although rates of slope erosion and lake sedimentation do not have a 1:1 relationship because of interim storage, sedimentation rates in Loch Etive, a deep sea loch in the SW Highlands surrounded by steep slopes comparable to those in Gleann Lichd, indicate very high activity since the end of the Younger Dryas Stadial (Howe *et al* 2001). Average sedimentation rates of 0.7cm a^{-1} in Loch Etive are at least in part attributable to Holocene processes such as deposition from rivers with short, steep catchments. For comparison, the rate of Holocene sedimentation in the Humber estuary has been calculated as 0.1cm a^{-1} (Andrews *et al* 2000). If inferences reported in Chapters 4 - 7 about strong variations in sediment transport rates on slopes in the Northern Highlands are reliable, the high average sedimentation rate in Loch Etive over the last 10ka may reflect large variations in the production and availability of mobile sediment during the Holocene in steep north-western catchments.

7.2.3 Landscape rejuvenation

In Northern Scotland, evidence for mid and late Holocene rejuvenation of both slope and floodplain activity, after a period of climate recognised as warmer and drier than today (Harrison *et al* 1997, Lamb 1977, Yu & Harrison 1995), and followed by several millennia of stability, is not obviously accounted for by continuing paraglacial relaxation. As far as can be established from the still fragmentary data, the frequency of destabilising events appears to have increased, not decreased, with time. In addition, the source of debris in lower slope landforms is, in some cases, demonstrably not

paraglacial. The origin of fan debris and palaeogully fills in bedrock previously stable for several millennia, points to intra-Holocene drivers of instability.

The regional picture of slope evolution is the sum of many local slope histories. No systematic temporal variation or pattern of events over a period of several millennia will arise if local histories arise from random environmental perturbations, and/or to threshold-exceeding events due to intrinsic changes in local slope systems, affecting pre-existing paraglacial sediment prone to failure. But a random model is at odds with field observations. If the pattern described in Section 7.1 is real, other forcing factors, need to be considered. The most likely external triggers are human impact, and climate variability, together with Holocene seismicity. All these may act singly, or in combination. Weathering may also vary with climate, progressively sensitising slopes to the impacts of other environmental triggers.

7.2.4 Anthropogenic forcing

The possibility of an anthropogenic trigger was excluded as far as possible through site selection (Chapter 3, Section 3.1.1). However, soil failures associated with charcoal, may be associated with human activity. Examples occur in the Allt Innis na Làrach fan Strathfarrar (Chapter 4, Section 4.4.1), and the Allt Crumaidh fan in Glen Cannich (Chapter 4, Section 4.2.2.3), which was accompanied by heather burning and erosion of a deep podzol profile. No evidence is available to disprove the anthropogenic origin of the charcoal in the small fan in upper Gleann Lichd, or on Creagan a' Chaorainn. However, the origins of the fan debris in the former case, and the temporal distribution of the charcoal in the latter, poorly support any such argument.

Increased rock-debris production in the late Holocene from sources protected from human impact, and for fan aggradation in the absence of any indicators of human activity, has been established in one study (Ballantyne & Whittington 1999). In this study, as in previous ones (Chapter 2, section 2.2.3), evidence for an anthropogenic trigger for landform evolution is sparse.

7.2.5 Humidity changes and slope failures

Several lines of evidence, both negative and positive, from Chapters 4 - 6 are consistent with increased humidity, rather than vegetation burning (anthropogenic or not) as the dominant influence on the timing of slope failures. The incidence of macroscopic charcoal on Creagan à Chaorainn is unexplained, but did not increase in parallel with known Bronze and Iron Age occupation of the area (Chapter 5, Table 5.7). Secondly, the evidence that burning was followed by slope instability is at best inconclusive (Chapter 5, Table 5.14). Thirdly, soil failures in Strathfarrar and Glen Cannich occurred at a time of widespread debris flows and slopewash, together with possible large rises in river level, and evidence for slope peat erosion. These features can be set alongside other evidence for high water levels in the latter part of the Holocene: poor peat humification; fining-up fluvial deposits onlapping slope debris on lower slopes; and extended transport of spreads of mineral sediment to low-lying areas. There is consequently no need to invoke anthropogenic activity as a critical factor in late Holocene events at the localities examined, and a limited basis for doing so.

7.2.6 Changes in precipitation

The proposition that the timing of mountain slope evolution in the Northern Highlands may have been, at least in part, triggered by changes in hydrological conditions and variable water tables, can be assessed against independent datasets which provide information about the palaeoclimate - specifically precipitation. They include palaeobotanical data, palaeohydrological data available from peat humification studies, and lake level reconstruction (Chapter 2, Section 2.2.2.5).

Changes in precipitation are difficult to infer directly from the stratigraphic record. However, responses to increased humidity inferred in this study have been linked to increases in precipitation in several ways. Sites were selected so that gullies were directly fed by high altitude, rocky areas, which were therefore probably little affected by changes in vegetation. At the same time, water and sediment sources were linked, with minimal interim storage, to lower slope sediment sinks. Fluvial events, related to exceptional discharge and exceptionally high water tables, were also logged.

A complication anticipated in resolving the existence of any regional response to climate change in Northern Scotland is the marked precipitation gradient from east to west, which may itself have been subject to secular variation (Chapter 1). As a result,

precipitation-driven landscape response might be expected to decline from west to east. Data from catchments as widely spread as Strathdearn in the Grampian Highlands area to the east, to Glen Lichd and Glen Shiel on the west coast, and from those listed in Table 7.1 overleaf, reveal no such bias. The table collates dated geomorphic and hydrological events in Northern Scotland, from which precipitation increases have been inferred.

	SITE	W/E/C[1]	EVENT	BP(^{14}C)	Cal BP	AUTHORS
1.	Cairngorms[2]	E	'pluvial'	7350±90	8120	Dubois & Ferguson 1985
2.	Cairngorms[2]	E	'pluvial'	6210±120-5790±80	7050-6600	Dubois &Ferguson 1985
3.	N. Skye	W	debris flow renewal	5055±50	5785	Hinchcliffe et al 1998
4.	NW Scotland	W	wet shift in peat (3 sites)		5120-5070	Anderson 1998
5.	Glen Etive*[3]	W	fan aggradation		5120	Brazier et al 1988
6.	Cairngorms	E	high level solifluction	4480 ±300	5090	Sugden 1971
7.	Cairngorms[2]	E	'pluvial'(very intense)	4480 ±135		Dubois &Ferguson 1985
8.	Beinn Dearg	C	reduced peat humification	4200±230-3940±60	4725-4370	Dubois &Ferguson 1985
9.	Gleann Lichd	W	river channel avulsion	4200-3940	4715-4350	Binney 1997
10.	Creagan a' Chaorainn	C	9-fold increase in debris flow volumes	4000±45	c. 4500	This study
11.	N W Scotland	W	wet shift in peat (3 sites)	after 3850±45	after 4200	This study
12.	Ross-shire/Sutherland	W/E	reduced peat humification		4020-3630	Anderson et al 1998
13.	Glen Feshie[3]	E	fan building		3900	Anderson et al 1998
14.	Cairngorms[2]	E	'pluvial'	3620±50	3870	Robertson-Rintoul 1986
15.	Achany Glen	E	lake level rise	3310±60	3540	Dubois & Ferguson 1985
16.	N W Scotland	W	wet shift in peat (3 sites)	3250±60	3480	Smith 1998
17.	Beinn Dearg	C	decline in peat humification		3340-3270	Anderson 1998
18.	Grampian Highlands	E	floods & net floodplain incision		2750	Binney 1997
19.	Scotland		more frequent of debris flows		2700-2400	Ballantyne & Whittington 1999
20.	Creagan a' Chaorainn	C	abrupt decline in peat humification	2455±50	2700-1700	Curry 1999
21.	Gleann Einich	E	decline in peat humification	c.2500	2535	This study
22.	Beinn Dearg	C	decline in peat humification	c.2500	c. 2560	Binney 1997
23.	Skye	W	debris flow renewal		c. 2560	Binney 1997
24.	Glen Feshie[4]	E	resumption of debris flows	2225±40	2230	Hinchcliffe et al 1998
25.	N Skye*[2]	W	resumption of talus accumulation	2020±50	1950	Brazier & Ballantyne 1989
26.	N W Scotland	W	wet shift in peat	1990±70	1920	Innes 1983
27.	N Skye*[2]	W	resumption of talus accumulation		940-800	Anderson 1998
28.	Gleann Lichd	W	overbank flooding initiated	720±40	665	Innes 1983
	Gleann Lichd	W	overbank flooding initiated	515±45	c. 500	This study
29.	Glen Feshie[4]	E	resumption of debris cone aggradation	320±50	400	Brazier & Ballantyne 1989

TABLE 7.1

Dated geomorphic and hydrological events in Northern Scotland from which precipitation increases have been inferred

With the following *caveats*, Table 7.1 provides a starting point for assessing the relationship of slope failures to other palaeoenvironmental data which suggest increases in humidity.

a)   The work of Dubois and Ferguson (1985) on stable isotope proxies for precipitation has been used as a standard against which to correlate other proxies, (e.g. Anderson *et al* 1998, Binney 1997, Haggart & Bridge 1992). However, more recent work (Chapter 2, Section 2.2.2.5) has undermined the reliability of $^{18}O/^{16}O$ ratios as palaeoprecipitation indicators, and hence correlations which use them.

b)   Some dates are based on peat thicknesses of up to 5cm which may span time intervals of one or two hundred years

c)   Carbon translocation in permeable slope sediments can lead to gross contamination and misleading dates (see Chapter 2, Section 2.2.2)

d)   The precision of most available radiocarbon dates does not allow discrimination of events separated by less than about 100 years.

e)   The more subtle moisture signals from peat humification may be obscured in high rainfall areas and/or high moisture sites (Barber *et al* 2000, Blackford 1993, Binney 1997).

f)   Topography, response to local drainage, and the control that thickening peat itself exerts on hydrology, may affect humification (Birks H. J. B. 1996, Charman 1993,

Haggart & Bridge 1992, Korhola 1995, Lowe 1993, Tallis 1965).

g) Progressive weathering makes talus slopes more susceptible to mass movement during rain storms (Hinchcliffe *et al* 1998). As long as the talus is suitably conditioned, debris flows could occur at any climatic stage of the Holocene and cannot therefore, individually, be interpreted as carrying a climatic signal. On the other hand, there may be long intervals of 'weathering time' between debris flows, even if climatic conditions are conducive to failure.

Omitting the very uncertain evidence for 'pluvials' (Dubois & Ferguson 1986), there is little evidence for changing water tables and slope destabilisation before about 5000 cal BP. Records 1, 2, and 7 in Table 7.1 are therefore considered uncertain indicators of palaeohumidity. Reliable evidence for increased humidity begins only at around 5 cal ka BP. Thereafter, only three records suggest increased humidity at around 4.5 cal ka BP, in upper Gleann Lichd, the adjacent west Glen Affric, and on Beinn Dearg, 20 km north of Gleann Chorainn. Between about 4.2 and 3.2 cal ka BP there are records in most centuries. After a gap of about 500 years, a further sequence centred on four records at about 2.5 cal ka BP spans a period lasting from about 2.7 - 1.9 cal ka BP.

There follows a gap of about one thousand years before the final scatter of dates. Late Holocene overbank deposits described from localities in this study were found in catchments from Strathdearn in the east to Gleann Lichd in the west. Several lines of evidence are consistent with destabilisation of slopes at this time being related to increased groundwater flow. Firstly, these horizons mark the inception of frequent flooding. Secondly, the overbank deposits from widely separated catchments across the region vary only in the relative proportions of peat and sand: the association of erosion and flooding is not easily explained as anthropogenic. Nor is it likely that the outcome of anthropogenic impacts would have declined with increasing grazing pressures in the last few centuries, or perhaps been so uniform. Thirdly, present day erosion of blanket peat is highly variable in grazed areas but increases with altitude (Grieve *et al* 1994) - a response more straightforwardly explained by exposure, temperature and precipitation increases, than by human impact. To this can be added the negative argument that, forestry monocultures excepted, there is no unequivocal evidence for human impact on erosion of slopes unsuited to agriculture (Chapter 2, section 2.1.6).

Scottish dates for enhanced solifluction (Table 7.2) provide an additional dataset linked to climatic triggers.

Site	W/E/C[1]	Date (^{14}C BP)	Date (Cal BP)	Authors
Arkle	W	5440±55-3985±50	6245-4460	Mottershead 1978
Cairngorms	E	4480 ±135	5090	Sugden 1971
Cairngorms	E	2680±120	2820	Sugden 1971
Fannich Mts	C	660±70	650	Ballantyne 1986

TABLE 7.2
Dated Holocene solifluction in the Northern Highlands
[1]**west/ east/close to, the precipitation divide**

The wider European record is inconclusive in relating episodes of solifluction to climate cooling as measured by glacier advances, which have complex causes (Berrisford & Matthews 1997, Gudmudsson 1997). Dated records from Scotland are few in number, but with the exception of the first of those in Table 7.2, they coincide with periods of time noted in Table 7.1 as marking enhanced or intensifying wetness and coldness, as expressed in environmental responses at lower levels.

7.2.7 Neoseismicity

The role of Lateglacial and Holocene seismicity as a trigger for slope evolution deserves consideration. Holocene seismicity in Northern Scotland may be rare (Chapter 2, Section 2.2.1), but some evidence from Glen Cannich (Chapter 4, Section 4.5.3) suggests that it has not yet been fully explored. In Chapter 4, Section 4.5.3, and Chapter 6, Section 6.3.1.1, a previously undescribed phenomenon was logged - *in situ* shattered bedrock beneath till. One such example was on a low angle glacial bench, in the eroded wave zone of the Loch Mullardoch reservoir, close to the Allt Coire Eòghainn postglacial fault trace. The other was beneath a till failure in the old (Caledonian) fault trace of upper Glen Lichd. In the wave zone of Loch Mullardoch, all the sediment had been washed out of the wide interstices which cross cut all the rock structures (Fig. 4.28c). In Gleann Lichd, where the barely coherent rock is exposed beneath a recent till failure on an oversteepened slope, wide fractures are filled by till which could have infiltrated either pre-existing or post-depositional spaces, to form a neobreccia. The precarious stability of the rock fragments on a 34-50° slope, and the fact that some fragments have already fallen (Fig. 6.3), suggests that the rock mass previously remained intact only thanks to the retaining blanket of till. Given the ubiquity of fault-controlled valley incision (Johnstone & Mykura 1989) and the association of neoseismicity with old fault traces (Owen *et al* 1993), there is a possibility that fault-

shattered bedrock of this kind is widespread. If the mid and late Holocene has seen processes capable of triggering new till failures, there may be many areas of till-covered, fault-brecciated bedrock, which constitute a reservoir of potential energy for slope evolution. In Gleann Lichd the cause of till failure was observed to be a spring line. Changes in groundwater flow paths on steep slopes are therefore potentially a direct trigger of continuing slope sensitivity.

7.3   INDEPENDENT CLIMATE PROXIES

Three independent sets of data from Northern Scotland permit some comparison of inferences about past climate. These are: the pollen record (Section 7.3.1), lake level data (Section 7.3.2), and the marine record (Section 7.3.3).

7.3.1   The pollen record

Pollen stratigraphy constitutes the most extensive palaeoenvironmental dataset for Northern Scotland (Ch. 2, Fig. 2.1, Section 2.2.2.5). Many vegetation histories are likely to embody a multi-proxy signal, but the palaeobotanical record consistently indicates widespread soil humidity and blanket peat from about 6.0 cal ka BP. Relatively dry conditions, free of persistent strong winds, which were favourable for woodland expansion to the north and west occurred at around 5.0 cal ka BP, while humidity increased at around 4.5 cal ka BP, (4.0 cal ka BP, Anderson 1995) and intensified after about 3.0 cal ka BP. In west Glen Affric (immediately east of Gleann Lichd) there were marked pine pollen declines at 3765±45 (about 4.1 cal ka BP) and 2315±45BP (about 2.3 cal ka BP) with the pollen succession indicating wetter conditions beginning a few hundred years earlier at 4000BP, (4465calBP) (A. Davies, University of Stirling, pers. comm.). At about 4.5 cal ka BP, river channel avulsion in upper Gleann Lichd occurred. Increased water-logging at this time is also reflected in pollen stratigraphy at Loch Maree which lies about 30km NW of Gleann Chorainn (Pennington *et al* 1972).

In Figure 2.3, which sets out the dated Holocene geomorphic record, the sequence of events described in Section 7.1 is less easily discerned. There are concentrations of slopewash, debris flows and cone/fan aggradation at around 2000 cal BP, again after about 700 cal BP, and possibly between about 6.0 and 4.0cal ka BP. But since only firmly dated events are included, and the investigations were methodologically diverse, it is difficult to draw firm conclusions from the comparison.

7.3.2   Lake level changes

Lake level changes in Northern Scotland have so far provided an uncertain record of changes in humidity (Chapter 2, Section 2.2.2.5). However, inferences from this study, added to other work discussed above, suggest that lake level data have further potential to improve understanding of past changes in humidity in Northern Scotland. In addition to the evidence from three fans in Glen Cannich noted in Section 7.1, rises and falls in lake level of 2 - 3m have previously been suggested in Strathfarrar (Chapter 4, Section 4.4), west Glen Affric, and further north in Sutherland (Smith 1996). There, a lake level rise (in a small lochan) at Achany Glen, Sutherland after about 3500 cal BP was described by Smith (1996, Table 7.1). In west Glen Affric, immediately to the east of Gleann Lichd, the sequence in the spit at the west end of Loch Affric suggests a high loch level followed by a significant fall, both undated (Tipping, 1997. pers. comm.). A fourth locality where there is clear evidence for a former higher shore line is Loch a' Chroisg (NH127587), 15km NW of Gleann Chorainn (Fig. 4.4). A fragment of a well-formed horizontal terrace on the north bank lies 1.5m above the contemporary shoreline. The terrace is not degraded, and there are no records of the loch level rising to flood the road (Highland Region Roads Department). The terrace appears to be a palaeoshoreline, formed when the water surface height was maintained for long enough to form a distinct marginal surface, before dropping to present levels. Unfortunately, the current paucity of dates makes it impossible, as yet, to test whether these records of lake level change form a time cluster.

7.3.3   The marine record

A further dataset with which terrestrial events in northern Scotland can be compared is the record of episodes of marine cooling in the north-east Atlantic and North Sea (Table 7.4) and Chapter 2, Section 2.2.2.2.

These data allow preliminary correlation of cooling of the northern oceans and terrestrial landscape evolution. A southwards advance of the Faroes-Iceland Front at 4700 cal ka BP (Bianchi & McCave 1999) matches the timing of increased precipitation and glacier advance in Iceland (Gudmundsson 1997) and of increased debris flow frequency in Scandinavia (Blikra & Nemec 1997). Historical records in Northern Scotland also suggest a correlation between increased precipitation and storminess and the arrival of cold Arctic water (Chapter 2, Section 2.2.2.5).

1.	5900
2.	5100*
3.	4700**
4.	4200
5.	2800
6.	1400
7.	Little Ice Age (no dates given)

Bond *et al* 1997;   *Jiang *et al* 1997; ** Bianchi & McCave 1999

TABLE 7.4
Marine cooling events (cal BP) in the Northeast Atlantic and North Sea

Broadly speaking, the timing of marine cooling in the early 5th millennium, and in the third and second millennia, coincides with the inferred timing of terrestrial increases in humidity and accelerated slope failure. However, since the resolution of terrestrial records does not match those obtainable from marine sediment sequences, and since the North Sea and Atlantic records diverge, no unambiguous inferences can be drawn.

7.3.4   Summary

In conclusion, several independent lines of evidence - palaeobotanical, hydrological, lake level changes - are consistent with the trend of increased slope instability described in Section 7.1 being contemporary with independent evidence for climate deterioration. This does not constitute evidence for cause and effect. However, in the absence of good evidence for alternative forcing factors, a climatic trigger for mid and late Holocene slope evolution in Northern Scotland is a hypothesis which needs to be tested, not least in the light of its importance for understanding the future sensitivity of mountain landscapes to global warming.

7.4   SLOPE SENSITIVITY TO CLIMATE CHANGE IN NW EUROPE

The timing of events and inferred climate changes in other areas of Northwest Europe, provides a context for evaluating a climatic trigger for the broadly defined pattern of landscape evolution in Northern Scotland. Such data must be treated with caution however, as there is no *a priori* reason for supposing that similar climate changes are synchronous over wide areas; indeed the reverse may hold true, in that the diversion of weather systems to produce wetter conditions in one area may result in drier conditions in other, even adjacent, areas. Interpretation of 'synchronous' may critically depend on

the timescale of observation and of the processes themselves (Chapter 3, Section 3.1.1).

Warmer and drier conditions in NW Europe from about 6.0-5.0 ka BP, and wetter, cooler conditions after c. 3.0kaBP, have been inferred by Cheddadi *et al* (1996), Harrison & Digerfeldt (1993), Houghton *et al* (1995), Huntley & Prentice (1988), Kutzbach *et al* (1993), Lamb (1995), Prentice *et al* (1998).

Some authors have inferred palaeoclimate on the basis of changes in debris flow frequency in Western Norway, whose oceanic climate provides a partial analogue for northern Scotland (Chapter 2, Section 2.2.2.3). As (tentatively) in Northern Scotland, 7.5 - 4.4 cal ka BP in Western Norway was a time of infrequent slope movement, but conditions became more conducive to debris flows after about 4.4 cal ka BP (Matthews *et al* 1997). However, a more ambiguous pattern emerges from the work of these authors when their data are considered as maximum (2σ) age ranges. Between about 3.8 and 2.0 cal ka BP they logged high water tables and frequent debris flows, while an overlapping period of almost continuous peat growth, with much reduced debris flow frequency occurred between 2.7 and 1.3 cal ka BP. In a separate study, Blikra & Selvik (1998) concluded that 5.4 cal ka BP saw a clear deterioration of winter climate in Western Norway, with frequent slope instability between 4.4 and 3.3 cal ka BP, a period of winter climate which was 'one of the most severe in the Holocene'.

Long tree ring records from Fennoscandia provide further information on palaeoclimate change. Tree-rings indicate that interannual variability increased during the latter part of the Holocene, 'presumably as a consequence of increased instability of the climatic system', and overall, humidity increased in northern Fennoscandia during the past five thousand years (Eronen *et al* 1999). According to these authors, a period of particularly harsh climatic conditions affected tree growth 2500-2000 years ago. At that time, pine woodland retreated behind its former limits and growth rates were inhibited to a greater extent than during the "Little Ice Age" whose dendrochronological maximum occurred in Fennoscandia between 1560 and 1650 AD (Briffa *et al* 1990), at the same time as an increase in debris flows in Norway (Grove 1972).

There is therefore some agreement between data from Scotland and Western Norway about early mid Holocene favourable climate and slope stability, which began to deteriorate after about 5.0 cal ka BP, with a major perturbation in the middle of the 5th millennium BP. There is also a common theme of further climate deterioration and increased slope instability thereafter, but the timing may have varied within the broad

scope of 'late Holocene'.

The difficulties associated with large compilations of rapid slope mass movements across Europe were noted in Chapter 2, section 2.2.2.3. Further questions over such large, disparate databases, which attempt to reconstruct palaeoclimate on the basis of regional patterns of mass movement, arise from the fact that a multi-proxy stratigraphic record is unlikely to show synchrony of slope response between sites on timescales of the order of tens to hundreds of years (Section 7.1). Changes in the characteristic frequency of geomorphic events may only be distinguishable on a timescale - in Northern Scotland, probably 10^3 years - which is a multiple of the interval between low frequency events. In consequence, comparisons between sites where 'low frequency' and 'high frequency' have not been defined, due to lack of high resolution dating, are likely to produce some spurious results. In addition, since dated sequences contain an unknown number of only locally significant events, the boundaries of inferred climate events as reflected in compilations of rapid mass movements are unlikely to be sharp.

For comparison with data from the north of Scotland Table 7.3 (overleaf) lists rapid mass movement and hydrological changes inferred in south and west Norway, and mountain areas of northern England and north Wales. There are parallels to the pattern in northern Scotland (Table 7.1) up to about 1.9 cal ka BP. The divergence thereafter could be attributed to both the diversity and patchiness of records and to differences in environmental conditions and response.

	Location	Events	Dates/yrsBP	Authors
1	Iceland	glacier advance	5000-4500	Gudmundsson 1997
2	W. Norway	increased winter storms; snow/debris avalanches	4700-4200	Blikra & Nesje 1997
3	Iceland	glacier advance	4200	Gudmundsson 1997
4	W. Norway	increased rockfall avalanches	4000	Blikra & Nesje 1997
5	W. Norway	severe snow/debris avalanches	3900-3100	Blikra & Selvik 1998
6	Alpine S.Norway	debris flows	3840-2370	Matthews et al 1997
7	W. Norway	snow/debris avalanches	3800-3000	Blikra & Nesje 1997
8	Iceland	glacier advance	3000	Gudmundsson 1997
9	W. Norway	increased slopewash & solifluction	3000	Blikra & Nesje 1997
10	W. Norway	rockfall	3000-2800	Blikra & Nesje 1997
11	W Norway	debris flows	2800	Blikra & Nesje 1997
12	Iceland	glacier advance	2000	Gudmundsson 1997
13	W. Norway	snow avalanches	2000-1800	Blikra & Nesje 1997
14	W. Wales	increased wetness and storminess	2000-1300	Lamb 1977
15	W. Norway	rockfall	1800-1200	Blikra & Nesje 1997
16	NW England	debris flows	1770±60	Harvey 1996
17	Alpine S.Norway	debris flows	1570	Matthews et al 1997
18	Iceland	glacier advance	1500-1200	Gudmundsson 1997
19	W Norway	debris flow increase	1400	Blikra & Nesje 1997
20	W.Norway	snow/debris avalanches	1400	Blikra & Nesje 1997
21	W Wales	extreme wetness	1200	Lamb 1977
22	N Pennines(W)	river incision	1200	Macklin et al 1992
23	NW Pennines(W)	fan building	1100-900kBP	Harvey & Renwick 1987
24	NW England	debris flows	940±90	Harvey et al 1981
25	Cumbria(W)	fan building	900BP	Harvey et al 1981
26	NW England	debris flows	890±50	Harvey 1996
27	W Norway	snow/debris avalanches	after 600BP	Blikra & Selvik 1998
28	Alpine S.Norway	debris flows	630	Matthews et al 1997

TABLE 7.3

Mid and late Holocene landscape evolution in mountain areas of sub-arctic NW Europe

for comparison with Northern Scotland

7.4.1 Summary

In summary, in NW Europe, mountain slope stratigraphy, pine woodland distribution and growth rates, and soil conditions, as well as lake levels, and the marine record, all offer some, though not always unambiguous, evidence of both increased precipitation and increased slope activity after about 5.0 cal ka BP, with later intensification after about 3.0 cal ka BP. The mid 5th millennium BP and the mid 3rd millennium BP, as well as the 'Little Ice Age' stand out as times when independent precipitation proxies are closely clustered. Each of the datasets suffers from uncertainties which cumulatively make for sometimes very large error bars. But at these times in particular, with the caveats spelled out in paragraph 1 of Section 7.4 in mind, the burden of evidence linking increased humidity and landform response appears to be substantial.

7.5   MECHANISMS OF SLOPE EVOLUTION

If climate has influenced landscape evolution in mountain areas of Northern Scotland, the question remains of how that influence operated. A fully specified process-response model incorporating interactions of anthropogenic activity, intrinsic change, weathering rates and vegetation cover with climate, would be extremely complex, and probably unrealistic in view of the co-variability of different forcing factors. It would also go beyond the data generated for this thesis. The aim here is therefore a limited one of drawing together evidence for specific, precipitation-related mechanisms in the catchments studied, and identifying gaps in understanding arising from the fact that the majority of studies have previously been done in glaciated areas. Processes associated with slope evolution are discussed first in terms of erosion and deposition, and second, in relation to lag times and/or cumulative effects.

7.5.1   Erosion and deposition

The slope evolution described in previous chapters arises from a combination of processes associated with erosive and stable conditions. During stable periods, podzols developed, gullies filled with peat, and blanket peat spread over the till substrate. Erosional processes include debris flows of till or destabilised bedrock, grain flow, massive till failure and consequent failure of fault-generated *in situ* breccia, truncation of well developed podzol profiles, water erosion of thin sandy soils and fines from drift, with deposition downslope as thin sheets of sand and silt, and inferred *in situ* peat disintegration and sub-surface erosion during leaching events. From floodplain

stratigraphy, exceptionally low water levels (upper Gleann Lichd, small fan, Ch. 6, Section 6.3.2), exceptionally high water levels (Strathfarrar, Glen Cannich, Ch. 4, Sections 4.4.5, 4.5.1 - 4.5.3)), periods of rare, large floods and channel avulsion (Strathfarrar, Strathdearn (Ch 4, Section 4.6.1.1) and upper Gleann Lichd), and a recent period of frequent, moderate flooding (Strathfarrar, Strathdearn, Gleann Lichd), were inferred. Setting aside faulting and possibly fire, the observed landform changes appear to be largely precipitation-driven through establishment of generally higher water tables and/or more frequent extreme rainstorms. An account of changes through time in the sediment:groundwater ratio will therefore reflect, (at least in part), climate-driven landscape evolution. This ratio addresses the aim stated above, in that it neither excludes nor specifies the unknown relative contributions of changing vegetation cover, precipitation and local conditions.

7.5.2   Blanket peat: a geological material

Blanket peat plays an important biogeochemical role in sediment and water cycling. Peat in northern Scotland has conventionally been regarded as climatically controlled ecological system with dates for its initiation generally between 8-5kaBP (Moore *et al* 1984). More recently it has been recognised that, within broad parameters of humidity, blanket bog is a poor climate indicator because, for instance, it is more strongly limited by topographic confines than changes in precipitation, can be encouraged by progressive soil leaching and paludification in the absence of specific humidity shifts, and itself acts to alter the water table (Birks 1996, Charman 1993, Haggart & Bridge 1992, Korhola 1995, Lowe 1993, Tallis 1965). Dates quoted for blanket peat initiation in northern Scotland are listed in Table 7.5 below. While there is some clustering of ages, few studies have specifically sought this information, and climatic control is not apparent.

	Location	Events	Dates/yrsBP	Authors
1	Iceland	glacier advance	5000-4500	Gudmundsson 1997
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Location	Dates/BP	Authors
near Pitlochry	9800-9200BP	Tipping 1995
Allt na Feithe Sheilich	before 9400BP	Birks 1975
Rannoch Moor	9300BP	Haggart& Bridge 1992
	8300BP	
	6500BP	
Cross Lochs, Sutherland	9170BP	Charman 1992
	intensification at 7000-6000 BP	
Skye	continuously since c. 9000BP intensification since c. 4000BP	Birks 1996
widespread	8000-5000BP	Moore *et al* 1984
widespread	7600BP	Tipping 1994
Cairngorms	6090BP and 5230BP	Dubois & Ferguson 1985
widespread	c. 5500 (Cal BP)	Bennett 1994
Loch Sionascaig	c 4500BP	Pennington *et al* 1972
Loch Maree	3800 BP	Birks 1972
locations in Sutherland	after 1500 BP	Pennington *et al* 1972

TABLE 7.5

Blanket bog initiation and expansion dates, Northern Scotland

As yet, little work has been done on run-off production from peatlands (Burt *et al* 1990). Chemical erosion beneath peat is even less well investigated (Fuchsman 1986, Tallis 1999 pers. comm.). As a result, its role as an Earth surface material affecting slope evolution has not been emphasised in the literature. Yet this material covers tens of thousands of hectares of mountain slopes in the Highlands.

The very low vertical hydraulic conductivity of peat (Clymo 1991) means that drainage is concentrated at the base where flow is towards the margins, creating conditions for expansion from nuclei in damp hollows. During peat growth, hydraulic conductivity above the water table decreases by several orders of magnitude, promoting increasingly efficient lateral drainage and marginal peat expansion. In Caithness, a hydrogeological study found that surface water took 10 years to descend through thick peat and begin to infiltrate bedrock (Robins 1990).

Rates of peat growth, as well as its hydrological status, are linked to the availability of mineral sediment. Peat may be truncated by debris, or its transmissivity may be altered by incorporation of sandy slopewash. Lateral and vertical expansion require both damp conditions and a stable substrate conducive to slow accumulation of dead plant material.

The spread of blanket peat on slopes and the growth of thick peat on floodplains both therefore imply an absence of mobile sediment.

Once established, blanket peat causes irreversible changes in local hydrology, (Clymo 1991), and its ability to retain three times its own weight in water alters infiltration patterns and reduces run-off on gentle slopes (Pitty 1982). The peat thus becomes a buffer in interactions between Earth materials, climate and slope stability. If the water balance changes in favour of evaporation, lateral drainage continues, but the raised water table within the bog sinks. If drainage is disturbed, irreversible drying of the peat begins (Clymo 1991). In the Cairngorms a middle to late Holocene rate of blanket peat growth on slopes of 1.4 - 3.4 cm/100 years has been calculated (Pears 1975), comparable to rates calculated on Creagan a' Chaorainn (Chapter 5). With such a slow growth rate, even minor erosion scars may be unable to develop a vegetation cover sufficiently fast to heal the damage (Pears 1975). Thus any scarring may result in a run-away process of erosion (Clymo 1981, Korhola 1995, Pears 1975). Once initiated, slope erosion creates positive feedbacks between groundwater and sediment availability. One such example is the enhanced erosion of peat found in catchments where the water is already charged with sand and silt. The Peatland Team of Scottish Natural Heritage (quoted in Grieve *et al* 1994) identified water containing mineral grains washing over peat as much more erosive than clear water run-off. This may explain why, in the Midland Valley of Scotland, current erosion rates of catchments with eroding blanket peat are greater than those in catchments with mineral soils (Duck & McManus 1990). As with well established parameters for sediment erosion in fluvial systems, the critical threshold for system transformation - the initial breach of the peat cover - may require a higher energy input than is needed to maintain transforming processes - that is, persistent erosion of the till substrate.

In addition to the internal dynamics of peat/water/substrate interactions, large peat bogs which grow beyond their stable topographic confines are intrinsically unstable (Evans 1993, Tallis 1965). But few studies in NW Europe with the exception of Korhola (1994) have specifically investigated such confines, and they have not been mapped in northern Scotland. In such circumstances an intense storm might be enough to trigger major disruption and instigate long-term erosion by overcoming internal thresholds. When peat is incised or eroded, the mineral substrate is exposed so that both peat and sediment are subject to further erosion.

There is some field evidence that the integrity of peat cover can additionally be damaged by two types of chemical weathering intrinsic to the peat system. There is an increase in the dissolution and removal of iron and manganese oxides with depth within a peat column (Fuchsman 1986, Clymo *et al* 1990). Iron-rich mineral flushes seen both in the field and as relict ferromanganese cements (Strathfarrar, Glen Cannich), confirm the long history of that process. The adsorbtion/desorbtion of large metal ions in peat can, in addition, substantially alter pore shapes and sizes and lead to dewatering (Fuchsman 1986). Iron in solution is normally present at a concentration of <1ppm in river and groundwater but these levels can be grossly exceeded in peatlands. In extreme cases, bog iron ores, in which manganese oxide may reach 40%, are formed in middle and high latitude sites, perhaps through groundwater seepages (Tucker 1981). Once leached into groundwater, ferric hydroxides are stabilised in the presence of organic matter (Tucker 1981). In Highland catchments, colloidal peat, which colours stream water a whisky brown, represents an abundant source of OM. An increase in Fe and Mn in lake sediments has also been interpreted as a sign of erosion of waterlogged peat and peaty soils (Pennington *et al* 1972).

Removal of ions in solution means volume loss in the peat column (unquantified by published studies) as well as dewatering. When iron flushes are sufficiently large to colour streams and coat their vegetation and sediment, or form concentrates which precipitate as substantial cemented horizons, sudden volume loss is indicated. Cements interpreted as resulting from Fe-Mn flushing have been logged in fans and laminated lake deposits in Glen Cannich. There is also some field evidence from the Allt Innis na Làrach fan in Strathfarrar that chemical erosion can cause physical disaggregation of peat. This seems likely to make it susceptible to mechanical erosion by increasingly channelled flow within the peat.

A second destabilising chemical process acts on minerals in the substrate rather than on the slope peat itself. Hydrolysis of Al-Si framework silicate minerals such as feldspars, micas and amphiboles, common in metamorphic rocks of the northern Highlands and their derived sediments, results in bleached boulders and pebbles beneath peat. The bleaching is due to chemical weathering removing Ca^{2+}, Mg^{2+}, Na^+, K^+ and SiO_2 ions in solution, leaving a residue of sheet silicates (micas and clay minerals). The operation of this process in the past has been confirmed in a study of interglacial deposits in Strathnairn which runs parallel to, and some 10-15 km north of, Strathdearn. There,

vermiculite was identified as a product of pedogenesis in bleached gravels beneath cryoturbated peat (Bloodsworth 1990). Quantifying the rate of weathering and the volume of the clay product is beyond the scope of this thesis. However, a substrate boundary with a percentage of clays will behave physically, chemically and hydrologically in very different ways from a sandy substrate: permeability and transmissivity will be reduced; anisotropic flow parallel to the slope will increase and act to enhance lateral sub-peat drainage; shear strength may also be reduced; and ion exchange between peat leachates and clay minerals will be significant. It has been estimated that estimated that the presence of 5% by volume of clays was sufficient to lubricate very coarse debris flows on low angle slopes (Rodine & Johnstone 1976).

In Glen Cannich (Chapter 4, Section 4.5.3) where other potential energy sources appear to be absent, these observations suggest an alternative (or addition) to Holocene seismicity as a mechanism for slope destabilisation. On the shores of Loch Mullardoch, on a slope with an angle of only 5-10°, apparently energetic debris flows, enclosing disrupted peat masses, coherent but unlithified segments of cemented laminates, and large boulders, have occurred in areas where there are no streams, springs, or gullies (Fig. 4.29).

Because of its status as a physically chemically and hydrologically active sub-system, blanket peat and its substrates together form an environmental system which may be inherently unstable at times of climate fluctuation, due to the fact that the peat itself can act to lower slope stability thresholds. Erosion of blanket peat in the Scottish uplands is commonly attributed to human impact, through a combination of grazing pressures, heather burning and tree removal (Chapter 2, Section 2.2.3). However, intrinsic mechanisms for peat destabilisation as described above, amplified by climate change, cannot, on this evidence, be discounted. Moreover, the positive feedbacks between peat, slope hydrology and substrate stability mean that peat and its erosion should be considered as elements of a wider system of inter-related slope thresholds and equilibria.

The mechanisms of two further types of slope instability remain to be discussed: fan incision and slope podzol erosion.

7.5.3 Fan incision

Where fan sediment and run-off are sourced in long structure gullies, whose origins are in outcrop on upper slopes, they, rather than human-induced changes in water table on the fan apron, are likely to be the locus of environmental control of fan apron incision. Fan incision some time after aggradation indicates a change in the sediment:groundwater ratio and has been described by Brazier *et al* (1988) as concurrent with anthropogenic activity operating through vegetation clearance and a rise in water table. But incision is not necessarily related to a rise in water table and could be caused by sediment exhaustion irrespective of the volume of run-off. This proposition can be tested against field observations.

The incision of smaller, finer-grained fans, and the lack of incising streams in very coarse fans in Gleann Lichd, Glen Shiel and Strathfarrar (Allt Innis na Làrach) may be systematic. Incision depends on the type of fan sediment and the sediment:groundwater ratio (Chapter 6, Section 6.4.1.3). Where fan aggradation has largely depended on rapid response mechanisms, consisting of flows hypersaturated with fines washed from till, and/or failure of oversaturated soils during periods of exceptionally high water table, sediment starvation, or small-scale, incremental processes in balance with the average transporting power of the fan stream, the norm will be channelled, water-dominated flow, which may occasionally be debris dominated (e.g. the small fan in upper Gleann Lichd). But massive flows of rock-derived debris are not the product of channelled flow. So larger, coarser fans, with no distal fining (e.g. the large fan in upper Gleann Lichd and the Allt Carnach fan in Glen Shiel, Fig. 6.21) do not have their origins in stream flow. In such fans, reverse grading, flow envelopes of non-fluidised debris, and flow stacking, ensure high surface and subsurface permeability. As a result, incision is not the norm. This explanation predicts that fan extension by water-dominated processes will initiate incision (e.g. the Allt Crumaidh and Allt Innis na Làrach fans). Incision of the Allt Carnach fan in Glen Shiel appears not to fit this model. However, where the fan source area is unusually large (as in this case), exceptional flow is apparently, on rare occasions, capable of disrupting even such a hydrologically robust system.

7.5.4 Slope podzol erosion

Slope podzol erosion in Northern Scotland is not yet well understood. Physical modelling (Brooks *et al* 1993, Brooks 1997) suggested that as podzols mature and differentiate into hydrologically distinct layers, there will be a decline in their stability relative to immature, more homogeneously drained precursors. The implication is that mature slope podzols have an inbuilt, and decreasing, resistance to erosion with age, even in conditions generally favourable to slope stability. The mechanism of decreasing shear resistance was described by these authors as the channelling of groundwater through more permeable layers with preferential flow through the illuviated, iron-rich B horizon. However, field observations on natural slopes have shown shallow groundwater flow in saturated conditions to be concentrated in near surface layers, notably the A horizon (Kirkby 1978). This observation better matches the truncation pattern seen in podzols in Sections GCh 7a & b (Figs. 5.12,, 5.13) where B-E horizon pairs are preserved, but A horizons have frequently been partly or wholly removed. Repeated podzol profile truncation on Creagan a' Chaorainn occurred on slopes of much lower angle than those derived from models by Brooks *et al* (1993) (10-15° as opposed to 34-41°).

Some freely draining soils in Scotland are reportedly underlain by indurated, low permeability horizons - fragipans - capable of generating a perched water table which lowers stability thresholds (Brooks *et al* 1993, Brooks 1997). An indurated basal horizon resembling fragipan was identified on Creagan a' Chaorainn (Chapter 5, Section 5.3.1.1). Any resultant perched water table would have enhanced backflow from shallow bedrock, and repeated podzol erosion may not have required exceptional hydrological perturbations. That is consistent with the repeated renewal of podzol development in leaching, but not generally destabilising conditions.

7.5.5 Summary

Mechanisms for climate-influenced slope activity therefore include:

- cumulative processes, especially bedrock, till and soil weathering
- long lag time processes particularly accumulation of bedrock-derived debris
- short lag time responses such as sediment failures in response to extreme events.

Chapter 8 concludes this thesis by reviewing models for Holocene landscape evolution in Northern Scotland, and suggesting how competing hypotheses can be further tested.

CHAPTER 8: MODELS OF MID AND LATE HOLOCENE SLOPE EVOLUTION IN NORTHERN SCOTLAND: CONCLUSIONS AND FURTHER WORK

8.0   EXISTING MODELS OF SLOPE EVOLUTION

Use of the paraglacial relaxation model to explain mid or late Holocene landform evolution in northern Scotland has partly relied on a scarcity of evidence capable of supporting alternative hypotheses (Ballantyne 1986, 1991, 1993, Innes 1997). For the early part of the Holocene, paraglacial sediment cycling successfully predicts a rapidly declining rate of drift redistribution, followed by sediment exhaustion. Thereafter, available evidence has principally supported random local response, with an anthropogenic overprint, as drivers of change. But although enhanced slopewash has been associated with evidence for human activity in a few studies, a primary, or sole, role for human impact in creating episodes of enhanced slope wash, remains uncertain. In addition, an anthropogenic trigger for rapid mass movement capable of transporting large volumes of sediment, has not been demonstrated.

A revision of this model is supported by the observation that patterns of mid and late Holocene activity identified at a number of sites across the region, may be non-random. Tentatively - due to limited evidence - that pattern is of several millennia of rare disturbances, followed by groundwater-generated perturbations which affected floodplains more immediately than sediment-starved slopes. Although thin mineral soils and friable till blankets on slopes were rapidly dislodged, longer-term processes such as granular disaggregation of bedrock, may have taken some time to produce large stores of potentially mobile sediment.

After the initial large disturbances, conditions conducive to slope destabilisation persisted. It remains unclear whether the inferred late Holocene 'intensification' of slope mobility is more strongly linked to frequent large precipitation events, or to generally higher annual water tables. Similarities between dated and undated sequences in this study suggest that a programme to test whether these perturbations are identifiable in a wide range of settings across the region at around the middle of the 5th millennium BP, after about 2.5 cal ka BP, and again, in a more muted form, during the latest period of severe climate, commonly referred to as the 'Little Ice Age', would be worthwhile. There are also indications that the conditions affecting slopes in Northern Scotland were, at roughly the same time, active in the wider region of Northwest Europe

(Chapter 7, Section 7.4).

It has not been possible to separate the contributions of intrinsic local response, random intense storms and human-induced changes in vegetation cover to landscape evolution from those of a climate trigger. Since they are interlinked variables, it may, in principle, be impossible to do so. However, the coincidences of timing of slope instability with palaeobotanical and stratigraphic evidence for increased precipitation, raised water tables on slopes and floodplains, with slope instability in other areas of NW Europe, and with southward incursions of cold Arctic waters (Chapter 7, Section 7.2) in the mid to late Holocene, emphasise the need for a model of landscape evolution which encompasses responses to climate change. Such a model is not in conflict with the concept of paraglacial relaxation. Instead, it extends the scope of landscape evolution beyond the final outcome of paraglacial sediment exhaustion - relict landforms, subsequently modified by intrinsic response and/or human impact. Those relict landforms have been shown to be susceptible to modification by intra-Holocene processes which are directly or indirectly driven by energy inputs independent of the glacial/interglacial transition, including precipitation-dependent groundwater flow.

In the light of temporal association with independent evidence for climate change, and in the absence of an alternative satisfactory explanation, a testable hypothesis is that natural slopes in mountain areas of Northern Scotland have been both directly and indirectly sensitive to low amplitude Holocene climate change on a number of in-phase and out-of-phase timescales. Thus, slope failures and related lower slope deposition might exhibit

- high frequency in the early Holocene due to paraglacial relaxation

- low frequency in the mid-Holocene due to sediment exhaustion and/or stabilisation

- increased frequency in the mid to late Holocene due to the action of increased precipitation and/or extreme rainstorms on stores of weathered material, and on spring-generated till failures exposing fractured bedrock

The use of slope failures, or compilations of slope failures, as palaeoclimate proxies in this non-glaciated area is problematic, since a) direct triggering mechanisms for distant events are likely to be both disparate, interdependent (and in most individual cases impossible to establish), and b) lag times between some kinds of slope response and the initiation of triggering conditions are unknown. Nevertheless, long-sequence studies of

landform evolution offer insights into persistent changes in the frequency of landscape-altering events, and such changes contain a multi-causal signal of environmental change.

8.1   SUMMARY OF CONCLUSIONS

Northern Scotland has a climate of extreme oceanicity, and lies at the crossroads of several oceanic and atmospheric systems. It has been deglaciated for 11,500 -13,500 years, providing a long time span during which a distinction can be made between paraglacial-driven processes and those whose triggers are independent of postglacial relaxation. It is therefore a critical site for improving understanding of landscape response to climate change.

1. Paraglacial adjustment in Northern Scotland may have been largely complete by about 7.5 cal ka BP, some four millennia after the end of the Younger Dryas Stadial.

2. Climate has influenced mid and late Holocene mountain landscape evolution in northern Scotland although it is unlikely to have operated independently of other factors such as weathering rates, human impact and changes in vegetation cover.

3. Despite the strong east-west precipitation gradient there is no evidence that climate-related landscape evolution was restricted to the wetter, western area.

4. The geomorphic impacts of the 'Little Ice Age' may have been more limited than those associated with earlier climate perturbations particularly in the mid 5th and mid 3rd millennia BP.

5. Slope sensitivity to precipitation changes is an evolutionary function of its geological and glacial history.

6. Changes in hydrological conditions which mark periods of enhanced slope activity can be inferred, though not always precisely specified, from stratigraphy, slope landform morphology, and sedimentology.

7. Many North Scottish debris fans and 'stable' slope profiles such as those in Upper Gleann Lichd and Gleann Chorainn may date from the mid or late Holocene, rather than being stable relics of paraglacial relaxation dating from the first few millennia after deglaciation.

8. Blanket peat is an agent in physical, chemical and hydrological slope system feedbacks, both positive and negative. In terms of landscape evolution in Northern

Scotland, it should be considered as a geological material, rather than simply a botanical system.

9. Mechanisms of slope landform rejuvenation described here have implications for future slope evolution, since the driving processes and sensitivities are not defunct. Landscapes in the Northern Highlands may therefore be sensitive to future climate change which could interact with the human impact now evident on an unprecedented scale.

10. Both large disparate sets of records of slope failures aimed at identifying palaeoclimatic 'episodes', and the extrapolation of quantitative, short-term process-response studies to palaeoevents, are potentially beset by very large errors.

8.2 FURTHER WORK

This study has raised many questions as well as proposing testable hypotheses. In particular, the typical nature and timing of the pattern of landscape evolution detected needs to be tested by further dating in the undated catchments and elsewhere. Specifically, the question arises of whether large, but infrequent, perturbations in the mid 5th millennium BP, followed by intensifying slope and floodplain remodelling after about 3 cal ka BP, due to increasingly frequent, though not necessarily very high magnitude storms, is indeed a regional effect. A second question is whether the slope sensitivity described, and the inferred mechanisms, apply throughout the region. There is a possibility that 7.5 cal ka BP marks an environmental cusp in northern Scotland as well as western Norway. If so, common causal mechanisms might be inferred. But the number of dated landform changes in Scotland is still too small for such generalisation, and the existence of older buried horizons has not been excluded.

Further work to extend understanding of possible lake level changes in a trio of east-west valleys (Glen Cannich, Glen Affric and Strathfarrar) would underpin testing of the proposition that a) the mid or late Holocene saw regional cluster of lake level changes and that b) slope activity occurred in parallel with, and in response to, increased humidity. Such dates would make a useful contribution to the UK and European lake level database.

Sites already described in Strathfarrar, Gleann Chorainn and Glen Cannich, contain datable material whose age would clarify the timing of slope failure, river channel avulsion (whose timing has been dated only in Upper Gleann Lichd), pine growth and

retreat. Undated pine buried beneath peat was found in each of the above-mentioned catchments, as well as in Strathdearn. There are indications that the disappearance of these trees has more than local significance, and dates are needed to test that hypothesis.

In more recent times, the region-wide occurrence of late Holocene overbank deposits may be climate driven, but the basal date of 500 cal BP is known for only one of four sequences described and requires substantiation.

The chemical, physical and hydrological role of blanket peat in stabilising and destabilising slopes covered by sandy till, is an issue whose relevance is apparent in the contemporary context of increasing human impact and changing climate.

In conclusion, both dated and undated sites investigated for this thesis have the capacity to yield further information capable of testing the proposed model of partially climate driven landscape evolution.

APPENDIX 1

SEDIMENT LOGS: FIELD DESCRIPTIONS

Descriptions follow in the order that sites appear in the text. All boundaries between horizons are sharp unless otherwise specified.

Chapter 4

Allt Innis na Làrach fan (Section 4.4., Figs. 4.17 - 4.19)

Depth/m Section SF1

0-0.2	Immature, sandy soil.
0.2-0.3	Black, humified peat.
0.3-0.6	Black peat with thin, sandy flushes.
0.6-0.85	Black humified peat with wood fragments.
0.85-1.4	Black peat with thin sandy flushes and wood fragments throughout.
1.4-	Water-worn bedrock.

Depth/m Section SF2

0-0.2	Immature, sandy soil.
0.2- 0.5	Iron-mottled sand with 1-2cm impersistent, peaty horizons.
0.5-0.62	Fibrous peat with thin sandy flushes.
0.62-0.87	Peaty sand grading into:
0.87-1.3	Sandy fibrous peat with wood fragments.
1.3 -	Water-worn bedrock.

Section SF3

0-0.2	Unlayered, iron-stained, sandy soil.
0.2-0.65	Iron-mottled sand with 1-2cm impersistent, peaty horizons.
0.65-0.7	Grey-buff, plastic silt immediately overlying, and sheared and folded with:
0.7-0.8	Distorted and discontinuous charcoal layer 2-5cm thick within
0.7-1.65	Iron-stained, medium to coarse, structureles sand with two, single-clast layers of cobbles
1.65-	clast-supported, rounded cobbles

Section SF4

0-0.25	Orange-brown, sandy, slightly organic soil.
0.25-0.95	Iron-mottled sand with 0.5-2cm impersistent, peaty horizons.
0.95-1.13	Black peat; wood near top poorly preserved;
1.13-1.68	Brown sand with disseminated organic matter and small decomposed plant fragments.
1.68 -	Pebbly sand.

Section SF4a   (gouge augur hole)

0-0.75	Dark brown, fibrous peat.
0.75-1.25	No sample.
1.25-1.35	Black peat with coarse sand.
1.35-1.42	Black humified peat.
1.42-1.6	Black sandy peat
1.6-1.8	Black peat with lenses of sand and sandy peat.
1.8-1.93	Silt with twigs at the base and irregular particles of eroded peat.
1.93-2.25	Unstructured (?) homogenised peaty organic mud; twigs in central portion
2.25 -	No sample.

Allt Innis na Làrach fan contd.

Section SF5

0-0.22	Orange-brown, sandy, slightly organic soil.
0.2-0.45	Iron-mottled sand with 0.5-2cm impersistent, peaty horizons.
0.45-0.5	Pebbly, cobbly sand.
0.5-0.8	Iron-mottled sand with 0.5-2cm impersistent, peaty horizons.
0.8-0.85	Pebbly, cobbly sand
0.85-1.2	Iron-mottled sand with 0.5-2cm impersistent, peaty horizons.
1.2-1.35	Iron-rich, pebbly, cobbly sand; some clast support.
1.35-2.15	Dark brown fibrous peat.
2.15- 2.3	Pebbles with a sandy matrix.
2.3-?	Amorphous peat, liquefying on sampling.

Section SF5a   (gouge augur hole)

0-0.1	Orange-brown, sandy, slightly organic soil.
0.1-1.1	poorly sorted, buff, coarse sand with Silt.
1.1-1.2	Coarse sand + granules & pebbles.
1.2-1.4	Black, amorphous peat with sand
1.4-1.45	Coarse sand + granules & pebbles.
1.45-1.7	Fibrous, sandy peat.
1.7-1.9	Peaty sand with twigs at base.
1.9-?	Partial sample of disintegrating peat.

Braulen Fan (Section 4.4.4, Fig. 4.21)

0-0.5	Buff, matrix-rich, coarse, angular, unsorted debris.
0.5-0.7	Black humified peat.
0.7-0.95	Clast-supported, sub-angular cobbles
0.95-1.4	Black, humified peat with *Calluna* roots
1.4-1.7	*In situ Pinus sylvestris*
1.7-?	not seen

Liatrie Fan (Section 4.5.1, Fig. 4.23)

0-2.2m	Buff, matrix-rich, coarse, angular, unsorted debris. Crudely cemented, horizontal layers near base, 5 & 20 cm thick respectively. Cement consists of blackish-red Fe-Mn oxides + variable amounts of eroded peat.
2.2-?	Soft, grey-brown organic silt and mud with plant rootlets.

Allt Coire Eòghainn Fan (Section 4.5.2 Fig. 4.25

0-0.8	Winnowed boulder deposit, matching lithology of till walls of source gully.
0.8-0.85	Black organic mud.
0.85-1.05	Coarse sand with lenses of water-lain twigs and pebbles.
1.05-1.5	Sandy, black humified peat with twigs and root fragments.

Stratigraphy unseen, but giving way down-fan to:
(minimum depths; measured thicknesses)

1.5-3.6	clast supported cobbles and boulders
3.6-3.9	black, humified peat with wood fragments

Stratigraphy unseen, but giving way down-fan to:
(minimum depths; measured thicknesses)

3.9-4.65	grey-brown organic, silty, mud with rootlets and dark brown organic patches of eroded peaty silt.

Allt Crumaidh Fan (part) (Section 4.5.3, Fig. 4.26)

0-0.15	Unlayered sandy soil
0.15-0.95	Intensely iron-stained, unsorted, silty sandy, unstratified deposit with randomly scattered *Calluna* roots, sparse charcoal, and rounded peat masses with randomly oriented 'tails'. Many roots and peat masses preserved as carbon films, even within the uneroded body of sediment.
0.95-?	Clast-supported, rounded cobbles base not seen.

Chapter 5

Creagan a' Chaorainn (Figs. 5.9 - 5.15)

Section GCh4

0-0.08	Thin peaty soil with sparse boulders on surface.
0.08-1.18	Brown fibrous peat with identifiable plant remains in lower section.
1.18-1.3	Very poorly humified light brown peat; preserved plant remains (grasses, *Sphagnum*)
1.3-1.95	Black humified peat with numerous poorly preserved woody fragments.
1.95 -2.0	Silty sand with charcoal and twigs.
2.0->2.3	Barely clast-supported debris with sandy silty matrix. Base not seen.

Section GCh5

0-0.0.1	Poorly differentiated soil developed in till.
0.1-0.7	Very angular debris, sandy matrix
0.7-0.75	Thin charcoal and sub-horizontal thin tree roots (*in situ*?).
0.75-1.57	Black humified peat.
1.57->1.85	Clast-supported debris with coarse sandy matrix, grading up to pebbly, coarse sand.

Section GCh6

0-0.1	Peaty soil.
0.1-0.4	Thin (1-3cm) peats alternating with sandy Pebbly horizons. Palaeosurface angle same as today's (20°).
0.4-1.4	Matrix-rich debris with downslope-pointing clasts.
1.4	Charcoal on surface dipping at 20°.
1.4->2.0	Matrix-rich debris.

Section GCh8

0-0.1	Immature sandy soil.
0.1-0.55	Very angular debris, sandy matrix
0.55-0.65	Black humified peat.
0.65-0.95	Very angular debris, sandy matrix
0.95-1.12	Black humified peat.
1.12-1.35	Very angular debris, sandy matrix
1.35-1.4	Black humified peat.
1.4	Charcoal layer
1.4->1.8	Matrix-rich debris; some water sorting.

Section GCh7a (All horizons gradational and developed in coarse, angular, matrix-rich till; downslope angle similar to modern surface)

upslope		mid slope		lower slope	
0-0.15	Peaty, sandy soil.	0-0.15	Peaty sandy soil	0-0.2	Peaty sandy soil
0-0.25	Poorly defined E&B horizons of podzol.	0.15-2.25	Matrix-rich debris; clasts point downhill; base not seen. Indistinct flow boundaries.	0.2-1.5	Matrix-rich debris; clasts point downhill;
0.25-0.65	A-E-B podzol horizons.			1.5-1.8	A-E-B podzol horizons; boundaries fade upslope.
0.65-1.3	A-E-B podzol horizons.				
1.3->1.6	Matrix-rich debris.			18	Charcoal; woody, 1-25mm; extends upslope for 1m, along adjacent wall of modern gully, and downslope on opposing side of modern gully.

Lower slope contd.

1.8-2.3	E-B podzol horizons, boundary marked by faint colour/texture change in debris. Truncated about 1m upslope against another, less laterally persistent A-E-B sequence. Repeated fragments of single podzol? Fragments of different podzols?

Creagan a' Chaorainn contd.

Section GCh7b (All horizons gradational unless specified, and developed in till as for GCh7a)

0-0.2   Sandy soil.
0.2-0.65 A-E-B podzol horizons.
0.65-1.15 E-B podzol horizons with trace of A.
1.15-1.6 E-B podzol horizons with trace of A.
1.6-1.7   Friable, degraded black peat; erosional upper surface; contains macroscopic charcoal.
1.7->2.3 Till
Podzols stacked or repeated by shearing?

Section GCh9

0-0.15   Immature sandy organic soil.
0.15-0.3   Very angular debris, sandy matrix.
0.3-0.4   Black humified peat.
0.4-0.55 Very angular debris; sandy matrix.
0.55-0.67 Black humified peat.
0.67-0.76 Very angular debris; sandy matrix.
0.76-0.8   Black humified peat.
0.8-1.0   Very angular debris; sandy matrix.
1.0-1.06   Black humified peat.
1.06-1.13 Very angular debris; sandy matrix.
1.13-1.36 Black humified peat.
1.36-1.52 Very angular debris; sandy matrix.
1.52-1.6   Black humified peat
1.6-1.67 Very angular debris; sandy matrix.
1.67-1.9 Black humified peat.
1.9-2.0   Single layer of very angular debris.
2.0- 2.3   Black humified peat
2.3-2.36 Single layer of very angular debris.
2.36-2.55 Black humified peat.
2.55-2.6 Silt with coarse sand; residual
           layering and charcoal; burnt appearance.
2.5-2.8   *In situ* tree stump; surface burned
2.7-?   Matrix-rich till.

Section GCh10

0-0.2   Sandy peat.
0.2-0.65 Black humified peat; rare twigs.
0.65-72 Very angular debris,; sandy matrix.
0.72-1.1 Black humified peat; twigs and some
           fibrous areas.
1.1-1.3   Angular debris with sandy matrix
           containing irregular patches of OM.
           Erosional contact with
1.3-1.8   Black humified peat with decomposed
           wood fragments.
1.8->2.0 Matrix-rich till; some water sorting.

Chapter 6

Upper Gleann Lichd (Figs. 6.9 -6.12)

Section GL1 (small fan)

0-0.1	Sandy soil with fine root mat.
0-0.2	Barely matrix-supported angular debris; reverse grading.
0.4-0.46	Lens of unsorted angular silt to coarse sand; iron-stained.
0.46-0.8	Barely matrix-supported angular debris; reverse grading.
0.8-1.15	Unsorted, Fe-stained, angular silt to coarse sand.
1.15-1.5	Barely matrix-supported angular debris.
1.5	Charcoal layer
1.5-1-8	Shallow channel filled with unsorted, Fe-stained, angular silt to coarse sand. Erosional boundary with
1.6-2.0	Barely matrix-supported angular debris.
2.0-2.3	Fe-stained, angular silt to coarse sand containing 2cm thick lens of dispersed charcoal.
2.3->3.5	Clast supported rounded gravels.

Section GL2 (large fan)

0-0.0.1	Immature sandy soil.
0.1-0.3	Barely matrix-supported angular debris; reverse grading.
0.3-0.1.2	Black, slightly fibrous peat with wood fragments.
1.2-1.32	Matrix-supported gravels.
1.32-1.5	Black humiffed peat with thin flushes and small lenses of sand, and twigs.
1.5-1.76	Amorphous peat/sand mix; sandy lenses and wood fragments. erosional boundary with:
1.76-1.85	Black humified peat; wood fragments at base.
1.85-?	Clast-supported, rounded gravels.

Section GL3 (large fan)

0-0.3	Barely matrix-supported angular debris; reverse grading; minor soil at surface.
0.3-0.55	Barely matrix-supported angular debris; reverse grading.
0.55-0.9	Black humified peat with wood remains throughout: some *Betula* bark.
0.9-1.05	Black, sandy peat
1.05-1.25	Matrix-supported gravels; erosional boundary with:
1.1-1.3	Black humified peat.
1.1-1.4	Matrix-supported gravels; some down-valley imbrication.
1.6-1.85	Matrix-supported gravels with eroded peat masses and wood.
1.85-2.08	Black humified peat with wood fragments throughout.
2.08->2.5	Rounded, clast-supported cobbles.

Section GL4 (large fan)

0-0.12	Immature sandy soil
0.12-0.23	Barely matrix-supported angular debris;
0.23-0.82	Black humified peat; wood at base.
0.82-1.35	Matrix-supported gravels.
1.35-2.05	Black plastic mud and re-sedimented peat; no layering; randomly oriented wood fragments throughout.
2.05->2.2	Pebbly sand.

Section GL5 (floodplain)

0-0.1	Immature sandy soil.
0.1-0.97	Laminated overbank sands with Thin, irregular peaty stringers.
0.97-1.5	Black humified peat; horizontal root fragments oriented downstream, at base of peat.
1.5->2.0	Rounded, clast-supported cobbles.

Glen Shiel (Section 6.4.1.3, Fig. 6.20)

Allt a' Mhuing fan (All horizons dip 10-15°)

0-1.0	Parallel-bedded sands and gravels.
1.0->1.2	Coarse, matrix-supported debris
	Stratigraphy unseen, but giving way down-fan to:
	(estimated depths; measured thicknesses)
2.0-3.0	Coarse, matrix-supported debris; some
	downward pointing clasts.
3.0-3.5	Black humified peat with wood fragments
	and some fibrous areas.
3.5->3.8	Grey/buff compacted silt with dispersed, angular
	pebbles and coarse sand; small pebbles very decomposed;
	small plant fragments randomly dispersed throughout; no
	layering or oriented particles.
	Stratigraphy unseen, but giving way down-fan to:
	(estimated depths; measured thicknesses)
5.0-5.3	Peaty soil overlain by coarse, debris.
5.3-5.6	Black humified peat.
5.3-5.4	Clast-supported cobbles.
5.4-5.9	Coarse, unstratified sand.
5.9-6.35	Clast-supported cobbles.
6.35- ?	Coarse, matrix-supported debris; base unseen
	Grading to floodplain.

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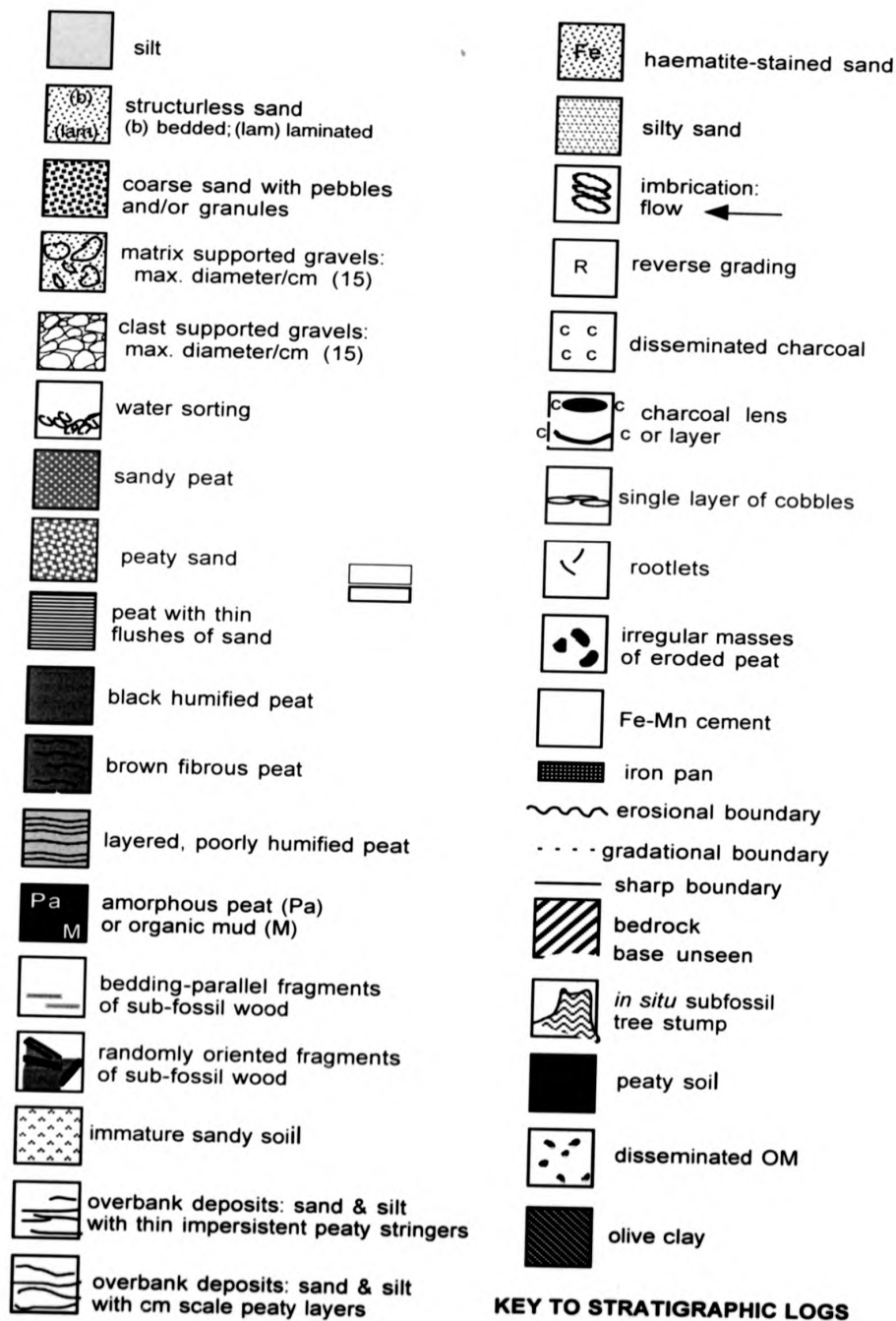Magny M. 1992. Holocene lake-level fluctuations in the Jura and the northern subalpine ranges,

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silt	haematite-stained sand
structureless sand (b) bedded; (lam) laminated	silty sand
coarse sand with pebbles and/or granules	imbrication: flow ←
matrix supported gravels: max. diameter/cm (15)	R reverse grading
clast supported gravels: max. diameter/cm (15)	c c / c c disseminated charcoal
water sorting	c c / c c charcoal lens or layer
sandy peat	single layer of cobbles
peaty sand	rootlets
peat with thin flushes of sand	irregular masses of eroded peat
black humified peat	Fe-Mn cement
brown fibrous peat	iron pan
layered, poorly humified peat	∿∿∿ erosional boundary
Pa / M amorphous peat (Pa) or organic mud (M)	- - - - gradational boundary
	——— sharp boundary
bedding-parallel fragments of sub-fossil wood	bedrock base unseen
randomly oriented fragments of sub-fossil wood	in situ subfossil tree stump
immature sandy soil	peaty soil
overbank deposits: sand & silt with thin impersistent peaty stringers	disseminated OM
overbank deposits: sand & silt with cm scale peaty layers	olive clay

KEY TO STRATIGRAPHIC LOGS