Loch Lomond Stadial plateau icefields in the Lake District, northwest England

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Submitted for the degree of Doctor of Philosophy

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May 1998

Abstract

Detailed geomorphological mapping has revealed evidence for the development of plateau icefields in the central fells of the English Lake District during the Loch Lomond (Younger Dryas) Stadial (c. 12.9–11.5 ka BP). The largest plateau icefield system, which covered an area of approximately 55 km^2 (including outlet glaciers), was centred on High Raise. To the west, smaller plateau icefields developed on Grey Knotts/Brandreth, Dale Head and Kirk Fell, covering areas of 7 km^2, 3 km^2 and 1 km^2 respectively.

The geomorphological impact of these plateau icefields appears to have been minimal on the summits, where the survival of blockfields and other frost-weathered debris (mostly peat-covered) implies the existence of protective, cold-based ice (the Loch Lomond Stadial was the last major episode of periglacial activity to have affected upland Britain). As such, these represent the first reported occurrences of Loch Lomond Stadial ice masses which were not wet-based throughout. Cold-based conditions would have been promoted by a combination of thin, slow-moving ice plus the influence of low mean annual air temperatures on the summits. Ice-moulded bedrock at some plateau edges, however, document a transition to wet-based, erosive conditions. At these locations, steeper slopes would have resulted in increased strain heating within the ice. In many cases, prominent moraine systems were produced by outlet glaciers which descended into the surrounding valleys where their margins became sediment traps for supraglacial debris and inwash. In some valleys, ice-marginal moraines record successive positions of outlet glaciers which actively backwasted towards their plateau source. Given the virtual absence of periglacial trimlines within the area, reconstructed palaeo- ice margins constitute the single most important line of evidence in the identification of plateau icefields in the geomorphological record.

The virtual absence of ice-marginal control points in the upper reaches of these plateau icefield systems means that their reconstructions are somewhat speculative. Ice thicknesses on the summits (40–50 m) are estimates based on the theoretical relationship between plateau icefield depth and summit breadth. A regional Loch Lomond Stadial firn line of 500 m OD is suggested for the central Lake District. This corresponds to a mean annual precipitation of 2,000–2,500 mm, assuming a mean July temperature of 9°C (derived from published coleopteran studies).

This interpretation differs from those of previous workers, who assumed an alpine style of glaciation, with reconstructed glaciers emanating from corries and valley heads. The recognition of Loch Lomond Stadial plateau icefields is critical to the correct geomorphological and glaciological interpretation of this event. In particular, the failure to account for former plateau icefields is significant where they were vital for the maintenance of glaciers in the surrounding corries and valleys, and will result in an overestimation of equilibrium line altitude lowering. Conventional approaches to reconstructing Loch Lomond Stadial glaciers do not take account of the subtle geomorphological impact of plateau icefields on summits and thus it is likely that plateau icefields were more common at this time than has hitherto been appreciated.

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Acknowledgements

I very much enjoyed geography at school, particularly under the guidance of Mr John MacFarlane at the Vale of Leven Academy in Alexandria, to whom I'm very grateful, but the relevance of a geography degree to the 'real world' (whatever that is) escaped me at the time. On leaving school, I entered the services of the Bank of Scotland in Helensburgh, but quickly realised that a geography degree would provide a greater insight to the 'real world' than the Bank could ever offer. One year later, I started my degree at the Department of Geography and Topographic Science at the University of Glasgow. I had a great time during my four years as an undergraduate and I'm very grateful to the staff for all their hard work. To them, I'd like to say a big thank you. Of course, none of this would have been possible without the unstinting support and encouragement from my parents (although they did resort to hiding the chocolate biscuits).

My enthusiasm for the Loch Lomond Stadial glaciation was fired initially by Bob Price's 'Palaeoenvironments' option, which enabled me to better appreciate the glaciogenic landform assemblages in the vicinity of my home in Balloch, at the southern end of Loch Lomond. In turn, this prompted me to investigate the Loch Lomond Readvance end moraine in nearby Glen Fruin for my dissertation, an exercise which I thoroughly enjoyed. My interest in palaeoenvironments developed further with the 'Quaternary Environments' option in my final year, run by the newly-recruited Dave Evans (who introduced me to the quaint Quaternary custom of food-patting). The opportunity to pursue a similar line of research as a postgraduate student was therefore a dream come true. To this end, I am extremely indebted to Dave Evans and Jim Hansom for encouraging me to apply for a postgraduate position and suggesting potential research topics, and the Faculty of Science of the University of Glasgow for the Studentship which made it financially viable.

I'd like to thank my supervisor, Dave Evans, for his help and encouragement during my somewhat protracted period as a PhD student. I'm sure there were times, particularly during my second year, when he must have wondered whether I'd ever get it together! Indeed, the present research topic – the identification of Loch Lomond Stadial plateau icefields in the Lake District – didn't emerge until late 1993. I knew this topic was worth pursuing (even at the expense of putting aside much of the fieldwork I had completed up until then) when the nature and distribution of these glacial landforms suddenly began to make sense. If Dave had given me less freedom and insisted that I completed within three years, I am in little doubt that I would have produced a very conventional, alpine style reconstruction of these former glaciers, differing little from the earlier work of J.B. Sissons (1980a). Thanks for your patience, Dave! Can I have that header now, please?

I would also like to thank the various individuals who have helped me in one way or another. They include Jim Hansom, John Briggs, John Jowett, Trevor Hoey, Gordon Dickinson and Stella Lowder (all at Glasgow), Brice Rea (Cardiff), Doug Benn (Aberdeen), John Boardman (Oxford), Alan Fishwick (Lake District National Park, Kendal) and John Fagg (Worcester). I'm also pleased to acknowledge the support of the School of Sciences research fund at University College Worcester for fieldwork expenses and the purchase of aerial photographs.

I owe my greatest debt of gratitude to Isobel, my long-suffering wife, who has endured my PhD for almost six years now. Her patience and understanding has gone well beyond what could reasonably be expected. Having said that, she threatened me with divorce four weeks ago unless I finished within the month. A close call indeed! We can now look forward to spending more time together; I'm sure we'll get on just fine.

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Introduction

1.1 INTRODUCTION

The abrupt thermal amelioration that marked the termination of the last major glaciation in Britain at c.14.7 ka BP effectively represented a false start to the present (Flandrian) interglacial. Temperatures subsequently declined during the Lateglacial Interstadial, with a final sharp deterioration at c.12.9 ka BP (Lowe *et al.*, 1995). This marked the return to full glacial conditions and the onset of the Loch Lomond (Younger Dryas) Stadial, c.12.9–11.5 ka BP (c.11–10 ^{14}C ka BP), an event that appears to have been both precipitated and terminated by abrupt reorganisations of the ocean-atmosphere subsystem in the North Atlantic (e.g. Wright, 1989; Broecker and Denton, 1990a, b).

Glaciers developed in the mountains of northern and western Britain during the Loch Lomond Stadial, with the Western Highlands of Scotland hosting an extensive icefield (Figure 1.1) (Thorp, 1991; Bennett and Boulton, 1993a, b). Elsewhere, the glaciation style is believed to have been predominantly alpine in nature, with glaciers restricted to valleys and cirques (Sissons, 1979a; Gray and Coxon, 1991). This glaciation has been intensively studied and palaeoclimatic inferences have been derived from the reconstructed surface profiles of these former ice masses (e.g. Sissons, 1972, 1974, 1980a, b; Sissons and Sutherland, 1976; Cornish, 1981; Ballantyne, 1989; Mitchell, 1991, 1996; Shakesby and Matthews, 1993).

There is a consensus within the literature that the geomorphological record is both clear and complete enough to have facilitated accurate reconstructions of Loch Lomond Stadial glaciers at maximal extents (e.g. Ballantyne and Harris, 1994, p.18; Lowe and Walker, 1997, p.30). Research in contemporary glacial environments, however, has cast doubt on the reliability of the procedures that have been employed in these reconstructions, particularly with regards to ice-marginal delimitation in former accumulation zones where geomorphological evidence may be poorly developed or absent. Extrapolated ice margins have generally been based on the assumption of an alpine style of glaciation, with glaciers emanating from cirques and valley heads (e.g. Sissons, 1980a; Cornish, 1981). This assumption may not, however, be appropriate for areas characterised by broad, rounded summits (the Cairngorms and parts of the Lake District, for example) where plateau icefields may have developed. Recession of some

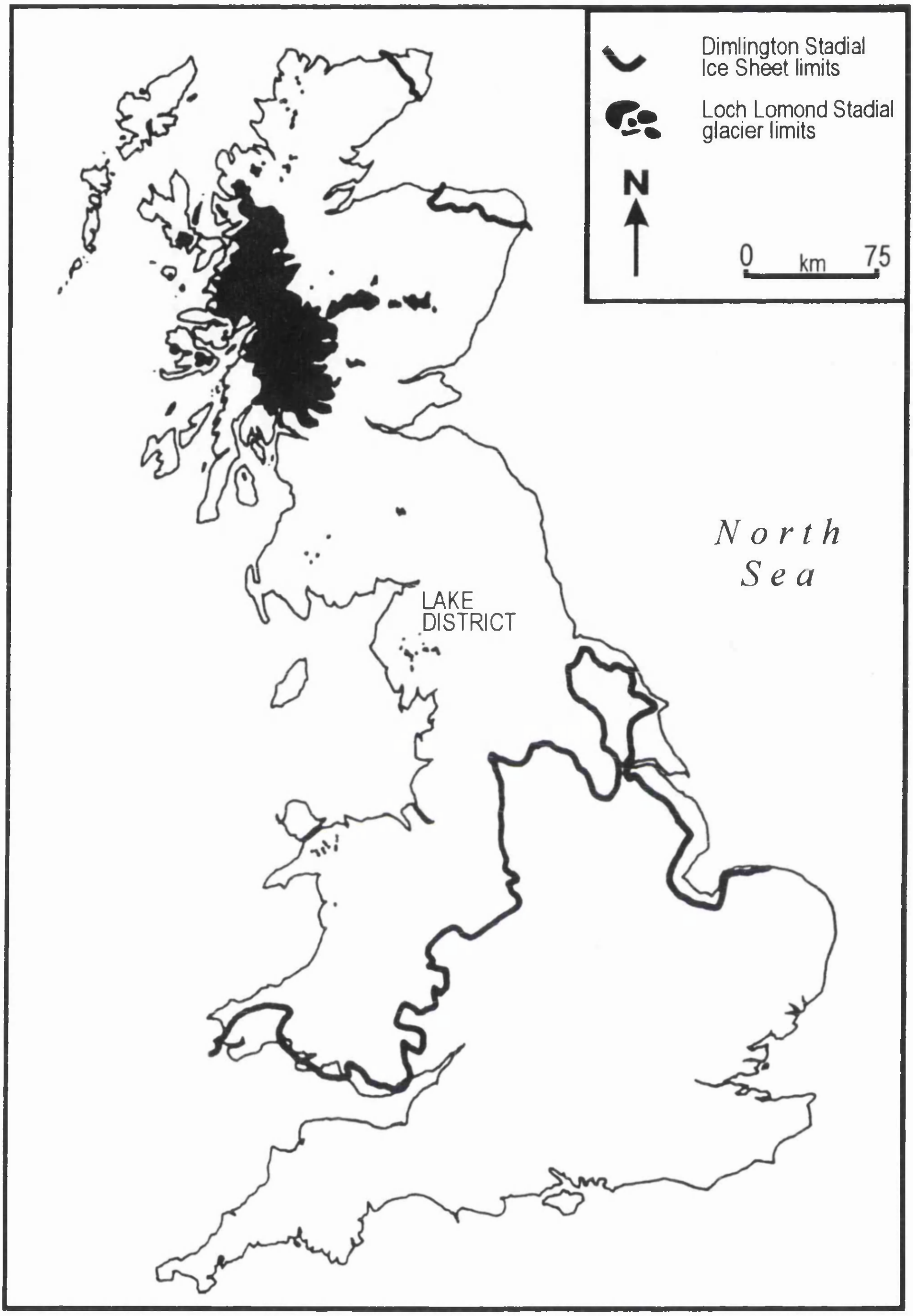


Figure 1.1 The Loch Lomond Stadial glaciation in Britain.
After Gray and Coxon (1991)

The major area of glacier development was in the high ground of the Western Highlands of Scotland, from Loch Torridon in the north to Loch Lomond in the south, a distance of approximately 175 km. Note the apparent marginality of the Cairngorms, the Southern Uplands, Wales and the Lake District for glacier development.

contemporary plateau icefields in north Norway has revealed areas that have experienced little or no subglacial erosion, a situation which has been attributed to low basal shear stresses and, in some places, cold-based ice (Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988; Rea *et al.*, in prep.). This implies that the identification of former plateau icefields in deglaciated areas may be problematical (Gordon *et al.*, 1987).

This study investigates whether the Loch Lomond Stadial glaciation in the Lake District was characterised by plateau icefields in addition to the corrie and valley glaciers already reported in the literature (cf. Manley, 1959; Sissons, 1980a). The recognition of Loch Lomond Stadial plateau icefields is critical to the correct geomorphological and glaciological interpretation of this event. In particular, the failure to account for former plateau icefields is significant where they were vital for the maintenance of glaciers in the surrounding corries and valleys, and will result in an overestimation of equilibrium line altitude (ELA) lowering. Furthermore, the possibility that such plateau icefields may have been cold-based and thus non-erosive must be addressed when assessing the significance of relict major periglacial landforms on Lake District summits. For example, some blockfields may have survived beneath relatively thin cold-based ice and thus pre-date the Loch Lomond Stadial. Weathering contrasts in such circumstances may therefore reflect variations in basal thermal regime rather than the presence or absence of glacier ice during the Loch Lomond Stadial (*cf.* Ballantyne, 1984; Ballantyne and Harris, 1994).

The remainder of this chapter is divided into three sections. Section 1.2 outlines the aims and rationale of this study in greater detail. This is followed by an overview of climate change during the last glacial–interglacial transition, with a particular emphasis on the Loch Lomond (Younger Dryas) Stadial. Finally, Section 1.4 provides an overview of the geological history of the Lake District.

1.2 AIMS and RATIONALE

1.2.1 Aim

The Lake District is assumed to have been characterised by an alpine style of glaciation during the Loch Lomond Stadial, with reconstructed glaciers emanating from corries and valley heads (Manley, 1959; Sissons, 1980a). This is despite the existence of numerous broad, rounded summits which, given suitable conditions, would have the potential to support small icefields during a restricted glaciation. The aim of this research is to establish whether there is any geomorphological evidence for the development of Loch Lomond Stadial plateau icefields on the relatively rounded summits investigated. If so, an assessment will be made of their palaeoclimatic significance.

1.2.2 Rationale

Investigations in contemporary glacial environments have demonstrated that some plateau icefields have a minimal geomorphological impact. In the Lyngen Peninsula, for example, patterned ground and blockfield has been revealed by the recession of the Jiek'kevárri, Bálgesvárri and Bredalsfjellet plateau icefields (Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988). The survival of periglacial phenomena beneath former plateau icefields has also been described by Evans (1988, 1990) in NW Ellesmere Island, where deglaciated summits lack obvious glacial erosional features and are often characterised by well-developed patterned ground and tors.

The preservation of periglacial phenomena beneath some plateau icefields requires cold-based, non-erosive ice. Such conditions are promoted by the dynamics of relatively thin, slow-moving ice, plus the impact of low mean annual air temperatures on the plateau (Section 2.2). That some Loch Lomond Stadial plateau icefields *may* have been cold-based must be addressed when assessing the palaeoenvironmental significance of relict major periglacial landforms in upland Britain. It is generally assumed that there is a mutually-exclusive relationship between the distribution of relict major periglacial landforms, such as blockfields, and that of Loch Lomond Stadial glaciers (e.g. Ballantyne, 1984; Ballantyne and Harris, 1994). Indeed, various researchers have employed contrasts in weathering and periglacial landform development to assist in the delimitation of the higher reaches of some Loch Lomond Stadial glaciers (e.g. Sissons, 1974, 1977a, b; Ballantyne, 1989). Nevertheless, such an interpretation assumes warm-based conditions throughout and does not take into account the possibility that some of these periglacial landforms may have survived beneath relatively thin, cold-based ice on summits and thus pre-date the Loch Lomond Stadial.

The recognition of former plateau icefields, irrespective of thermal regime, is also necessary for the reconstruction and palaeoclimatic interpretation of ELAs. The failure to account for plateau icefields becomes particularly significant where they nourished glaciers in the valleys below, either contiguously or by avalanching. Palaeoclimatic inferences derived solely from the reconstructed valley glacier component of such a system will result in an overestimation of ELA lowering. As the size of the plateau icefield increases in relation to the valley glacier component, so the magnitude of the error will increase (Rea *et al.*, in prep.).

In his reconstruction of the Loch Lomond Stadial glaciation in the Lake District, Sissons (1980a) invoked three separate snowfall intensity zones, in combination with the effects of snowblow and aspect, to account for considerable variations in palaeoglacier extents and ELAs (Section 2.4). However, these variations may also have reflected topographic variability and specifically, on some of the broader summits, the development of hitherto unrecognised plateau icefields. When compared with corrie glaciers incised into narrow summits, outlet glaciers draining extensive plateau icefields are potentially capable of contributing greater volumes of ice into valleys by virtue of their larger accumulation areas.

This is the first study to attempt to address the concerns expressed by Gordon *et al.* (1987, 1988) and Gellatly *et al.* (1988) regarding the identification of former plateau icefields in deglaciated areas. It is, in essence, an exercise in geomorphological mapping. Although it is acknowledged that the potentially equivocal nature of the geomorphological record may present problems of interpretation, the importance of recognising former plateau icefields in palaeoenvironmental reconstructions provides the basis for pursuing this investigation.

1.3 THE LOCH LOMOND (YOUNGER DRYAS) STADIAL

1.3.1 Hemispheric Patterns

In those countries bordering the North Atlantic, the most recent glacial-interglacial transition was not straightforward, with a return to full glacial conditions following initial thermal amelioration, effectively constituting a 'false start' to the present interglacial. This transitional period is referred to in north-west Europe as the Lateglacial, c 14.7–11.5 ka BP. Figure 1.2 shows a comparison of accumulation-rate data from the GISP-2 core (Greenland ice sheet) with palaeotemperature data derived from fossil coleoptera in Northeast England (Lowe *et al.*, 1995). A number of very general points can be made. The termination of the Dimlington Stadial was very rapid indeed, with the thermal maximum of the Lateglacial interstadial being achieved almost immediately thereafter at *c.* 14.7 ka BP. The Lateglacial interstadial, 14.7–12.9 ka BP, was characterised by a downward trend in temperatures, with a final rapid decline *c.* 12.9 ka BP. This final cooling marked the return to full glacial conditions and the onset of the Loch Lomond Stadial/Younger Dryas, an event which lasted approximately 1.3 $\pm$0.07 ka. As with the Dimlington Stadial before it, the Loch Lomond Stadial/Younger Dryas terminated extremely rapidly; snow accumulation-rates in GISP-2 doubled rapidly in the space of between one and three years. This marked the onset of the present interglacial, known in Britain as the Flandrian, at *c.* 11.5 ka BP.

There is now very good evidence to suggest that the Loch Lomond (Younger Dryas) Stadial was precipitated (and terminated) by relatively rapid changes in North Atlantic circulation, particularly in the process of deep water formation in the vicinity of Iceland (Broecker and Denton, 1990a, b). The process of North Atlantic Deep Water (NADW) formation is the mechanism by which, every winter, water in the vicinity of Iceland sinks to the bottom of the Atlantic and flows south and around the Cape of Africa (Figure 1.3). It is part of a global deepwater circulatory system, and the North Atlantic is a major source of this water; the vertical transfer involved averages about twenty times the combined discharge of all the worlds rivers. The water which sinks originates as an intermediate-depth, high salinity north-flowing current which, south of Iceland, is brought to the surface where it quickly cools from 10°C to 2°C. A combination of high salinity and low temperature makes the water unusually dense, causing it to sink. The

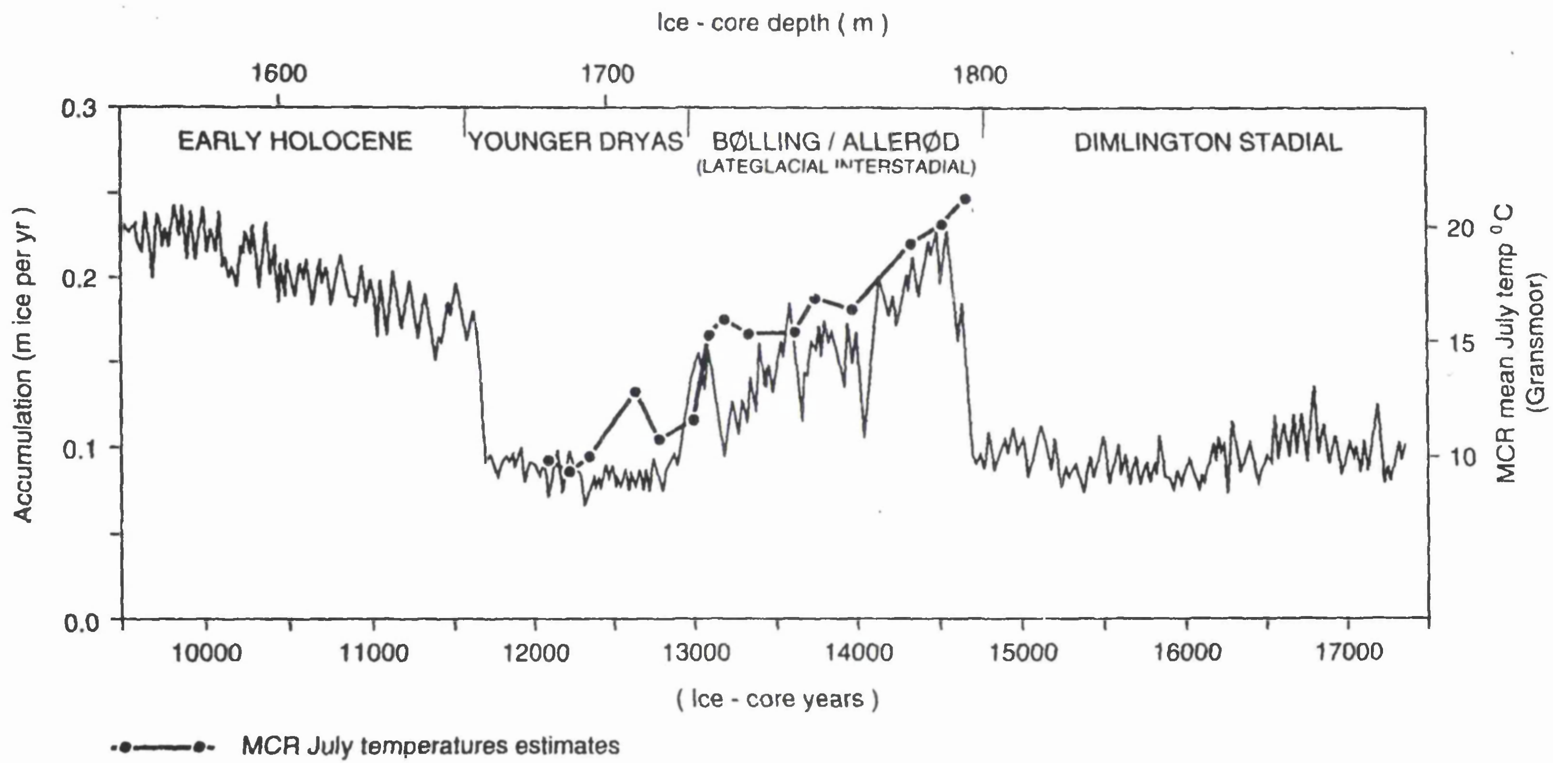


Figure 1.2 A comparison of accumulation-rate data from the GISP-2 ice core (Greenland Ice Sheet), and palaeotemperature data derived from fossil Coleoptera (Mutual Climatic Range) from Gransmoor, northeast England (after Lowe *et al.*, 1995).

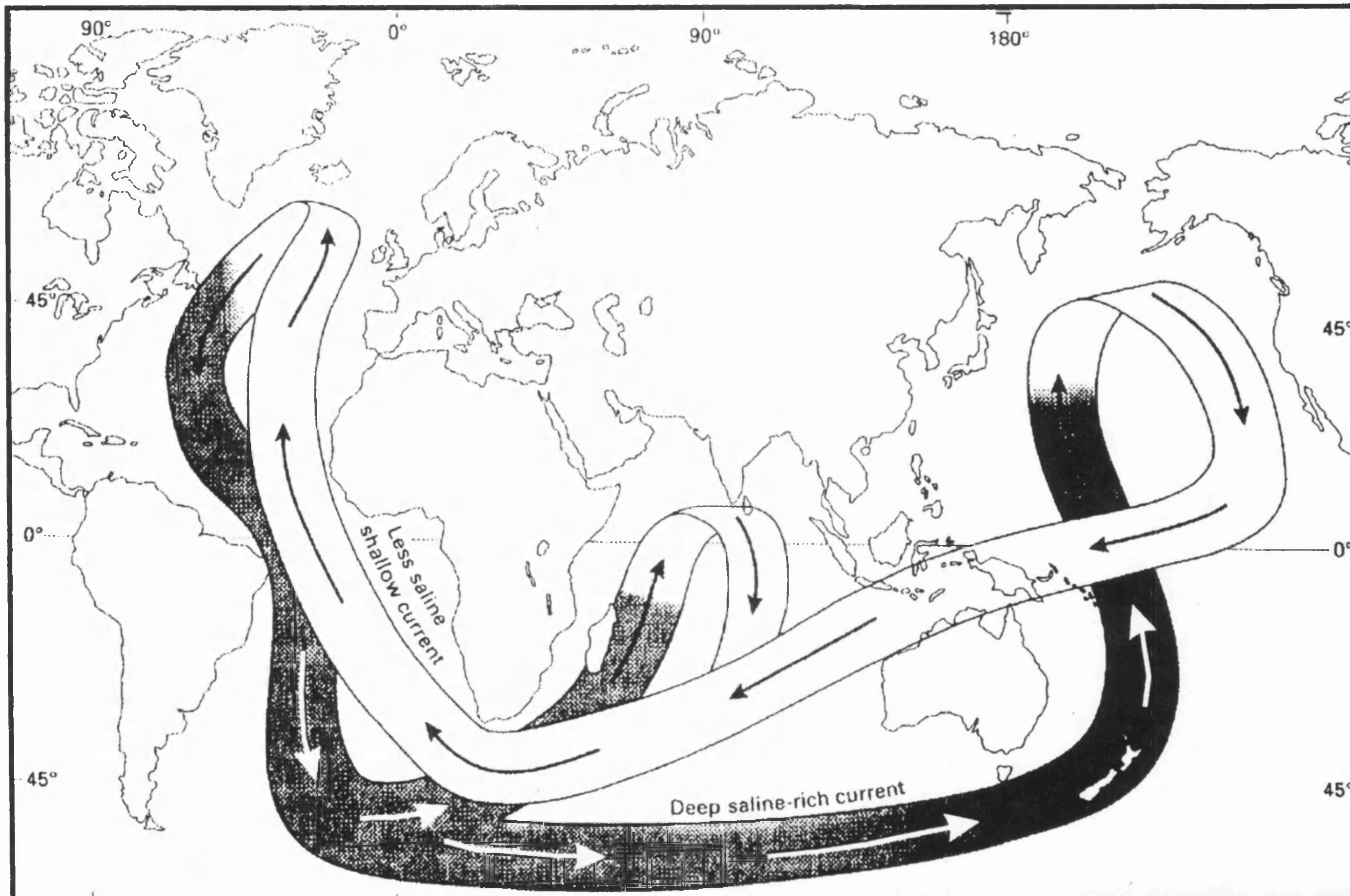


Figure 1.3 The large scale salt transport system operating in the present oceans. See text for explanation.
Source: Lowe and Walker (1997, Figure 7.27, p. 8)

climatological significance of this process is that, prior to sinking, the water releases a great deal of heat as it cools. It has been estimated that the heat provided is equal to about 30% of the yearly direct input of solar energy to the surface of the northern Atlantic, and it is this that is responsible for the very mild winters of western Europe north of the Alps and Pyrenees (Rind *et al.*, 1986). Thus, if NADW formation were to cease, then winter temperatures in western Europe would probably fall quite dramatically to levels more typical of its latitudinal range. It is considered that this is exactly what has happened on numerous occasions in the past, resulting in the oscillatory nature of the North Atlantic climate observed from ice cores, which is superimposed on longer term glacial–interglacial cycles (Alley *et al.*, 1993).

Evidence for the cessation of NADW formation in the past has been obtained from the study of benthic foraminifera in North Atlantic sediment cores. The cadmium measurements of Boyle and Keigwin (1987) and the carbon isotopic measurements summarised by Duplessy *et al.* (1988) demonstrate that the pattern of deep circulation was quite different during glacial times (Broecker and Denton, 1990a, b). Cadmium measurements are considered to reflect former nutrient and phosphate levels, which in turn are strongly controlled by ocean currents. At present, there is an increasing gradient in nutrient levels away from the source of NADW. This is because the source of the deep water is at the surface south of Iceland, where biological activity depletes nutrient levels. However, as the water flows south at depth, it collects sinking nutrients and phosphates, resulting in the enrichment of the current. This nutrient and phosphate gradient is mirrored by cadmium concentrations in benthic foraminifera, although it is not altogether clear why this should be so (Boyle and Keigwin, 1987). The absence of such a gradient during the Dimlington Stadial and Loch Lomond Stadial, but its presence in the Lateglacial interstadial and present interglacial, has been interpreted as demonstrating the significance of NADW formation in determining the climate of north-west Europe in the past as well as the present.

During the Dimlington Stadial, the Loch Lomond Stadial and, presumably, earlier cold episodes, the absence of NADW formation south of Iceland meant that there was no additional heat being provided to the North Atlantic, and this resulted in the southwards migration of the Polar Front. The end of the last glacial, *c.* 15 ka BP, witnessed a resumption in NADW formation and was associated with the northwards movement of the Polar Front. However, about 12.9 ka BP, analyses of both sea surface temperatures as well as cadmium measurements reveals that deep water formation had ceased yet again (Broecker and Denton, 1990a, b). This event, which lasted approximately 1.3 ka, corresponds with the Loch Lomond (Younger Dryas) Stadial.

Although relatively short-lived, there is abundant evidence which testifies to the severity of this episode in the North Atlantic region. Greenland ice cores record the return to mid-glacial conditions (e.g. Alley *et al.*, 1993; Taylor *et al.*, 1993a, b), and a range of terrestrial evidence records the transition from forest to tundra vegetational communities, and the return to cold climate geomorphological processes. In Britain, it witnessed the return of glacial and periglacial conditions for the last time before the onset of the present interglacial.

1.3.2 Evidence from the British Isles

A variety of lithological, morphological and biostratigraphical evidence exists for the Loch Lomond Stadial, particularly in northern and western Britain (Sissons, 1979a). Perhaps the most distinctive evidence is for the expansion of glaciers at this time. Evidence for a glacial event was first recognised in the south-eastern Loch Lomond basin by Jack (1875) and placed in a regional context by Simpson (1933), who clearly identified the limits of these glaciers along the Highland border west of the River Tay. The distinctive landforms and deposits associated with the palaeoglacier that occupied the Loch Lomond basin allowed him to map its piedmont margin. The limits of the Loch Lomond Readvance were subsequently mapped elsewhere, most notably during the 1970s by J.B.Sissons and several of his co-workers (see review in Gray and Coxon, 1991). It is generally accepted that these glacigenic landform assemblages remain precise and clearly defined, and hence heavy reliance has been placed on morphostratigraphy in attempts to delimit these former glaciers.

Evidence for the Loch Lomond Readvance has been recognised in many parts of upland Britain. The major area of glacier development was in the high ground of the western

Highlands, from Loch Torridon in the north to Loch Lomond in the south, a distance of some 175 km, and from Loch Shiel in the west to Loch Rannoch in the east, a distance of 150 km (Sissons, 1979a). Ice thicknesses in excess of 400 m occurred on Rannoch Moor and in the Great Glen (Thorp, 1984, 1986). In most places, the configuration of this ice mass took the form of a complex network of transecting glaciers interrupted by nunataks. The longest individual ice streams, each around 50 km, occupied the troughs of Loch Lomond, Loch Awe, Loch Rannoch and Loch Garry. The Loch Lomond glacier is estimated to have had a volume of 80 km^3, and whose terminus took the form of a piedmont lobe some 20 km wide (Sissons, 1979a, 1983; Price, 1983).

In addition to the Western Highlands icefield, 200 other individual ice masses have been identified. Of these, only four were plateau icefields (which occupied the Gaick and Glen Mark areas of the eastern Highlands, Mull, and the Cuillins of Skye), with most of the remainder believed to be cirque or valley glaciers. However, glaciers were not ubiquitous in the Highlands. In the Monadliath Mountains, where large areas of ground exceed 600 m, there were apparently only two (Sissons, 1979a). And in the north-west Cairngorms, only three out of seven cirques at 1000 m altitude contained glaciers (Sissons, 1979b). The Southern Uplands, Lake District and Wales are considered to have been very marginal for glaciation at this time, with ice masses generally occupying the most favourable sites (Sissons, 1977a, b, 1979a, 1980a, 1983; Cornish, 1981; Gray, 1982).

Altitudinal distributions of Loch Lomond Readvance glaciers at their maximal extents, and particularly that of reconstructed ELAs, have been used to derive palaeoclimate inferences for the period. These have been used to estimate mean July temperatures for this period of 6–8°C in northern Britain (Gray and Coxon, 1991). These figures are in close agreement with those calculated on the basis of coleopteran assemblages, which suggest mean July temperatures of around 10°C in southern England, 9°C in central England and north Wales, and 8°C in southern Scotland and the Lake District (Coope, 1977a, b). The widespread distribution of permafrost indicated by the periglacial evidence suggests that mean annual temperatures were rarely above freezing point and may have been as low as, or even lower than, –6° to –8°C. Hence, mean January temperatures of –17° to –20°C are suggested for much of the British Isles (Gray and Coxon, 1991).

Generally low levels of precipitation are implied for much of the British Isles during the Stadial, with glacier development resulting from a band of snowfall associated with increased cyclonic activity along the polar front which affected, in particular, northern and western parts of the country as it moved southwards. Glacier distributions, dimensions and altitudes suggest that the principal snow bearing winds were south-easterly or south-westerly (Sissons, 1979a), and that marked precipitation gradients developed in highland Britain between the maritime margin and the inland areas. Across the Scottish Highlands, palaeo-ELAs generally rise eastwards, from about 300 m in the Inner Hebrides and Southwest Highlands, to the Cairngorms where ELAs may have been as high as 1000 m (Ballantyne and Harris, 1994). In addition, there appears to have been a steep northwards rise parallel to the Highland Boundary Fault in the area of the South-eastern Highlands. These trends imply marked aridity in the Cairngorms during the Stadial, and a very steep decrease in precipitation both east and north of the Highland Boundary Fault (Figure 1.4).

There is some evidence to suggest that precipitation decline towards the end of the Loch Lomond Stadial resulted in the active retreat of the palaeoglaciers prior to rapid thermal amelioration (Pennington, 1978; Benn *et al.*, 1992). This hypothesis is supported by the occurrence of ice wedge casts in sediments within the Loch Lomond Readvance limits at a number of sites in Scotland (Sissons, 1974), and by the fact that in the Lake District several glaciers had withdrawn from their maximal positions before the end of the Artemisia pollen assemblage zone (Pennington, 1978). More recently, a combination of geomorphological and palynological research in Skye by Benn *et al.* (1992) has also confirmed that deglaciation may have occurred in two phases, one of interrupted retreat in response to precipitation decline, and the second of uninterrupted retreat, associated with rapid thermal amelioration. Ice core evidence suggests that thermal amelioration was abrupt, and may have occurred in as little as five years (Alley *et al.*, 1993).

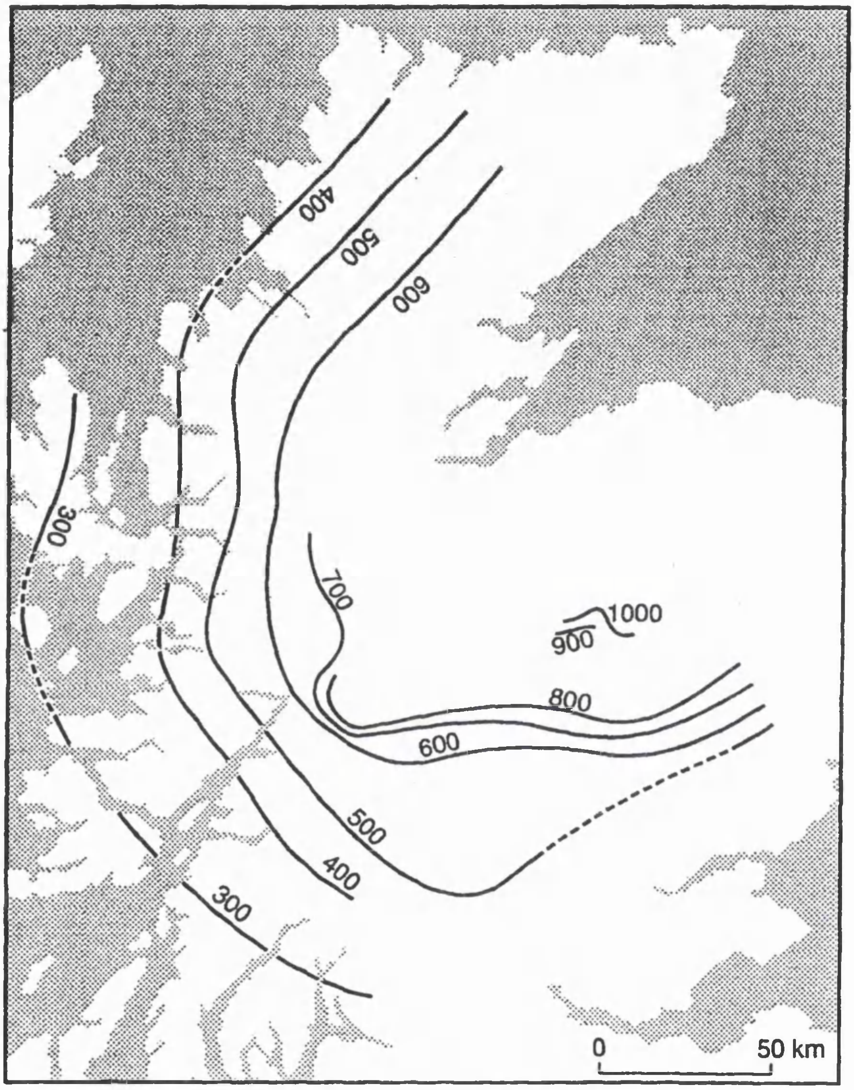


Figure 1.4 Regional pattern of Loch Lomond Readvance ELAs at maximal extents.
Source: Ballantyne and Harris (1994, Figure 2.4, p. 20)

1.4 LAKE DISTRICT GEOLOGICAL HISTORY

The Lake District is located in north-west England between 54 and 55°N, bounded by the Solway Plain to the north, the Vale of Eden to the east, and the Fylde Lowlands to the south (Figure 1.5). It is a small dome of Lower Palaeozoic rocks, an inlier protruding from beneath a cover of Carboniferous and Permo-Triassic rocks (Smith, 1992) (Table 1.1 and Figure 1.6). Although only 50 km in diameter, it exhibits considerable geological and topographic diversity, and includes the highest peak in England, Scafell Pike (NY215072), which attains a height of 978 m.

The Lake District massif contains three lithostratigraphic units, separated by unconformities, in an overall southeast-younging sequence (Kneller and Bell, 1993) (Figure 1.6). The Tremadoc-Llanvirn Skiddaw Group, a turbidite-dominated marine succession, crops out in a WSW-ENE belt in the north, and are the oldest rocks present. Unconformably overlying the Skiddaw Group is the Borrowdale Volcanic Group, an Ordovician (Llandeilo-Caradoc) volcanic succession over 4 km thick, and this makes up the central Lake District. The Borrowdale Volcanic Group is, in turn, unconformably overlain by the Windermere Supergroup, a folded and cleaved sequence of predominantly marine sedimentary rocks, which crops out in the south of the district. Each group gives rise to slightly different topography. The mountains of the Skiddaw Group are typically smooth-sided and vegetated, contrasting markedly with the more rugged appearance of the Borrowdale Volcanic Group, south of which are the low rolling hills of the Windermere Supergroup.

1.4.1 The Skiddaw Group

The Tremadoc-Llanvirn Skiddaw Group, the oldest rocks in the district, comprises almost one third of the mountain core, and crops out in four inliers. The largest of these is the Skiddaw Inlier, which extends from Cleator Moor in the west to Troutbeck in the east. Smaller inliers occur around Ullswater, Bampton, and Black Combe. The Group consists of over 5 km of greywacke, siltstone, and mudstone beds, intensely folded in places, and has been interpreted as possibly representing evidence for fore-arc sedimentation (Moore, 1992).

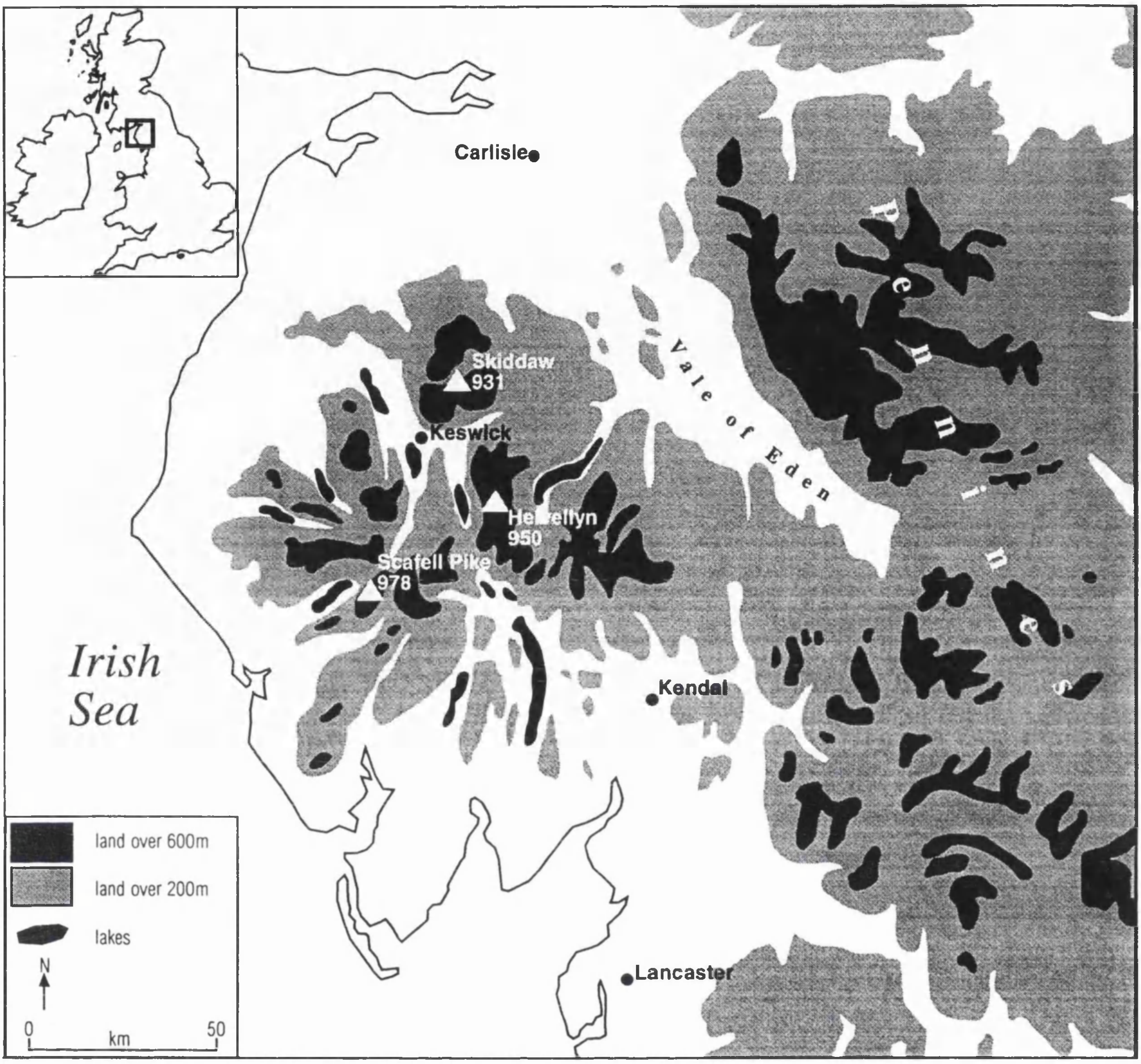


Figure 1.5 Location and topography of the Lake District and adjoining areas.

Stratigraphic Groups	Principal Local Formations	Major Events
Lower Palaeozoic (600-405 Ma)		
ORDOVICIAN 510–450 Ma	The Skiddaw Group (510-460 Ma) of the northern and western Lake District.	Deep-sea mudstones and sandstones laid down as the Iapetus Ocean narrowed.
	The Borrowdale Volcanic Group (460-450 Ma) of the central Lake District.	Increasingly violent eruptions in land and in shallow water as the ocean crust was subducted.
SILURIAN 450–405 Ma	The Windermere Supergroup of the southern Lake District.	Shales and sandstones deposited in seas deepening as the Iapetus Ocean closed.
Upper Palaeozoic (405-250 Ma)		
DEVONIAN 405–35 Ma	Caledonian mountain building and granite intrusions.	Continental collision produced mountain ranges followed by rapid erosion.
CARBONIFEROUS 355–290 Ma	Carboniferous Limestones around and possibly over the Lake District (355-320 Ma)	Shallow, clear seas in tropical latitudes with coral reefs.
	Coal measures in West Cumbria.	Tropical rain forests and extensive deltas.
290 $\pm$ 10 Ma	Hercynian mountain building.	Continental collision and uplift of Lake District.
PERMO-TRIASSIC 290–205 Ma	The New Red Rocks of West Cumbria, the Vale of Eden, near Carlisle and around Barrow.	Hot desert climates. Deposition on land, in shallow salt lakes and seas.
MESOZOIC and TERTIARY 205–2 Ma	No rocks of these ages remain in the Lake District.	The Lake District was land.
QUATERNARY c. Last 2 Ma	Glacial and interglacial deposits.	Rapid climatic change. Glacial, tundra and interglacial climate.

Table 1.1 An outline geological history showing principal formations and events.

After Smith (1992, Table 2, p.4)

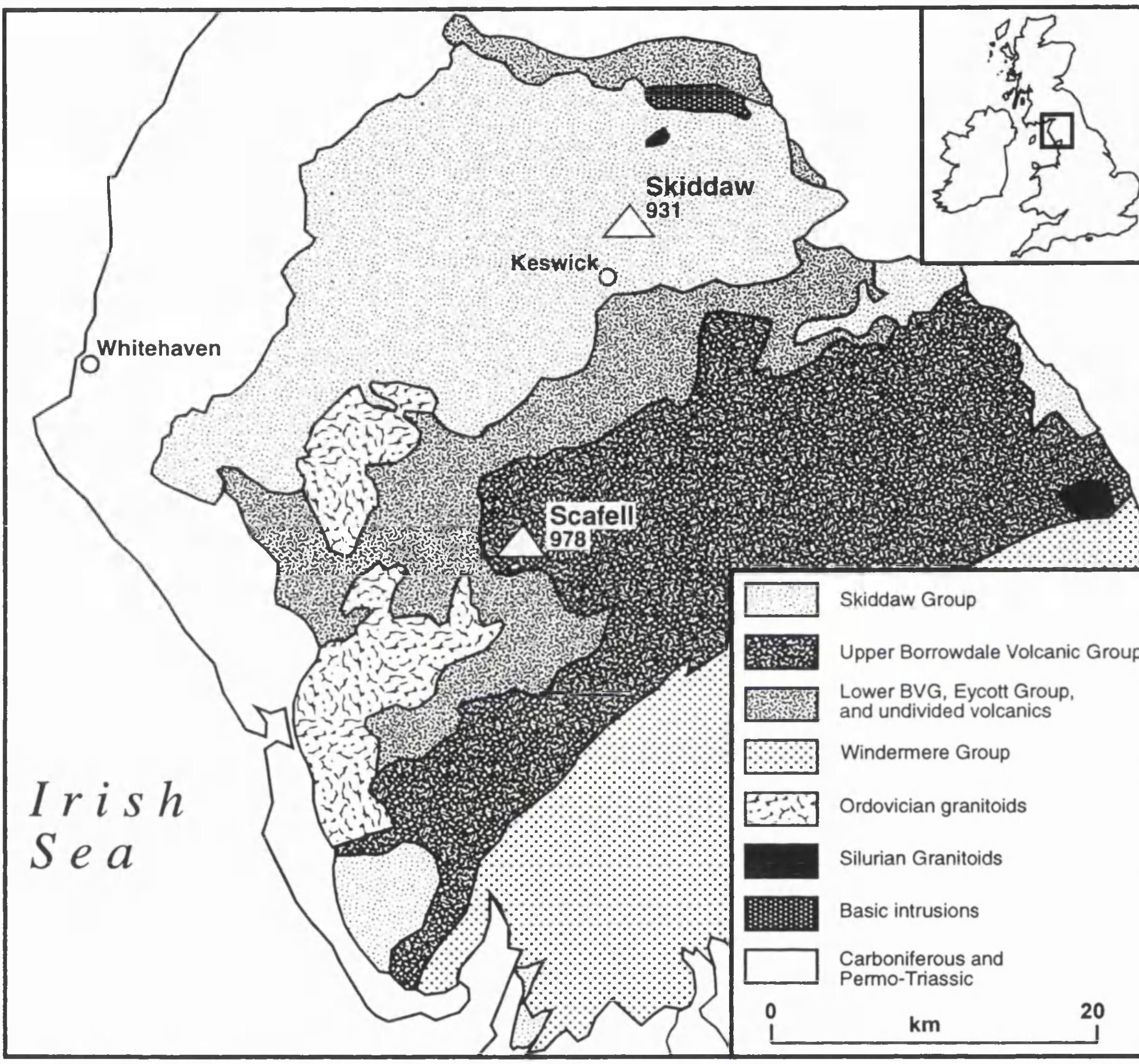


Figure 1.6 Lake District Geology
After Branney and Kokelaar (1994)

The Skiddaw Group possesses a very complex structure. In addition to having undergone many phases of folding, they are also highly cleaved and altered by thermal metamorphism and mineralisation. Major north-easterly trending faults delimit distinct zones in the Skiddaw Group. In the north-western fells, a thrust fault crosses from the Crummock Water area through to Causey Pike. To the north of this fault, slump structures due to movement on the continental slopes dominate. To the south, complex mixtures of rock with many olistoliths are usual (Smith, 1992).

The Skiddaw Group are relatively homogenous with respect to erosion, being rather weak, resulting in their characteristic smooth-sided, vegetated slopes. Crags are relatively rare, and are associated with the greywackes. Despite their relative weakness, two of the highest peaks in the Lake District occur in Skiddaw Group rocks. Skiddaw (NY260291) and Blencathra (NY323277) attain altitudes of 931 m and 808 m respectively.

1.4.2 Ordovician Volcanics

The Ordovician volcanics are highly variable. The oldest is the Eycott Group, which comprises mainly basaltic and basaltic-andesite lavas, and may have been erupted partly in submarine conditions as some of the Skiddaw Group sediments are interbedded with them. There then follows more than 2.5 km of basalts, andesites, and subsidiary dacites of transitional theoleiitic-calc-alkaline composition (Fitton and Hughes, 1970). The Eycott Group crops out over a relatively small area ($<50km^2$) in a narrow belt across the extreme northern fringe of the district. It is likely that some of the basic intrusives belong to this Eycott episode, in particular the Carrock Fell Complex, and minor intrusions such as those of Castle Head, Keswick and Embleton (Firman, 1978).

The Borrowdale Volcanic Group succeeds the Eycott Group. It comprises the central Lake District and is part of an elongate chain of Caledonian magmatism lying south of the Iapetus Sutre, stretching from Dingle in south-west Ireland to the northern Pennines. Within the Lake District, the Borrowdale Volcanic Group crops out over an area of 800 km^2 and attains a maximum thickness of 6 km, and extends from the Duddon valley in the west through to Haweswater in the east. It comprises mostly non-marine basaltic, andesitic, dacitic, and rhyolitic lavas, sills and tuffs with interstratified volcaniclastic sedimentary rocks. Recent research has demonstrated that the rocks record subaerial ensialic arc volcanism with rapid synvolcanic subsidence that preserved them from erosional removal. The subsidence is interpreted as due to caldera collapse and to mild extension across the arc (Branney, 1988; Branney and Kokelaar, 1994).

The Borrowdale Volcanic Group can, broadly speaking, be divided into lower and upper sub-groups, reflecting differences in the types of volcanic processes. The lower BVG, known as Phase 1, is dominated by the products of effusive volcanism whereas the upper BVG (Phase 2) is dominated by products of explosive volcanism (Petterson, 1990). Many faults within the BVG are volcano-tectonic in origin.

There is a considerable amount of evidence to suggest that the BVG was erupted in a subaerial environment. Negative evidence exists in the absence of marine sediments, pillow lavas and marine fossils. Positive evidence is provided by the abundance of unconformities within the Phase 2 succession, mantling and draping structures within air-fall tuff sequences, common occurrences of fluviatile sediments, and the existence of pyroclastic surge and thin welded pyroclastic flow horizons (Branney, 1988). Some tuff formations, however, were deposited subaqueously, probably in ephemeral lakes (e.g. parts of the Honister and Seathwaite Fell tuffs).

The heterogeneity of the Borrowdale Volcanic Group gives rise to the rugged, stepped topography of the central Lake District, with resistant beds standing out as steep crags. Even so, the landscape cannot be described as alpine; many of the summits are rounded, with some of them (e.g. High Raise – NY281095) being better described as plateaux.

1.4.3 Windermere Supergroup

The Windermere Supergroup crops out in the southern part of the Lake District, in a small inlier at Cross Fell, a fault bounded inlier at Dry Gill and in a series of small inliers at the southern margin of the Askrigg block. It is a folded and cleaved sequence of predominantly marine sedimentary rocks that unconformably overlies the mid-Ordovician Borrowdale Volcanic Group, and is itself unconformably overlain by gently dipping

Lower Carboniferous or possibly Upper Devonian rocks. The fauna contained in the Supergroup indicates almost continuous marine deposition from the Longvillian (locally) to the Pridoli.

The Windermere Supergroup are less resistant than the Borrowdale Volcanic Group, and more varied than the Skiddaw Group. The result is topography on a smaller scale than that of the central Lake District; the south lacks the relief to show the full effect of highland glaciation.

1.4.4 Regional glaciation

Despite over a century of research, very little is known about the flow directions and detailed dynamics of the last (Dimlington Stadial) ice sheet in Cumbria (Mitchell and Clark, 1994). Reconstructions have been based on a range of field evidence, including striations, erratics, till distribution and drumlins. The established model is one in which a small ice dome developed over the Lake District, with neighbouring domes centred over the NW Yorkshire Dales and the Alston Block. These ice domes are believed to have made relatively minor contributions to an ice sheet which was dominated by a major dome centred on the Scottish Highlands. Indeed, the lack of diamicts on the higher fells has been interpreted as evidence for nunataks, even at maximal ice sheet conditions.

In this model, Lake District ice is believed to have flowed radially outwards from the central mountain core. This is based on striations and the distribution of Borrowdale Volcanic Group rocks in the surrounding lowlands, with a lack of foreign erratics in the Lake District mountains. Erratic evidence also suggests that Southern Upland ice was deflected around the northern Lake District. To the east of the massif, drumlins mapped by Trotter (1929) and Hollingworth (1931) indicate convergence of former ice flow eastwards through the Stainmore Gap in the Pennines towards Holderness. However, the overall pattern of drumlins in the Vale of Eden indicates major ice flow in the *opposite* direction, towards Carlisle and the Solway Firth. Around Carlisle, ice flow directional indicators demonstrate convergence of this ice with east-moving Southern Upland ice to flow through the Tyne Gap, and with west-moving ice to flow towards the Solway Firth and Irish Sea (Figure 1.7).

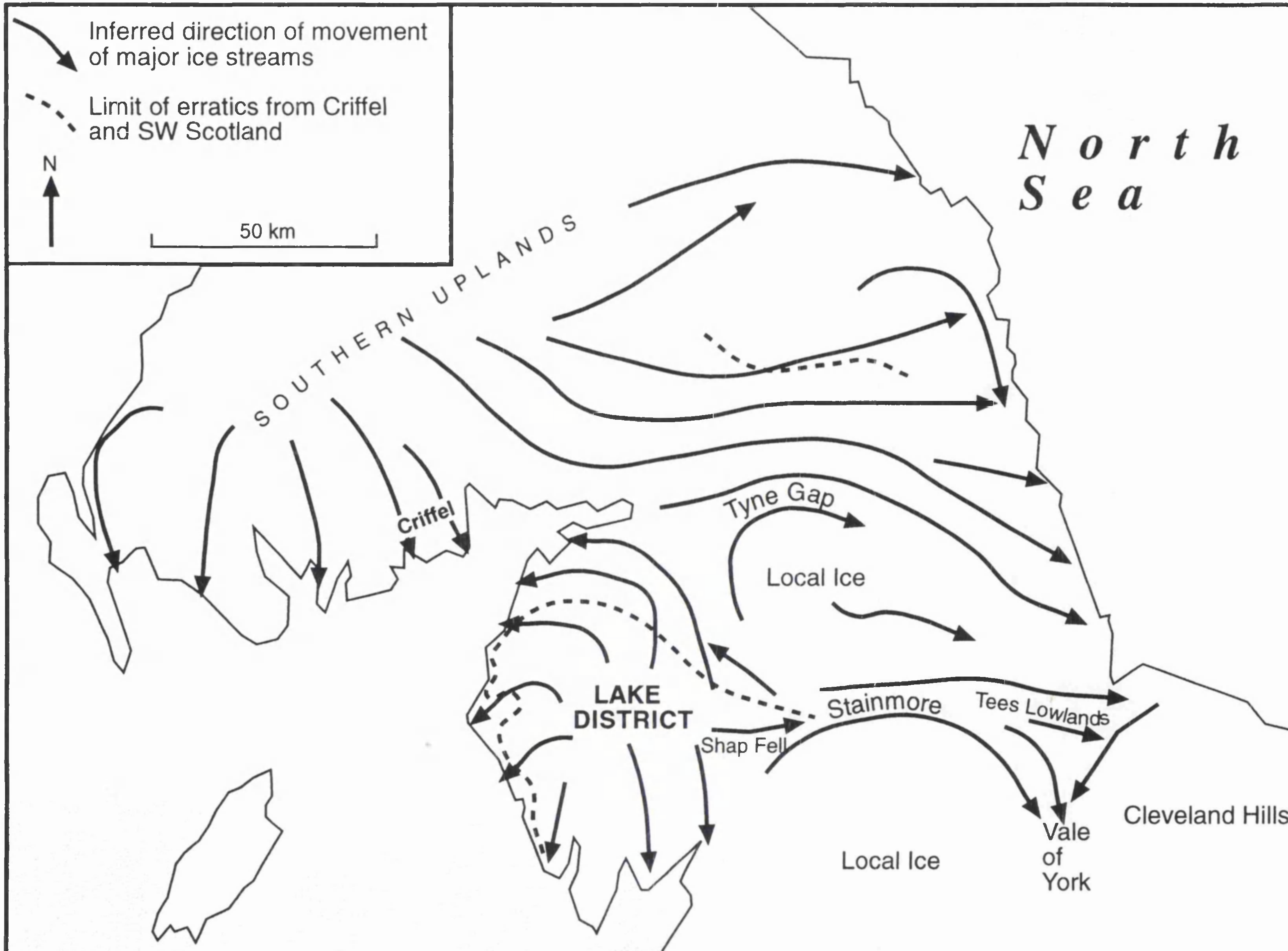


Figure 1.7 General pattern of ice flow directions in northern England.
After Mitchell and Clark (1994)

An alternative model for the last ice sheet in Cumbria is in the process of being developed (Mitchell and Clark, 1994). Whereas the established model features relatively minor ice domes over the Lake District, the NW Yorkshire Dales and the Alston Block, evidence is mounting to support the former existence of a large regional ice mass. This apparently had a linear ice divide which was independent of topography, extending from Dentdale northwestwards over 50 km to the Derwent Fells in the central Lake District. The ice divide is believed to have migrated northwards during glaciation/deglaciation, with a corresponding reduction in northerly ice flow catchment but with an extension of ice flow down the Rawthey valley and into Wensleydale (Mitchell, 1991, 1994; Mitchell and Clark, 1994).

The development of this model follows the identification of superimposed drumlin forms in the southern part of the Vale of Eden (Letzer, 1978, 1981, 1987; Whiteman, 1981) and the Western Pennines (Mitchell, 1991, 1994). These indicate different flow events during the course of the glaciation, and have allowed the identification of a linear ice divide which migrated over time. In addition, the distribution of Shap granite erratics can only be explained by the presence of a migrating ice divide over the area of the outcrop. This differs from the established model in which seemingly contradictory ice-directional indicators in certain areas were accommodated by postulating complex contemporary ice flow, with basal layers of the ice sheet able to move independently from the upper layers. However, recent advances in knowledge of ice sheet dynamics precludes such an explanation. A substantial thickness of ice in the region is also indicated by the identification of drumlins at altitudes in excess of 600 m OD (Mitchell, 1991).

Although the configuration and dynamics of the last ice sheet in Cumbria may be poorly understood, it is generally accepted that it decayed by areal stagnation. The sequence in the Solway lowlands, however, is considered to be more complex, with an upper till attributed to the 'Scottish Readvance' (e.g. Huddart, 1970, 1994). This upper till has also been interpreted as a glaciomarine mud drape associated with marine drawdown within the Irish Sea basin (Eyles and McCabe, 1989). Within the Lake District mountains, kame and kettle topography in the Keswick area has been interpreted by Boardman (1981) as evidence for stagnant ice in the valleys during deglaciation.

Nevertheless, well-defined moraines are present in several valleys only short distances downvalley of the Loch Lomond Stadial limits proposed by Sissons (1980a). Examples include those at Rosthwaite and Thornythwaite in Borrowdale, at Wythburn and at the northern foot of Kirkstone Pass (Clark, 1990; Clark and Wilson, 1994).

The age and significance of the moraines downvalley from the presently accepted limits of Loch Lomond Stadial palaeoglaciers have yet to be established. It may be that they represent evidence for more extensive Loch Lomond Stadial glaciers in these areas. Alternatively, these moraines may have been produced during stillstands and/or readvances which interrupted the decay of the last ice sheet. Although Pennington (1978) has stated that there is no biostratigraphical evidence for this, the high resolution Greenland ice cores (GRIP and GISP2) clearly reveal spikes in the oxygen isotope record which document short-lived climatic deteriorations (Figure 1.2). Whether these climatic oscillations had a significant impact on the margins of an otherwise decaying British ice sheet is unclear. Although a number of readvances have been proposed (Figure 1.8), relating these to the Greenland ice core record demands greater dating accuracy and precision than is available at present. Until then, the climatic significance of some of these events will remain uncertain and the possibility that they merely represent local readjustments of the ice margin cannot be ruled out. Assuming the latter *not* to be the case, the Wester Ross Readvance (Robinson and Ballantyne, 1979) is the only event other than the Loch Lomond Readvance which is potentially capable of explaining the Lake District moraines referred to above. The other readvances shown in Figure 1.8 occurred at times when the Lake District was presumably still largely submerged by ice whereas the moraines in question imply a restricted glaciation not substantially more extensive than the Loch Lomond Readvance.

1.4.5 The Loch Lomond Stadial glaciation (Loch Lomond Readvance)

Following the climatic amelioration of the Lateglacial Interstadial, the Loch Lomond Stadial (12.9–11.5 ka BP) witnessed renewed glaciation in the Lake District and other parts of upland Britain. The Lake District is assumed to have been characterised by an alpine style of glaciation during the Loch Lomond Stadial, with reconstructed glaciers emanating from corries and valley heads (Sissons, 1980a). The small size of many of

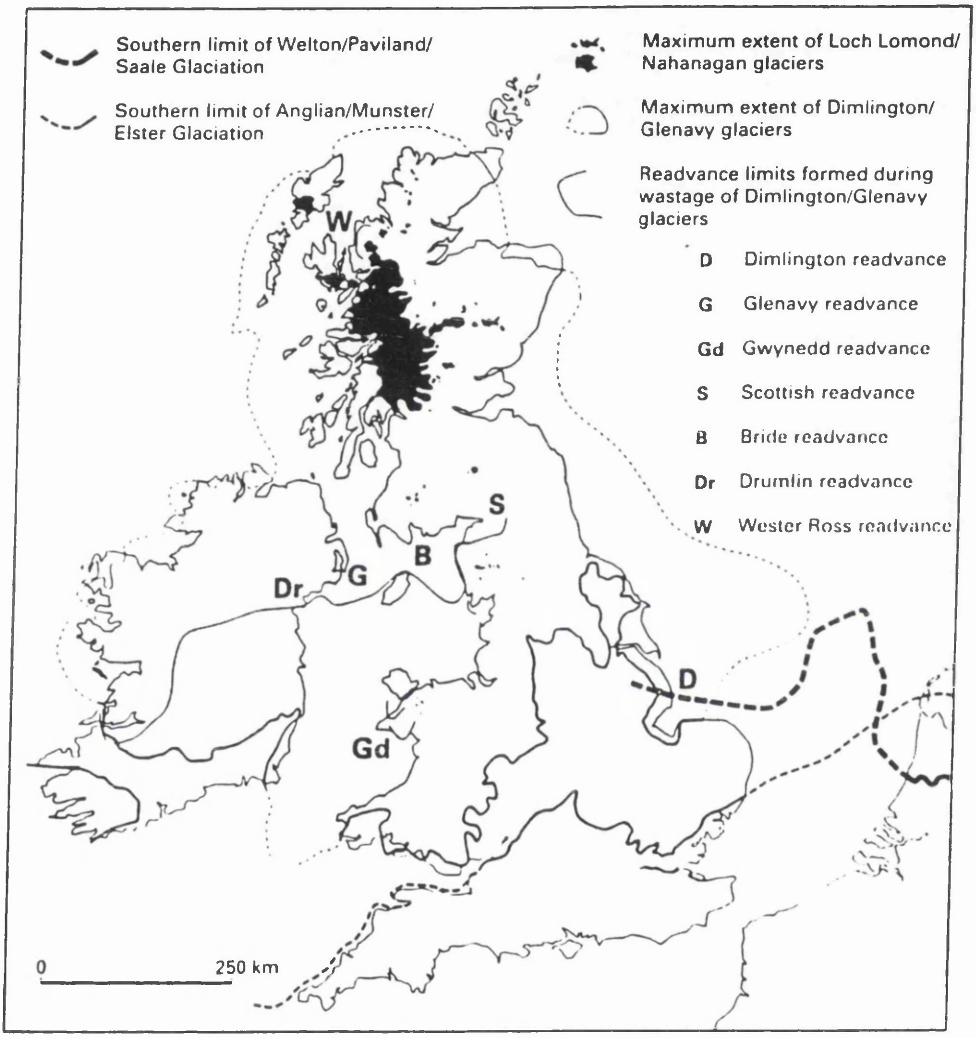


Figure 1.8 Glaciation limits and readvances in Great Britain and Ireland.

Source: Lowe and Walker (1997, p. 29)

these former glaciers has been interpreted as evidence that the Lake District was marginal for glaciation at this time, most probably due to low levels of precipitation (e.g. Sissons, 1980). The literature on this topic is reviewed in Section 2.4.

2

Literature review

2.1 INTRODUCTION

This research seeks to establish whether geomorphological evidence for Loch Lomond Stadial plateau icefields exists within the Lake District. It is thus appropriate to commence this chapter by reviewing the literature on the geomorphological impacts of contemporary and former plateau icefields (Section 2.2). This is followed in Section 2.3 by a consideration of the topoclimatic controls on plateau icefield development, an important topic given that direct geomorphological evidence for plateau icefields may be poorly developed or absent altogether. In such circumstances, reconstructions of plateau icefields will require an assessment of topographic and palaeoclimatic factors (Gellatly *et al.*, 1988). A review of Loch Lomond Readvance studies, with an emphasis on Lake District research, is presented in Section 2.4. This provides the basis for site selection since failure to account for plateau icefields may be evidenced by anomalously low ELA reconstructions (Chapter 3). Finally, a chapter summary is presented in Section 2.5.

2.2 GEOMORPHOLOGICAL IMPACT OF PLATEAU ICEFIELDS

2.2.1 Introduction

The identification of former plateau icefields in deglaciated areas, such as the Lake District, demands an appreciation of the geomorphological impact of contemporary examples. Investigations in north Norway provides the only substantial basis for this (e.g. Whalley *et al.*, 1981, 1995; Gordon *et al.*, 1987, 1988, 1995; Gellatly *et al.*, 1988, 1989; Rea and Whalley, 1996). It is not clear to what extent Norwegian plateau icefields may constitute appropriate analogues. Therefore, this section attempts to highlight those observations relating to the geomorphological activity of these plateau icefields which may have general applicability.

2.2.2 Geomorphological impacts

Research in north Norway has shown that some plateau icefields have a minimal geomorphic impact. In the Lyngen Peninsula (70°N, 20°E), for example, it has been inferred from field investigations that small plateau icefields at altitudes in excess of 1500 m are probably cold-based at present (Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988). Recent recession of the Jiek'kevárri, Bálgesvárri and Bredalsfjellet plateau icefields has revealed undisturbed blockfield and patterned ground (Whalley *et al.*, 1981; Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988) (Figure 2.1). There is no evidence for subglacial erosion on the main plateau surfaces, although localised erosion has been observed at the edge of an outlet glacier on the northern side of Bálgesvárri (Gordon *et al.*, 1987; Gellatly *et al.*, 1988). A trench excavated into the edge of the Bálgesvárri icefield in 1979 showed the margin to be frozen to its substrate, with basal layers free of debris (Whalley *et al.*, 1981). Furthermore, patterned ground was observed to extend beneath the glacier. Stable-isotope analyses support the interpretation that this plateau icefield is predominantly cold-based at present (Gordon *et al.*, 1988).

The preservation of periglacial phenomena beneath plateau icefields requires cold-based, non-erosive ice. It was believed, until fairly recently, that basal sliding does not occur under cold based conditions, a conclusion based on field observations and theoretical analyses (e.g. Goldthwaite, 1960; Holdsworth and Bull, 1970; Boulton, 1972; Hughes, 1973; Paterson, 1981). However, recent laboratory and theoretical investigations

Figure 2.1 Blockfield exposed by the recession of the Bredalsfjellet icefield, north Norway. *Photograph courtesy of W. B. Whalley.*

(Shreve, 1984; Fowler, 1986) have concluded that basal sliding processes should operate at subfreezing temperatures, albeit at very low rates. This has been confirmed by field observations (Echelmeyer and Wang, 1987). Nevertheless, there is a tendency for cold-based ice to protect the underlying substrate (with the exception of the margins, where locally impressive quarrying can occur) (Benn and D.J.A. Evans, 1998).

In plateau icefields, cold-based conditions are promoted by the dynamics of relatively thin, slow-moving ice, plus the impact of low mean annual air temperatures. The relationship between plateau icefield thickness and summit breadth is shown in Figure 2.2. This was calculated by Rea (1994a), who reasoned that a first approximation of plateau icefield profiles could be obtained by employing a parabolic equation used to estimate ice sheet profiles (Orowan, 1949; Nye, 1952):

$$h = (2\, \tau_b\, x / \rho\, g)^{1/2} \qquad (1)$$

where h is the ice thickness at the centre

τ_b is the basal shear stress

x is the distance from the margin to the ice divide

ρ is the density of ice and g is the acceleration due to gravity.

Actual thicknesses are likely to be even lower than those indicated in Figure 2.2 due to the draw-down effect of outlet glaciers (Rea, 1994a). The combination of thin ice and low bed slope angles means that ice velocities, and thus the generation of frictional heat, will be relatively low. The insulating effect of thin ice is also minimal, and thus basal ice temperatures will be strongly influenced by plateau air temperatures. Therefore, plateau altitude can be a key variable in determining basal thermal regimes, as is the case in Lyngen (Gordon *et al.*, 1987).

Investigations of contemporary plateau icefields in the Lyngen Peninsula illustrate the potential difficulties surrounding the reconstruction of former plateau icefields which were partly or wholly cold-based. In such circumstances, marginal meltwater channels produced during deglaciation may represent the only signs of glacial modification. These

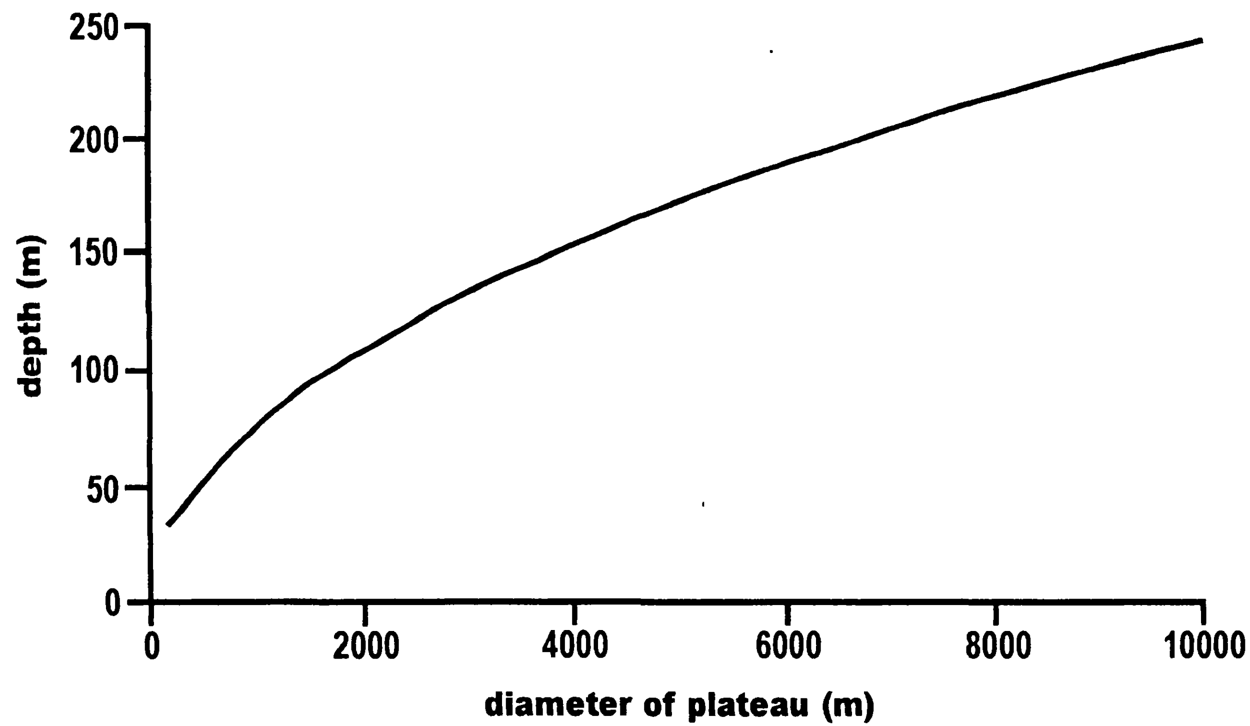


Figure 2.2 Theoretical relationship between plateau icefield depth and summit breadth. From Rea (1994b)

channels are often well-developed in sub-polar glaciers because meltwater is unable to penetrate to the frozen bed (Benn and D.J.A. Evans, 1998). For example, there are weathered landscapes preserved within the limits of late Wisconsinan ice caps in the central Canadian Arctic archipelago in which the only clear evidence for glacial modification are the ice-marginal meltwater channels produced during deglaciation (Dyke *et al.*, 1992; Dyke 1993). In NW Ellesmere Island, D.J.A. Evans (1988, 1990) has described how the wastage of former cold-based plateau icefields is documented by shallow ice-marginal meltwater channels which are incised into residuum and bedrock. Meltwater channels apart, these former plateau icefields appear to have had a minimal geomorphic impact; summits lack obvious glacial erosional features and are often characterised by well-developed patterned ground and tors (D.J.A. Evans, 1988, 1990). Nevertheless, it should be noted that lateral channels are absent from the margins of contemporary cold-based plateau icefields in Lyngen. This has been attributed to insufficient meltwater production for the development of channellised drainage (Gellatly *et al.*, 1988).

The boundaries of relict periglacial surfaces preserved beneath frozen patches of the Fennoscandian ice sheet have been investigated by Kleman and Borgström (1990, 1994). These patches, which are of pre-late Weichselian age, occur on low uplands within the core area of the Fennoscandian ice sheet and are surrounded by late Weichselian flutings. Boundaries are often sharp and erosional, with flutings truncating patterned ground. On the basis of three type areas, Kleman and Borgström (1994) have defined four thermal boundary landforms which they consider to be characteristic of the frozen patch environment (Figures 2.3 and 2.4). A *lateral sliding boundary* is well-defined (a sharp boundary or narrow transition), separating a zone of subglacial lineations (e.g. flutings) from a truncated relict surface. Where this takes the form of a ridge, which may be parallel or sub-parallel to a flow line, this is termed a *lateral shear moraine*. The proximal edge of a relict surface may be characterised by a *stoss-side moraine*, which is a transverse single- or multi-crested moraine ridge. The distal edge of a relict surface may be defined by a transverse *lee-side till scarp*, which is a till scarp oriented transversely to ice flow.

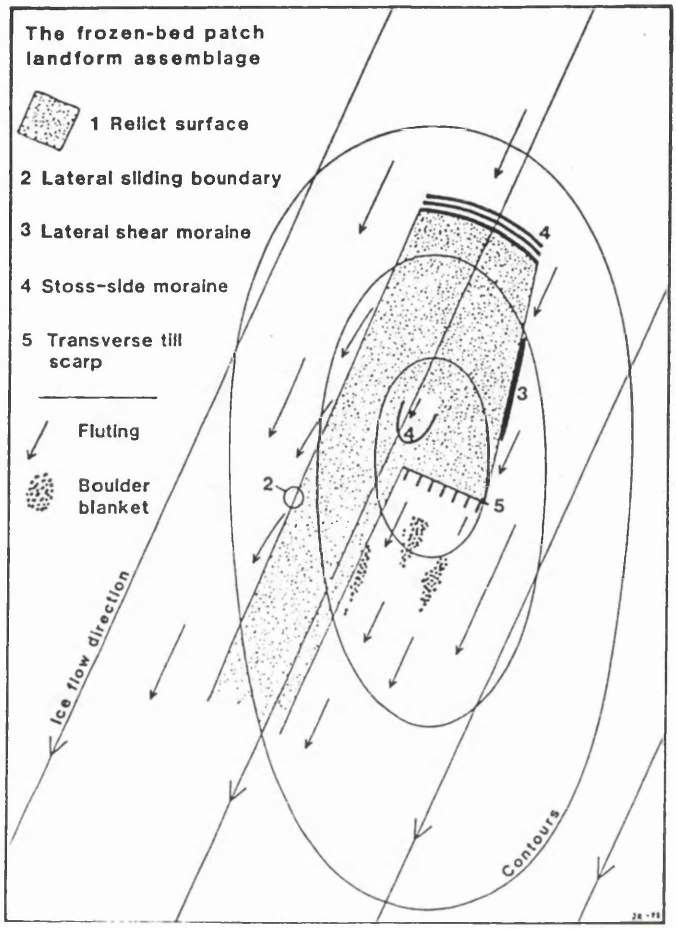


Figure 2.3 Thermal boundary landforms characteristic of the frozen-bed patch environment. See text for explanation.

Source: Kleman and Borgström (1994, p. 261)

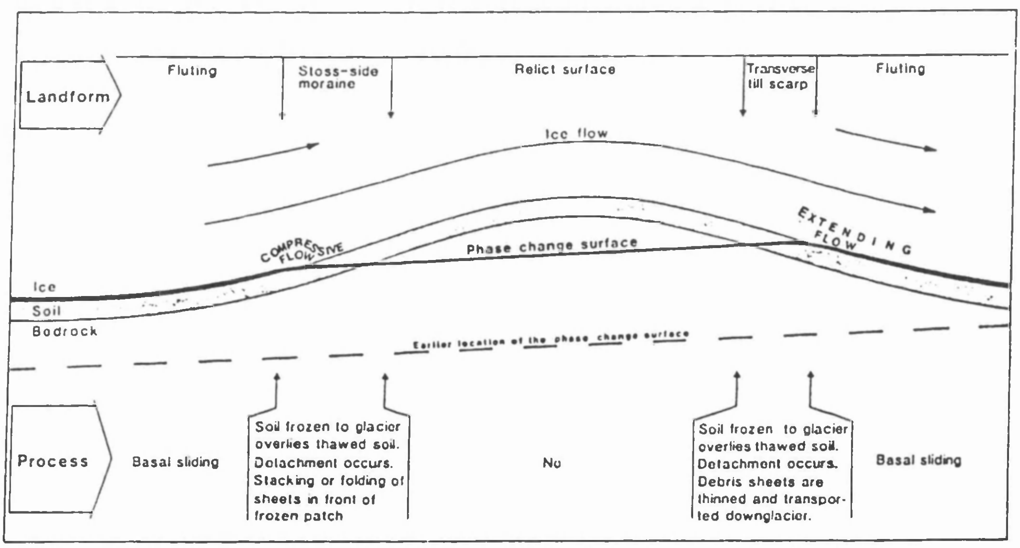


Figure 2.4 Geomorphological processes associated with a phase-change surface.

The phase-change surface (PCS) is the boundary between warm-based, sliding ice and cold-based protective ice. A PCS which rises over time creates two detachment zones where soil frozen to the glacier base overlies thawed soil or the soil/rock interface. Kleman and Borgström (1994) suggested that the stoss-side ridges may have formed by folding or stacking of debris sheets in front of frozen-bed patches, while the transverse till scarps mark the sites where debris sheets were entrained in an extending flow zone.

Source: Kleman and Borgström (1994, p. 262)

Kleman and Borgström (1994) suggested that, in addition to relict surfaces, the presence of these thermal boundary forms be taken as positive evidence for patchy thermal conditions during at least one interval of a glacial. However, the applicability of this work to the reconstruction of partly cold-based plateau icefields in deglaciated areas is uncertain. The lower basal shear stresses and radial flow associated with plateau icefields would substantially modify the landform assemblage shown in Figure 2.3. Transverse till scarps could conceivably still form, but lateral sliding boundaries, lateral shear moraines and stoss-side moraines would be absent.

Whilst the geomorphological impact of cold-based icefields on plateaux may be restricted to the development of marginal meltwater channels, a former warm-based plateau icefield may be evidenced by landforms of subglacial erosion, moraines and meltwater channels. Due to the combination of low ice velocities and low basal shear stresses, the erosional impact of thin (< 200m), warm-based plateau icefields on sound bedrock is likely to be limited to surface scouring and moulding (Gellatly *et al.*, 1988). Recent recession of the Øksfjordjøkelen plateau icefield in north Norway, for example, has revealed a forefield which predominantly comprises an inner zone of ice scoured bedrock and an outer zone, a few tens of metres wide, with blockfield-derived boulder moraines (Gellatly *et al.*, 1988).

Subglacial abrasion of bedrock requires the presence of basal debris, which may occur when an advancing plateau icefield margin incorporates unconsolidated debris (e.g. blockfields). For abrasion to be effective, basal debris must be replaced. Given that the input of supraglacial debris to a plateau icefield will be minimal at best, abrasional tools must be derived subglacially by quarrying processes. The recently deglaciated foreland of a small plateau-terminating outlet of the Øksfjordjøkelen icefield shows evidence of extensive abrasion and quarrying, including the development of rock steps (Rea, 1994a, b; Rea and Whalley, 1996). Nevertheless, modelling of flow-induced stresses for the inferred ice dynamics of the outlet glacier has shown that it is highly unlikely that the small rocksteps were formed by lee side fracturing mechanisms (Rea, 1994b). Instead, it appears that quarrying was facilitated at this location by a favourable rock structure.

Intersecting weathered joints produced blocks which could be removed from the crest of an up-glacier dipping rock step.

It is clear, therefore, that the overall erosional impact of a warm-based plateau icefield will reflect a number of factors. Apart from those locations where rock structure is conducive to quarrying processes, it is likely that the erosional impact of wet-based plateau icefields may be limited to surface scouring and moulding. The effects are likely to be most pronounced where the bed steepens towards the valley head (Rea, pers. comm., 1997). In some situations, perhaps where the coverage by warm-based ice is of limited duration, erosion may be limited to a clearing out of pre-existing regolith and blockfield (Gellatly *et al.*, 1988).

The distinction between wet- and cold-based plateau icefields, although convenient, represents a simplification of reality. For example, the basal thermal regime of the Øksfjordjøkelen plateau icefield is complex at present (Gellatly *et al.*, 1988; Rea and Whalley, 1996). In some areas, the ice is at pressure melting point and actively sliding over its bed. Elsewhere, the retreat of protective, cold-based ice has revealed blockfield and patterned ground. Given that basal thermal regime is determined by both environmental and internal dynamic factors (which are themselves interrelated), it should be expected that icefields may be polythermal.

The geomorphological impact of warm-based plateau icefields is not limited to erosion; moraines may develop at their margins. For example, a substantial Little Ice Age moraine has developed at a small southern outlet of Øksfjordjøkelen (Rea and Whalley, 1994). Due to the paucity or absence of supraglacial debris inputs, moraine development at plateau icefield margins requires debris to be incorporated subglacially. As outlined above, the erosional impact of warm-based plateau icefields on sound bedrock is likely to be very limited. Accordingly, the availability of pre-existing material which can be readily entrained, such as blockfield, may be critical in moraine development.

Whereas moraine development on the plateau margins of icefields is likely to be debris-limited, considerably more substantial moraines may be associated with outlet glaciers which descend into the surrounding valleys where their margins become sediment traps for extraglacial debris. In the Troms-Finnmark area, for example, plateau icefield outlet glaciers have produced latero-frontal moraines (Rea *et al.*, in prep.). These moraines comprise material which has been incorporated and transported along the valley, both subglacially and from valley sides. The presence of subglacial material indicates that these outlet glaciers (or part of them) are above PMP and sliding over their beds (Rea *et al.*, in prep.).

The geomorphological impact of former plateau icefield outlet glaciers in northwest Ellesmere Island has been emphasised by D.J.A. Evans (1988, 1990). In fjord/trough ice-marginal settings, considerable thicknesses of pre-existing sediments were available for direct glacial erosion, transport and deposition. Sediment deposited by ice-marginal and proglacial streams accumulated as thick valley-bottom alluvium and alluvial fans in valley systems surrounding plateau icefields. Some glaciers proglacially thrust this sediment, although in other situations glaciers advanced over it without disturbance (D.J.A. Evans, 1989a, b). The availability of sediment in the main U-shaped valleys contrasts with the situation where plateaux are surrounded by undulating bedrock lowlands, where a lack of thick sequences of pre-existing sediments, plus paucity of supraglacial debris from the source area, restricted moraine development (D.J.A. Evans, 1988, 1990).

2.2.3 Summary

Investigations in contemporary glacial environments highlight the potential difficulties surrounding the identification of former plateau icefields in deglaciated areas. Where a plateau icefield is cold-based, the geomorphological impact may be minimal, with preservation of blockfield and patterned ground (Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988). Evidence for wet-based plateau icefields on summits is likely to be limited to surface scouring and moulding. Although Rea (1994b) observed that a recently deglaciated foreland of a small plateau-terminating outlet of the Øksfjordjøkelen icefield is characterised by extensive quarrying and abrasion, he concluded that this was facilitated by the nature of the discontinuous rock mass, with favourably inclined, pre-weathered jointing. Once removed, these blocks provide ideal abrasion tools. Plateau icefield outlet glaciers which descend into surrounding valleys may have a more substantial geomorphological impact, since not only do their margins become sediment traps for extraglacial debris but they may also entrain valley-floor alluvium (D.J.A. Evans, 1988, 1990). By way of contrast, moraine development on plateaux is likely to be debris-limited, with supraglacially-derived debris minimal or non-existent. In such circumstances, pre-existing debris (e.g. blockfield) may constitute an important source of debris for moraine development (as well as abrasional tools) since the potential for a warm-based plateau icefield to erode its bed is likely to be limited due to low basal shear stresses and slow moving ice (Gellatly *et al.*, 1988; Rea 1994a, b; Rea and Whalley, 1996; Rea *et al.*, in prep.).

2.3 TOPOCLIMATES AND GLACIATION STYLES

2.3.1 Introduction

The identification of former plateau icefields in deglaciated areas may be problematical, particularly where they were cold-based. By itself, the absence of direct geomorphological evidence is inconclusive (Gordon *et al.*, 1987; Gellatly *et al.*, 1988). Reconstruction of plateau icefields in such circumstances will require an assessment of topographic and palaeoclimatic factors. This section presents a brief overview of topoclimates and glaciation styles, with a particular emphasis on the topoclimatic controls on plateau icefield development.

2.3.2 Topoclimates

The controlling factors in the initiation and maintenance of a restricted glaciation are climatically and topographically related, and this interaction determines the style of glaciation. The two most important climatic factors in glacier development are the rate of snow accumulation, a function of both temperature and precipitation, and the rate of ablation, largely a function of temperature. A glacier will only develop where snow can collect and survive. This is above the snowline, where snowfall accumulation is balanced by ablation. The climatic snowline in any region or site is strongly controlled by latitude, altitude and continentality.

At the local scale, topographical effects may result in significant variations in snowfall accumulation and ablation. These variations can be critical for the development of glaciers in marginal situations (I.S. Evans, 1969). The effect of topography on climate is to create topoclimates, which are primarily manifestations of slope angle, aspect and horizon (Barry, 1992). These topoclimates result in the local snowline displaying an irregular surface, which is termed the orographic snowline (e.g. Flint, 1971). Topography, therefore, acts both to modify and complicate the macroclimatic parameters that determine the altitude of the climatic snowline (Flint, 1971) (Figure 2.5).

In the middle and high latitudes of the northern hemisphere, the best known examples of orographic snowline lowering occurs in north easterly facing cirques in the vicinity of the regional snowline. Not only are these the coolest sites (with lowest ablation), but they occupy the lee side slopes in areas of prevailing south-westerly winds and, as such, their

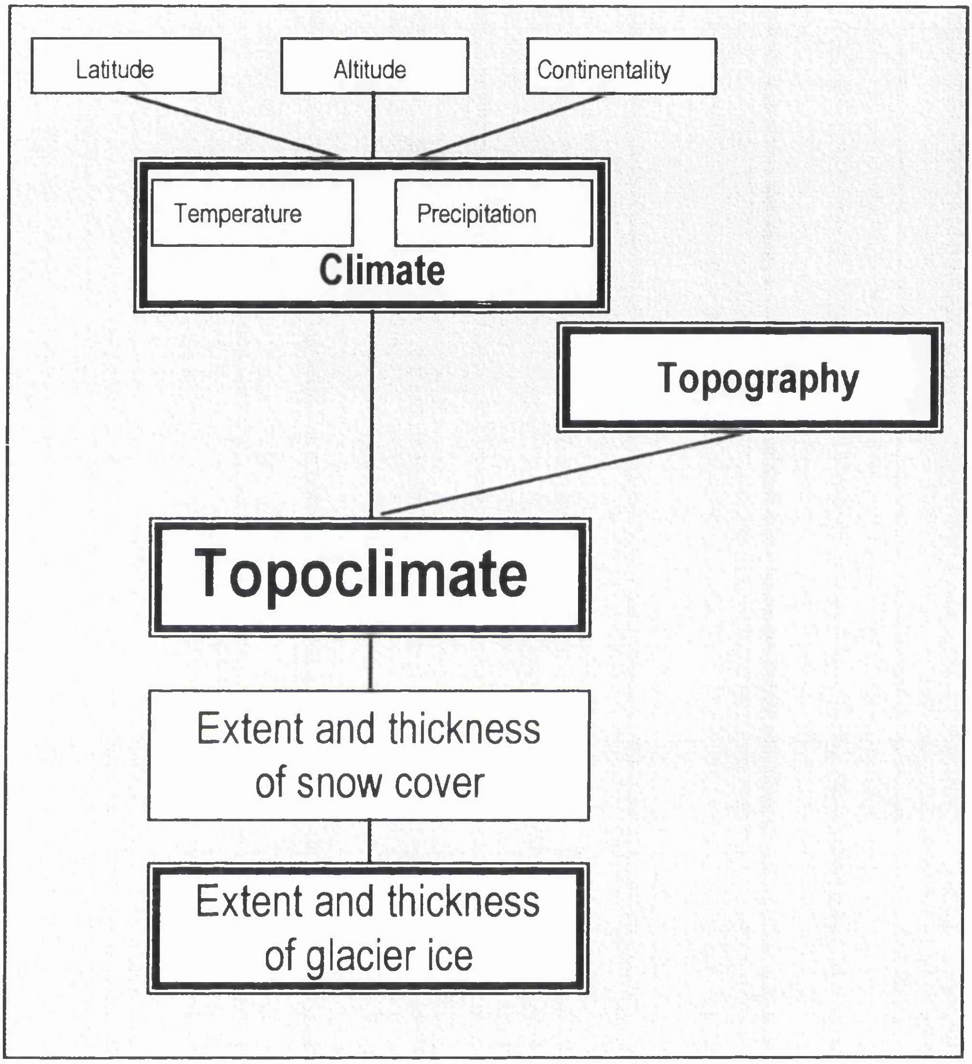


Figure 2.5 Environmental factors which are important in determining the extent and thickness of glacier ice.
After Summerfield (1991, Figure 11.3, p. 262)

topoclimates may be characterised by relatively high levels of snow accumulation through the operation of winds which transfer snow from adjacent higher ground, particularly plateaux. In the Polar Urals, Dolgushin (1961) has described the occurrence of cirque glaciers in an area where the climatic snowline lies above the highest summits.

The accumulation-ablation balance on an exposed summit differs from that of a sheltered, lee side cirque. A higher degree of exposure and consequent greater wind strength leads to the direct transfer of momentum to ice crystals, which results in the drifting of snow over the edge of the summit where it accumulates in lee side cirques or hollows. Thus, it is quite normal for amounts of accumulated snowfall on the summits to be less than that of lee side locations. Furthermore, with little or no protection from solar radiation, ablation on summits will be correspondingly higher than the cool, sheltered north-easterly orientated cirque (in the Northern Hemisphere), although the significance of this will partly depend on cloudiness (Manley, 1955; Barry, 1992). Thus, plateaux are often less favourable for snowfall accumulation and preservation than north-easterly facing cirques. For example, the broad plateau in NW Iceland which once supported the Glamajökull icefield is now ice free, although small cirque glaciers occur at lower altitudes in the lee side (John, 1976).

In his investigation into the topoclimatic controls on European icefields, Manley (1955) demonstrated that there is a close, non-linear relationship between summit breadth and altitude above the regional firn line. Essentially, as summit breadth decreases, the altitude above the firn line that a summit must attain if it is to support an icefield increases. Manley arbitrarily defined summit breadth as the horizontal distance between the contours 30 m below the summit in the direction of the prevailing winds.

On the basis of his observations, Manley (1955) considered that, in the middle latitudes of the northern hemisphere, a summit 1000 m broad is likely to be permanently snow covered if it rises about 200 m above the regional firn line (Figure 2.6). If summit width is reduced to 300 m, then it would have to rise 400 m above the regional firn line. If a summit is only 100 m broad it will only retain a permanent ice cover if it attains a height of around 600–700 m above the regional firn line.

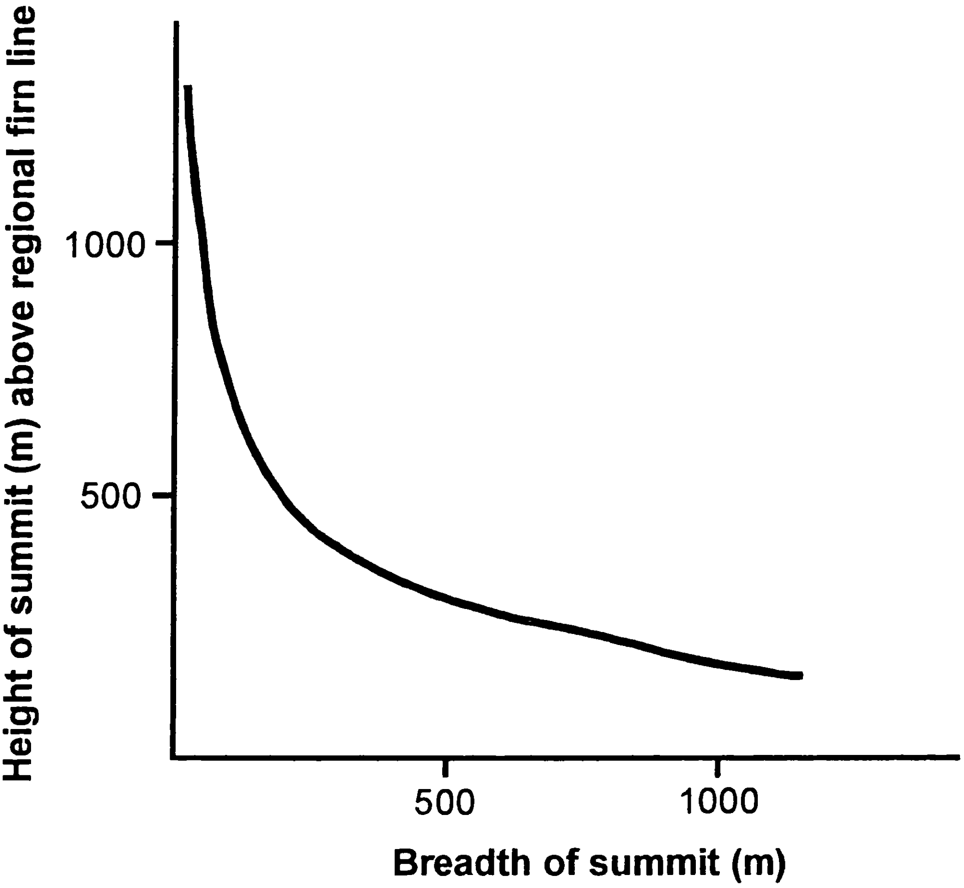


Figure 2.6 Topoclimatic controls on European icefields: relationship between summit breadth and altitude above regional firn line.
After Manley (1955).

In highly dissected alpine uplands, summit breadths are too narrow to permit any substantial snow cover. Steep mountain slopes may remain ice-free, and snow accumulation and glacier development is likely to be restricted to cirques and valleys. Indeed, highly dissected topography with precipitous slopes may well inhibit glacierization (Sugden and John, 1976).

It is quite clear, therefore, that topography, and its interaction with climate, can produce topoclimates which are highly significant for glacier development in marginal areas. Where the climatic snowline lies above the summits, the topoclimates of north-easterly facing cirques may be sufficiently favourable for the development of glaciers. Cirques with southerly aspects have less favourable topoclimates, and accordingly glaciers will not normally develop there as long as the climatic snowline lies above the summits. Where a glacier *has* developed in a south-facing cirque, and that cirque is not in receipt of avalanched snow or ice, then the snowline must lie below the summits. Whether an icefield develops on the summit itself, however, will depend on the breadth of the summit in relation to its altitude above the snowline; the narrower the summit, the higher above the climatic snowline it needs to be to be permanently covered by snow or ice.

2.3.3 Glacier morphologies

The range of glacier morphologies which will occur in any particular area will be determined by the interaction between topographic and climatic factors. This is illustrated by the work of D.J.A. Evans (1988, 1990) in the Phillips Inlet and Wootton Peninsula areas of north-west Ellesmere Island, where the glaciation level descends from 1100 m in the south of the field area, to sea level on the Alfred Ernest Ice Shelf in the north. He observed that the present day ice cover is largely controlled by topoclimate, and that the three bedrock-controlled physiographic zones which he defined are each characterised by different glacier morphologies. The first zone is an undulating plateau, less than 900 m asl., and hosts icefields with a few outlet lobes. It is a relatively ice-free zone because it is located inland, is of low elevation, and receives little precipitation. The second zone is quite different from the first, being a deeply fretted cirque terrain with many summits greater than 1200 m asl.. Accordingly, the style of glaciation differs, being characterised by cirque and trunk glaciers which terminate in or near the sea. This zone is

80% ice-covered, attributable to a combination of high elevations and greater precipitation. The third and final zone is a gently sloping plateau, decreasing from 300 m asl to sea level. This zone hosts the Alert Point ice mass, whose northernmost margins form an ice shelf.

Detailed geomorphological and sedimentological mapping enabled D.J.A. Evans (1988) to reconstruct glacier morphologies associated with the last glacial maximum in the area. He showed that trunk glaciers were created by the expansion and coalescence of piedmont lobes within the main valleys, with deglaciation resulting in the recession of trunk glaciers and reversion to piedmont lobes and then to plateau ice caps with no outlet lobes. This clearly illustrates the fact that an increase or decrease in glacierization can bring about a fundamental change in the topographical setting of a glacier and consequently the glacier morphologies present (Kuhle, 1988).

The work of Gellatly *et al.* (1986) and Gordon *et al.* (1987) in the Lyngen Peninsula of north Norway has also illustrated how the interaction between topography and climate determines the glaciation style of the area. On the western side of the peninsula, an alpine type of landscape changes eastwards into one of selective linear erosion, where plateau surface remnants survive between deeply dissected glacial troughs. The plateaux above 1600 m support icefields, and the lower plateau areas also held permanent snow and icefields at some time in the past (Figure 2.7). Precipitous cliffs more than 1000 m high separate the icefields from the valley glaciers below. These valley glaciers appear to lie entirely within the ablation zone, and can only exist because they receive a supply of avalanched ice from above. In other words, the occurrence of the valley glaciers is only possible because of the existence of plateau icefields above. Thus, although not contiguous with the icefields above, these valley glaciers are part of the icefield system and cannot be considered separately.

In the Lyngen Peninsula, topoclimates have created an irregular pattern of glaciation. Gellatly *et al.* (1986) observed that throughout southern Lyngen there are many ice-free valleys adjacent to valleys with glaciers, yet both are at similar altitudes. Topography, more than climate, controls the distribution of glaciers. They suggest that other areas,

Figure 2.7 Edge of the Jiek'kevarri plateau icefield, Lyngen Peninsula, north Norway
Photograph courtesy of W.B. Whalley

long since deglaciated, may also have been dominated by topographic controls. If this was so, they argue, then the resulting pattern of glaciation could have been quite irregular.

2.3.4 Summary

The foregoing is highly relevant to understanding glacier extents and morphologies associated with the Loch Lomond Readvance in the Lake District. It should be possible, prior to glacial geomorphological mapping, to identify on a topographic map potential sites for the initiation of glaciation which vary in their suitability. The higher, north-easterly facing cirques in the lee of broad plateaux will have been the most favoured sites for glacier development, and indeed the only ones to have been glacierized in a marginal event where the regional snowline was above the summits. In terms of identifying the most favoured location for an icefield, Manley (1955) suggests that this would have been a function of its breadth and its altitude in relation to the climatic snowline. It would seem reasonable to suggest from Manley's work that, for the Lake District, summit breadths could be plotted against their altitudes. The broadest and highest summits would be worthy of initial attention. This is developed in Chapter 3 (Methods).

2.4 THE LOCH LOMOND READVANCE IN THE LAKE DISTRICT

2.4.1 Introduction

Visually impressive glacigenic landform assemblages can be found in many of the corries and valley heads of the Lake District. These have long attracted the attention of geologists and geomorphologists working in the area. Marr (1895) made reference to what he considered to be a fine series of moraines below Angle Tarn at the head of Langstrath. Ward (1873) suggested that upper Ennerdale has *"...perhaps the most complete set of moraines to be found in the district."* Other workers who have commented on these landforms include Upham (1898), Raistrick (1926), Hay (1942, 1944), Manley (1959), Walker (1965), Boardman (1981), Sissons (1980a), and Oxford (1985, 1994). In most cases these distinct glacigenic landform assemblages have been attributed to the Loch Lomond Readvance (Manley, 1959; Walker, 1965, Pennington, 1978; Sissons, 1980a).

The Loch Lomond Readvance corresponds to a long-recognised and widespread readvance of glaciers in upland Britain (e.g. Geikie, 1863). During the late nineteenth century, the distinctive glacigenic landforms which had been recognised in many Highland valleys were attributed to a final phase of activity at the end of the last glacial (e.g. Geikie, 1864; Young, 1864; Jamieson, 1874).

The area after which the Loch Lomond Readvance takes its name is the south-eastern basin of Loch Lomond in west central Scotland. Evidence for a glacial readvance in the area was first recognised by Jack (1875), and placed in a regional context by Simpson (1928a, b, 1933), who identified the ice marginal positions of these palaeoglaciers along the Highland border west of the River Tay. The distinctive nature of these glacigenic landforms and sediments associated with this restricted glaciation enabled him to map the margins of the piedmont glacier which occupied the Loch Lomond basin. Subsequently, the glaciers of this period became known as the Loch Lomond Readvance. The use of the term 'readvance' reflected the belief that the last ice sheet did not completely waste away before this renewed episode of glaciation. Although it has since been argued more recently that complete deglaciation occurred prior to the Loch Lomond Stadial (e.g.

Sissons, 1972, 1976; Sissons and Grant, 1972), the evidence for this is at best equivocal (Sutherland, 1984).

This section reviews the literature on the Loch Lomond Readvance, with a particular emphasis on Lake District reconstructions. Following a general overview of the conventions that have been adopted in delimiting these former ice masses at maximal extents (Section 2.4.2), consideration is given to Loch Lomond Readvance studies in the Lake District (Section 2.4.3).

2.4.2 Delimiting Loch Lomond Stadial glaciers

In Britain, many workers have derived palaeoclimatic inferences from the reconstructed surface profiles of Loch Lomond Readvance glaciers (e.g. Sissons, 1972, 1973a, b, 1974, 1975, 1977a, b, c, 1978, 1979a, b, c, 1980; Sissons and Grant, 1972; Thompson, 1972; Gray and Brooks, 1972; Sissons, Lowe, Thompson and Walker, 1973; Sissons and Sutherland, 1976; Gray, 1975; Ballantyne and Wain-Hobson, 1980; Cornish, 1981; Gray, 1982; Ballantyne, 1989; Mitchell, 1991, 1996; Shakesby and Matthews, 1993). This is the most recent glaciation to have affected upland Britain and there is a general consensus within the literature that the geomorphological record is both clear and complete enough to have facilitated accurate reconstruction of these former ice masses at maximal extents (e.g. Lowe and Walker, 1984, p.27; Ballantyne and Harris, 1994, p.18).

A number of lines of geomorphological evidence have been used, usually in combination, to delimit the maximal downvalley extents of Loch Lomond Readvance palaeoglaciers. These include end and lateral moraines, hummocky moraine, drift limits, boulder limits, outwash spreads, and weathering contrasts (Price, 1983; Sissons, 1983; Sutherland, 1984; Gray and Coxon, 1991) (Figure 2.8). According to a review by Sutherland (1984), the majority of Loch Lomond Readvance glaciers have had part of their lower portions delineated by terminal or lateral moraines. In Snowdonia, for example, 30 out of the 35 glacial limits described by Gray (1982) utilise end or lateral moraines. On the other hand, the limit of the Gaick ice cap, as reconstructed by Sissons (1974), is only marked by end moraines for less than 15 km of the total ice marginal length of 180 km. Boulder limits occur where there is an abrupt limit to an arcuate spread of boulders, contrasting with

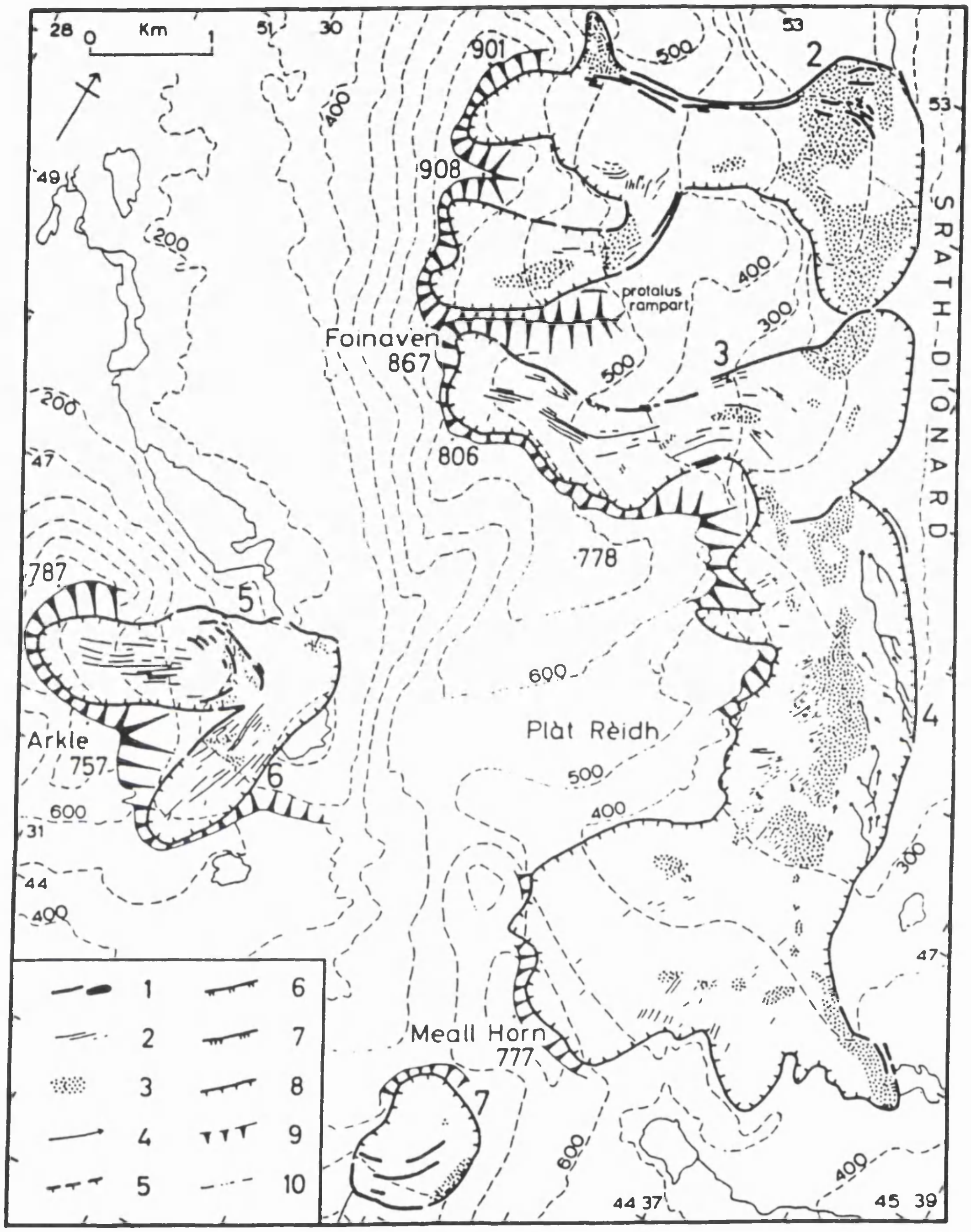


Figure 2.8 Loch Lomond Readvance glaciers in the vicinity of Foinaven, Arkle and Meall Horn, northern Scotland.

Symbols are as follows. 1, End moraines; 2, Flutings and other linear ridges not firmly identifiable as end moraines; 3, Hummocky moraine; 4, Meltwater channels; 5; Former ice limits inferred from boulder spread; 6, Former ice limit inferred from periglacial features; 7, Former ice limit based on limit of hummocky moraine or on limit of drift sheet; 8, Interpolated ice limit; 9, Very steep slopes of glacier source areas; 10, Contours on land and former glacier surfaces, interval 100 m.

Source: Sissons (1977b, p. 47)

either a boulder-free area outside the limit or with very different boulder lithologies. For example, in north-west Scotland, Sissons (1977a) reported spreads of white Cambrian quartzite boulders which terminate abruptly, contrasting sharply with the red Torridonian Sandstone bedrock beyond.

2.4.2.1 Hummocky moraine

Many former ice-margins have been defined on the basis of the downvalley extents of 'fresh' hummocky moraine, an all-embracing term which refers to *"...a highly irregular terrain comprising a large number of mounds that are often steep–sided and usually strewn with numerous boulders"* (Sissons, 1974, p.95). Manley (1959) was probably the first researcher to define Loch Lomond Readvance ice-margins in this way. Due to their chaotic appearance and close association with lateral moraines near the valley floor, Manley considered that hummocky moraine in the valleys heads of the Lake District represented evidence for the *in situ* stagnation of the terminal zones of Loch Lomond Readvance glaciers. He hypothesised that this may have followed topographically-induced disconnection of the lower (valley) portions of the former glaciers from their corrie sources, with corrie glaciers remaining active for a period thereafter. Additional support for their use as ice-marginal indicators, argued Manley, was provided by their coherent spatial distribution, occurring at lowest altitudes in those valleys which would have been most favourable for glacier development (Section 2.4.3)

As an aid to ice-marginal delineation, hummocky moraine has been used most extensively in Scotland, particularly during the 1970s (e.g. Sissons, 1974, 1977b) (Figure 2.9). The literature at that time tended to emphasise the often abrupt downvalley termination of hummocky moraine. In mapping, no attempt was made to distinguish individual elements of hummocky moraine due to the morphological complexity of these glacigenic landform assemblages and the paucity of sedimentological information (Price, 1983). Hummocky moraine was believed to be the product of areal stagnation of Loch Lomond Stadial glaciers at or near maximal extents in response to rapid thermal amelioration (Coope, 1977a, b).

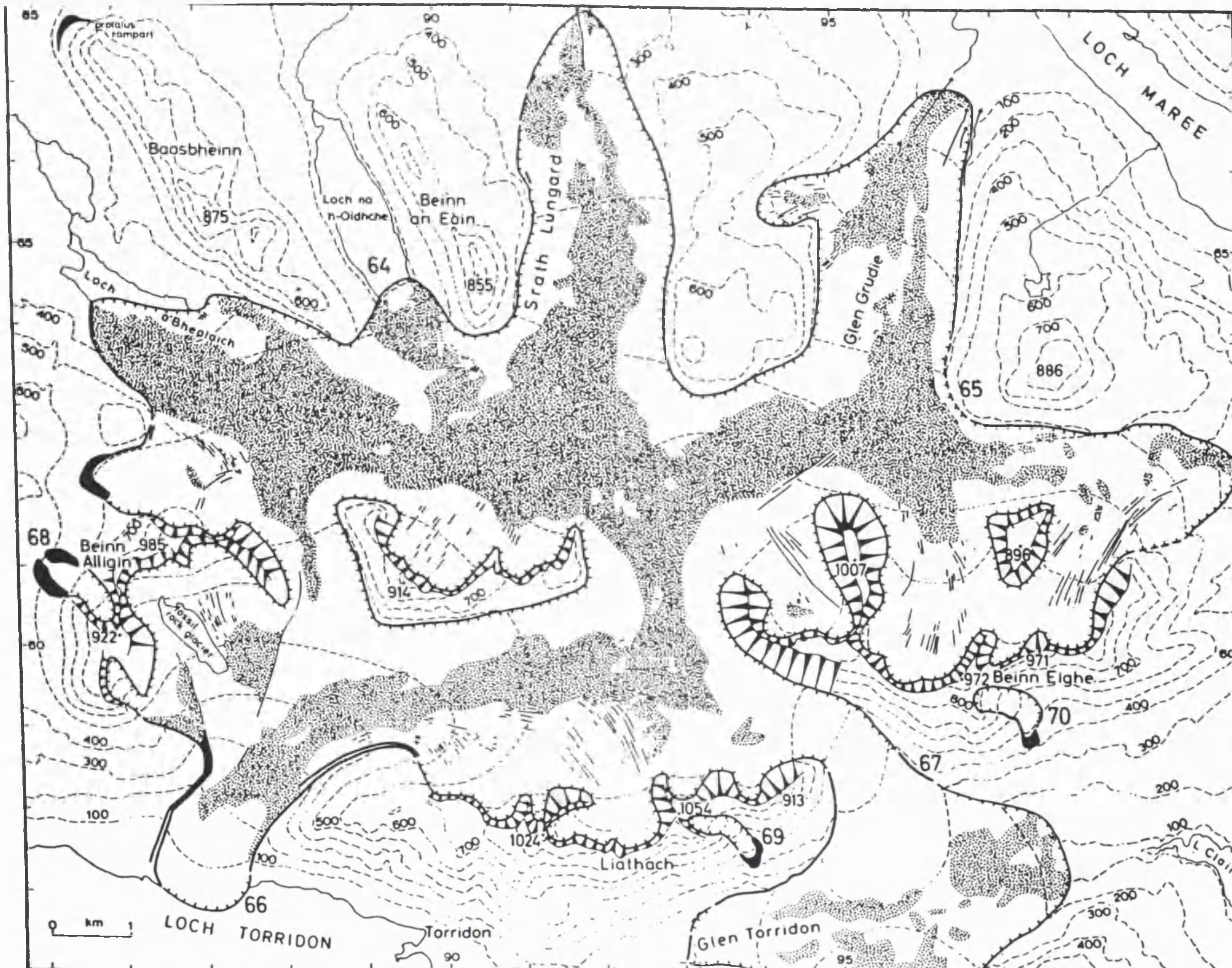


Figure 2.9 Loch Lomond Readvance glaciers between Loch Torridon and Loch Maree, northern Scotland. Symbols as for Figure 2.8.
Source: Sissons (1977b, p. 52)

Recent investigations have shown that hummocky moraine is polygenetic, and that many glaciers actively backwasted towards their sources (e.g. Benn, 1990; Bennett, 1991). These findings do not in themselves undermine previous reconstructions, which focused on the delineation of ice margins at maximal extent, although they do potentially provide a test as to their validity and may provide valuable information regarding centres of decay (Bennett, 1991; Bennett and Boulton, 1993a, b). This is particularly relevant in the context of the present research, where the former existence of plateau icefields may be inferred through detailed reconstructions of ice-margins during deglaciation.

Investigations by Benn (1990) and Bennett (1991) have demonstrated that hummocky moraine may comprise ice-marginal recessional moraines, subglacial moraines (flutings, drumlins), and localised ice-stagnation topography. Recessional moraines are typically sharp-crested ridges and chains of hummocks. Many ridges have an undulating or beaded long profile (particularly close to the valley floor), conveying a distinctly hummocky appearance. In plan, many of these moraines form converging cross-valley pairs, with similar sizes and gradients (Figure 2.10). Bifurcations and ridge asymmetry may be present, which are diagnostic properties of ice-marginal moraines (Bennett, 1991, 1994). Although linear elements within hummocky moraine had previously been recognised, they were interpreted as crevasse infills (e.g. Sissons, 1967). However, crevasse orientations in modern glaciers splay downvalley towards valley sides, in contrast with the arrangement in Skye and elsewhere (Benn, 1990). Finally, at the largest scale, the pattern of retreat revealed by these ice-marginal moraines may possess a regional coherency, with active-retreat towards decay centres (where these existed).

Many of these transverse ridges which have been interpreted as recessional moraines are composed of loose, rubbly diamictons, interbedded with silts, sands and gravels, interpreted as subaerial debris flow deposits (flow tills) and water lain sediments respectively (Benn, 1990). Clast analysis indicates that in many places subglacially-entrained debris was the dominant component. That these sediments were considerably reworked during deglaciation is implied by complex sedimentary geometries, and glaciotectonic activity is evidenced by the occurrence of steeply-dipping and folded diamict units (Benn, 1990, 1992).

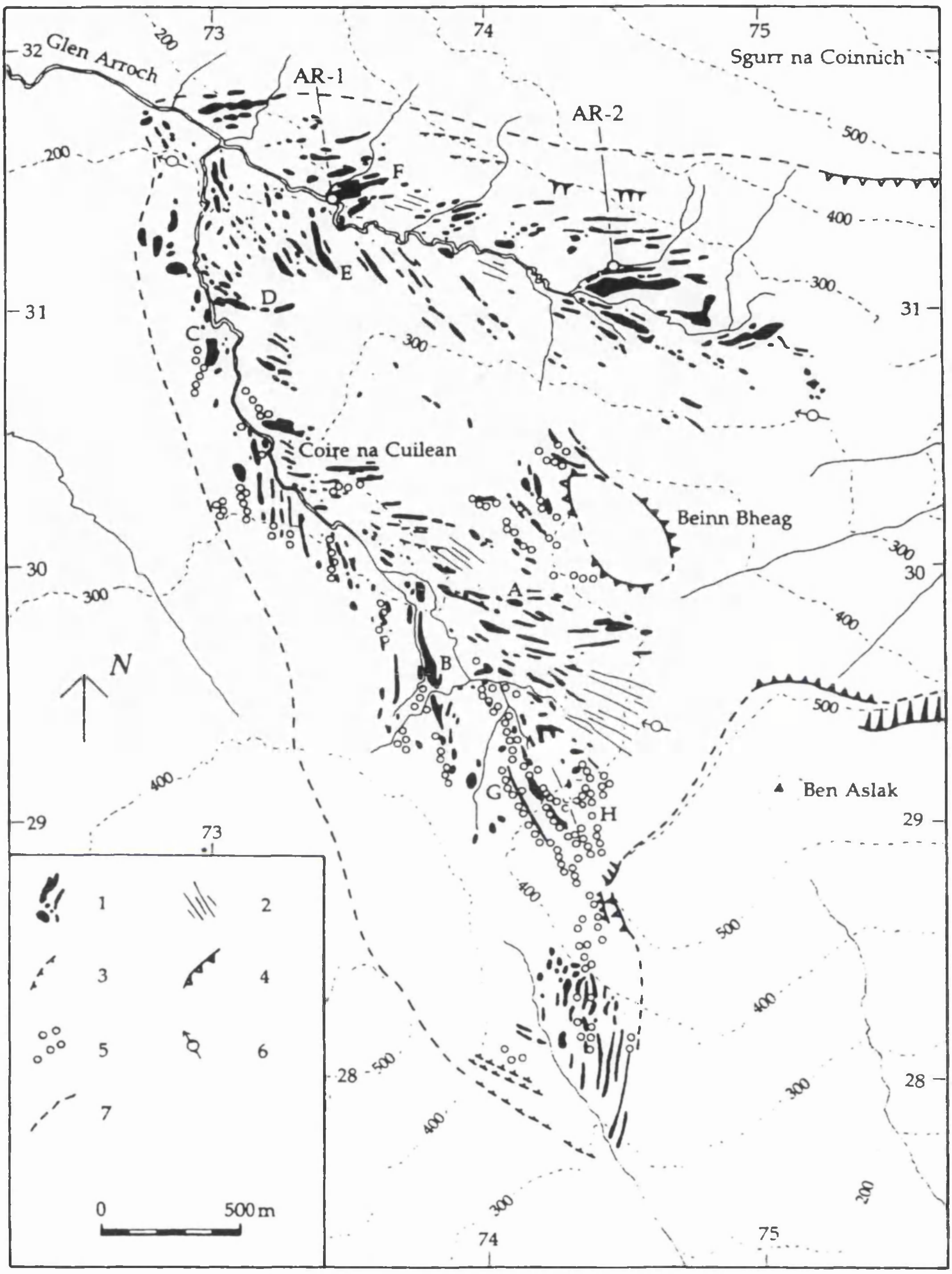


Figure 2.10 Glacial landforms in the Kyleakin Hills, Skye.

1. moraine ridges and mounds
2. flutings
3. drift benches
4. periglcial trimlines
5. glacially-transported boulders
6. striae
7. inferred glacier limit

Source: Benn (1992, p. 785)

Sections in randomly-oriented hummocks and mounds, with rim ridges enclosing hollows, typically reveal debris flows and other subaerial mass-movement deposits, interbedded with water-sorted material (Benn, 1990, 1992). There is usually evidence for several stages of reworking, and sedimentary structures (normal faults, slumps and other collapse structures) indicate deposition in contact with buried ice cores. Benn (1990) has argued that this does not necessarily constitute evidence for areal stagnation, and may instead indicate the decay of small sediment-covered ice cores. Whether this occurred in association with an otherwise actively retreating glacier, or whether areal stagnation of the glacier occurred, cannot be ascertained from the presence of hummocky moraine alone; adjacent landforms need to be taken into account. For example, although the valley floor of Gleann Torra-Michaig in Skye is occupied by chaotic moraines, fragmentary transverse moraines are extensive along the valley sides, suggesting that the chaotic moraines were formed in association with actively retreating ice. Elsewhere, however, chaotic moraines occur in association with stepped kame terraces and appear to have formed during the downwasting of stagnant glacier remnants (Benn, 1990, 1992).

Flutings and drumlins are aligned parallel to the direction of former ice flow, and sedimentologically comprise compact, sheared till containing abundant abraded and faceted clasts. Within tracts of hummocky moraine in central Skye, longitudinally-orientated subglacial bedforms range from flutings ca. 1 m high to small drumlins up to 400 m long and 10 m high (Benn, 1990, 1992).

2.4.2.2 Deglaciation style

Working in Skye, Benn *et al.* (1992) have argued that deglaciation occurred in two phases. The first was one of active retreat, and is represented in the geomorphological record by recessional moraines. Palynological investigations suggest that this first phase was driven by precipitation decline. Thereafter, uninterrupted retreat occurred, and was accompanied by local *in situ* stagnation. This second phase of deglaciation is tentatively associated with thermal amelioration (cf. Coope, 1977a, b). Benn *et al.* (1992) point out that there are many sites in Scotland where fluted moraines can be observed to trend obliquely across the mouths of corries (e.g. Robinson, 1977; Sissons, 1977b; Lawson, 1983, 1986), suggesting that they formed when the glaciers were at, or near, their maximal extents. They have interpreted the apparent absence of alteration as evidence that readvances did not occur during later stages of deglaciation.

In the Northwest Highlands, Bennett (1991) has argued that deglaciation of the Loch Lomond Readvance icefield was characterised by active retreat throughout [although the existence of recessional moraines in the upper valleys of the western Highlands has been challenged by Benn *et al.* (1992)]. It has been suggested by Bennett and Boulton (1993a, b) that the Western Highlands icefield may have behaved differently from the Skye icefield during deglaciation by virtue of its much larger size. They argue that it must have modified its local climate in much the same way as the similarly-sized Vatnajökull ice-cap does today. Thus, it may have decayed in a controlled and more prolonged manner, despite the rapid rise in air temperatures early in the Flandrian. Implicit in this reasoning is that the ability of the icefield to modify its climate will decrease over time, and thus rates of deglaciation should be expected to increase. In this respect, Bennett (1991) presented evidence which suggests that localised stagnation occurred in the vicinity of the decay centres.

2.4.2.3 Delimiting accumulation zones

Most research has focused on delimiting the downvalley extents of Loch Lomond Readvance palaeoglaciers (and, more recently, the genesis of the landforms within), where evidence is most abundant. Relatively little attention has been paid to accumulation zones reconstructions, where ice margins are often extrapolated throughout large sections. Extrapolated ice margins have generally been based on the assumption of an alpine style of glaciation, with glaciers emanating from corries and valley heads (e.g. Sissons, 1980a; Cornish, 1981). This assumption may not, however, be appropriate for areas characterised by broad, rounded summits (the Cairngorms and parts of the Lake District, for example) where plateau icefields may have developed.

A few workers have employed periglacial trimlines to reconstruct the surfaces of these former glaciers in accumulation zones (e.g. Thorp, 1981, 1986; Ballantyne, 1989). Periglacial trimlines mark the upslope transition from ice scoured bedrock to terrain that bears the imprint of contemporaneous frost action (Figure 2.11). They result from the removal of pre-existing regolith and/or periglacial weathering of the upper slopes during glaciation. Most periglacial trimlines associated with Loch Lomond Readvance glaciers have been mapped using the distribution of landforms such as roches moutonées, and periglacial blockfields and tors (e.g. Thorp, 1981; Ballantyne, 1989), together with the depth of rock joints, which tend to be deeper above the trimline (Ballantyne, 1982). Periglacial trimlines are most clearly developed on spurs composed of resistant crystalline rocks (Thorp, 1981). By contrast, frost-resistant lithologies such as lavas, coarse grained granites and massive schists tend to exhibit only limited evidence of Loch Lomond Stadial frost action (Thorp, 1981; Ballantyne and Harris, 1994). In such circumstances, periglacial trimlines may be difficult or impossible to recognise in the field.

Related to periglacial trimlines are so-called 'thermal trimlines' in which the transition from ice-scoured to frost weathered bedrock represents a change in basal thermal regime, from erosive wet-based to protective cold-based conditions (Benn and D.J.A. Evans, 1998). Thermal trimlines do not define former ice surfaces and so should be distinguished from periglacial trimlines. In the context of ice sheet reconstructions, Ballantyne and McCarroll (1995) have argued that periglacial trimlines have smooth regional gradients consistent with theoretical ice sheet profiles. By contrast, the effects of longitudinal stresses on the pressure melting point of ice should ensure that thermal trimlines are elevated on the upglacier sides of high ground and depressed on the downglacier sides (Ballantyne and McCarroll, 1995).

Whilst there is an increasing awareness of the geomorphological significance of cold-based conditions in ice sheet reconstructions, the same cannot be said for restricted glaciations such as the Loch Lomond Readvance. Indeed, the possibility that some periglacial phenomena may have been preserved beneath protective cold-based ice during the Loch Lomond Stadial has not been addressed in the literature.

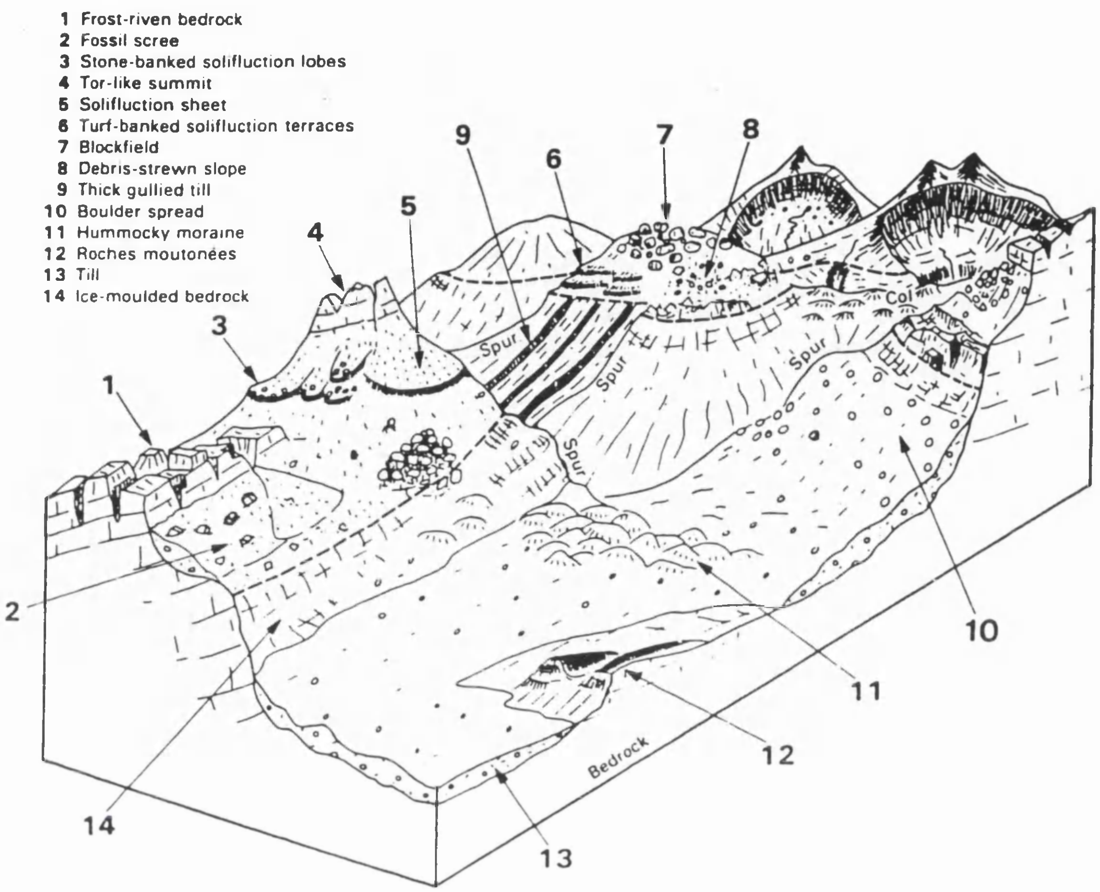


Figure 2.11 Periglacial trimlines and drift limits as an aid to palaeoglacier reconstruction

The broken line represents the upper limit of the most recent glaciation and separates frost weathered bedrock above from ice-scoured bedrock below.

Source: Thorp (1986)

2.4.2.4 Dating Loch Lomond Readvance limits

In general, it has been maintained that the clarity of the glacial geomorphological evidence has facilitated very accurate palaeoglacier reconstructions. For example, Sissons (1977b) was of the opinion that end moraines identify former ice margins to *c.* 10 m, and that boulder limits were probably accurate to within *c.* 1–2 m, giving an overall interpolated ice margin accuracy of 50–100 m. It is worth emphasising, however, that very few Loch Lomond Readvance glacier margins have been radiocarbon dated and/or palynologically constrained (see Gray and Coxon, 1991). In the Lake District, for example, there are no ice margins which have been radiocarbon dated and the spatial resolutions of palynological and lithostratigraphic investigations are too coarse to be useful in ice-marginal reconstructions.

In most studies, heavy reliance has been placed upon moraine morphology (or 'freshness') to distinguish between glacigenic landform assemblages of different ages. Nevertheless, different workers clearly have different conceptions of what exactly constitutes 'fresh.' Thus, Manley (1959) envisaged a slightly more extensive Loch Lomond Readvance in the Lake District than Sissons (1980a) did, even though both workers ostensibly employed the same approach. Although it has been suggested that a quantitative approach to 'freshness' could be developed (J. Boardman, pers. comm.), a more fundamental question is whether 'freshness' is a reliable relative dating technique. This approach assumes that there is a characteristic appearance to Loch Lomond Stadial glacigenic landform assemblages which makes them quite distinct from older glacigenic landforms. In turn, this demands that the variables influencing moraine morphology were more or less constant throughout upland Britain. These variables include glaciation style, lithology, ice-margin behaviour during deglaciation and the impacts of paraglacial reworking. This was clearly not the case. Whilst these issues have not yet been addressed in the literature, it is interesting to note that palynological investigations by Tipping (1989) in western Scotland implies a substantially more extensive Loch Lomond Readvance than is suggested by the geomorphological evidence.

2.4.3 The Loch Lomond Readvance in the Lake District according to Manley

Manley (1959) derived palaeotemperature estimates from the mapped extents of Loch Lomond Stadial glaciers in the Lake District, where the former existence of such glaciers had already been inferred by Pennington (1947) and Walker (1955) on the basis of varves recovered from Lake Windermere and the Kentmere valley respectively. The 400 paired varves from Lake Windermere occur as a 50 cm thick sequence lying conformably above Lateglacial Interstadial sediments and below Flandrian organic muds (Pennington, 1947). Although narrow, the varves displayed clearly graded bedding. This, according to Pennington, indicated that the lake at that time was turbid due to the inflow of glacial meltwater. In turn, this was considered to constitute evidence for a renewed period of glaciation in the high ground of the Windermere catchment during the Loch Lomond Stadial/Younger Dryas.

Walker (1955), working north of Windermere, described varved blue clay associated with Pollen Zone III (the Loch Lomond Stadial) which he extracted from a former lake bed above Kentmere. He interpreted this as representing evidence for an episode of glaciation in the valley, and suggested that it was possible that the varves were associated with the limits of a restricted glaciation tentatively recognised by Dahl (in a personal communication to Walker) a few years previously. During this glaciation, ice was considered by Dahl to have extended down the Kent valley as far as the Hartrigg moraine.

Manley (1959) argued that the glacial episode represented by the varves was clearly recorded in the geomorphological evidence. He recognised what he considered to be a distinct group of glacigenic landform assemblages located in the valley heads and corries of the Lake District. He was of the opinion that these corrie end moraines and valley-head hummocky moraine could be distinguished from other glacigenic landforms in the area on account of their 'freshness'. The 'freshness' of the hummocky moraine was particularly emphasised because Manley believed that these glacigenic landform assemblages belonged to the same event as the one which produced the 'tarn moraines', even though they occurred down-valley from them. He argued (p.190) that the hummocky moraine found in the heads of valleys such as Ennerdale, Easedale, Grisedale, Mardale, and Mickleden, are age-equivalent on the basis that they are all *"...very similar in appearance and contour... ."* That they related to the most recent episode of glaciation was suggested by the impression of 'freshness' which they conveyed, a quality he attributed to their morphology, with slopes frequently 20-25°, occasionally more, and with sharp breaks of slope with the valley-floor, which he took to imply little slumping or other modification prior to their surfaces becoming stabilised by vegetation (Figure 2.12). Hummocky moraine of greater age, Manley reasoned, would be less conspicuous since *"...for 2,000–3,000 years they would carry very scanty vegetation, and in a wet climate, with severe frost in winter, would be likely to weather rapidly... ."*

By mapping the distribution of these corrie end moraines and valley-head hummocky moraine, Manley (1959) constructed a map showing the inferred extent of the Loch Lomond Readvance in the Lake District. Using this information in conjunction with glaciological data derived from present-day southern Norway, he derived palaeoclimatic inferences for the Lake District during the Loch Lomond Stadial. The following sections briefly describe how he did this, and summarise the salient palaeoclimatic inferences. Manley's work is particularly interesting as it is possible to identify the origin of many of the ideas and approaches subsequently employed by workers such as J.B. Sissons during the 1970s and early 1980s.

2.4.3.1 Palaeoglacier reconstruction

Manley (1959) reconstructed Loch Lomond Readvance palaeoglaciers using corrie moraines and valley-head hummocky moraine. He was of the opinion that the corrie moraines did not represent evidence for the maximum extent of these palaeoglaciers (p. 193) *"...unless there is a complete absence of fresh-looking hummocky moraine in the valley below."*

Hummocky moraine was considered by Manley to represent the down-valley limits of Loch Lomond Readvance glaciers, the terminal sections of which, he believed, stagnated *in situ*. The stagnation interpretation appears to have been based on their morphology, which he described as 'irregular' and 'chaotic', and was influenced by the work of Hoppe (1952), Mannerfelt (1945) and Hollingworth (1951) on ice sheet stagnation,

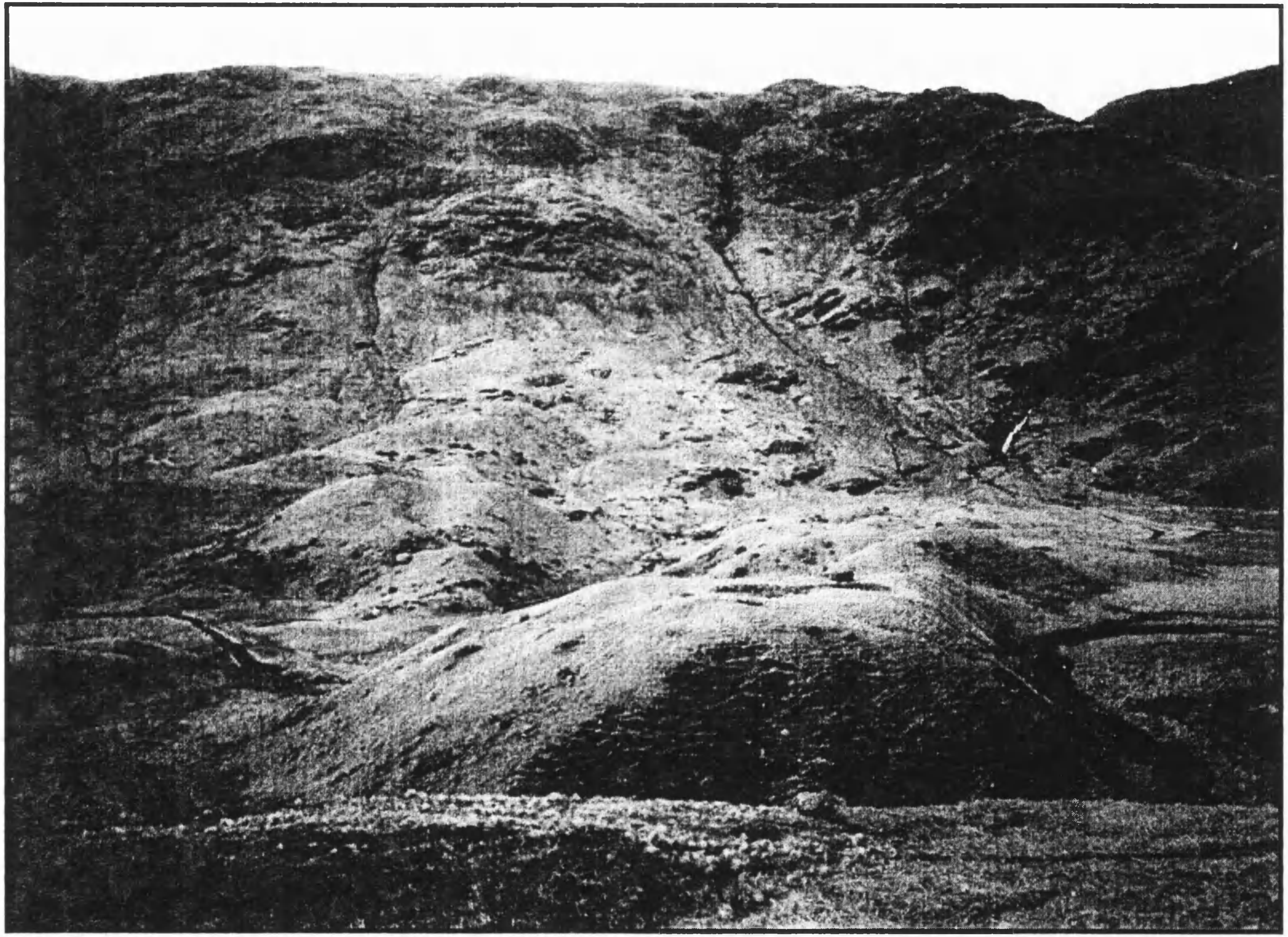


Figure 2.12 Hummocky moraine in Deepdale, east of Fairfield (NY123456)

which he refers to. Manley argued that a position of formation at or near the glacier terminus was suggested by the close association of 'hummocky moraine' with lateral moraines near the valley floor.

Manley (1959, p. 190) stated that the evidence to support his assertion that hummocky moraine represented the limits of a restricted glaciation could also be provided by plotting the average altitudes above sea-level *"...of these prominent dead-ice, terminal and lateral remnants in the wet central valleys..., all of which are very similar in appearance and contour"*(Table 2.1). If, in fact, these glacigenic landform assemblages did represent the former limits of Loch Lomond Readvance palaeoglaciers, he argued, then it would be expected that they would be lowest (i.e. furthest down-valley) in those situations which would have been most favourable for glacier development. He believed such a pattern was indeed evident. Further support for this interpretation was provided by the absence of further 'large' glacigenic landform assemblages, or lateral moraines near the valley floor, up-valley.

Location of Hummocky Moraine	Average Altitude (m)
Borrowdale; above Seathwaite	150
Ennerdale; near the Youth Hostel	275
Langdale; foot of Rossett Gill,	150
Easedale; just below the tarn	215
Grisedale; SE of Helvellyn	215
Mardale	245
Wastdale - Mosedale	140
Eskdale	350

Table 2.1 - Altitudes corresponding to the downvalley termination of hummocky moraine in some Lake District valleys according to Manley (1959, p190).

In addition to hummocky moraine, Manley noted the existence of many 'tarn-moraines' in the corries of the Lake District. However, he did not believe these could be used to reconstruct the maximum limits of Loch Lomond Readvance glaciers. Although he hypothesised that the hummocky moraine in the valley heads represented evidence for the *in-situ* stagnation of the terminal zones of Loch Lomond Stadial palaeoglaciers, he argued that the ameliorating climate would have lowered the surface profile of the former glaciers, ultimately resulting in the 'corrie-glacier' becoming disconnected from the stagnating ice in the valley below. Following this, Manley considered that these corrie glaciers would have remained active even if their budgets were sometimes negative.

Manley observed that corrie end moraines were so located in relation to their backwalls that the ratio between the horizontal distance of the end moraine and the altitude of the crags above it falls close to 1:3. He was of the opinion that a ridge located at the head of Keskadale was too close to the base of the crags to be glacigenic in origin and thus conferred a protalus rampart status. This ridge has subsequently been discussed by Sissons (1980a), Oxford (1985) and Ballantyne and Harris (1994).

Using the types of evidence outlined above, Manley (1959, p. 201) produced a somewhat small-scale map showing what he considered to be the extent of the Loch Lomond Readvance in the Lake District (Figure 2.13). Manley does not show any plateau icefields on his map and describes no geomorphological evidence for their existence. However, he does address the possibility that they may have existed in a consideration of the regional snowline. He believed it likely (p. 207) that a 'snow-dome' may have formed over High Raise in the central fells and stated that *"...a great deal of ice descending from this summit would probably reinforce the Easedale glaciers... ."* However, he doubted whether neighbouring Ullscarf would have played host to an ice-cap on account of its slightly lower altitude, although he believed that it would take only a slight further lowering of the snowline for one to develop. He did not address the implications, if any, that the existence of such plateau icefields would have for palaeoclimatic inferences.

2.4.3.2 Firn line Estimation

Manley estimated that the palaeo-ELA was located approximately half-way between the altitudes of the snout and the base of the upper crags. Such a figure, he claimed, takes into account that a glacier rarely has a uniform surface. The exception was in those instances where *"...allowance should be made for downward narrowing of the valley and diminishing precipitation..."* where he placed the firn line three-fifths of the way up

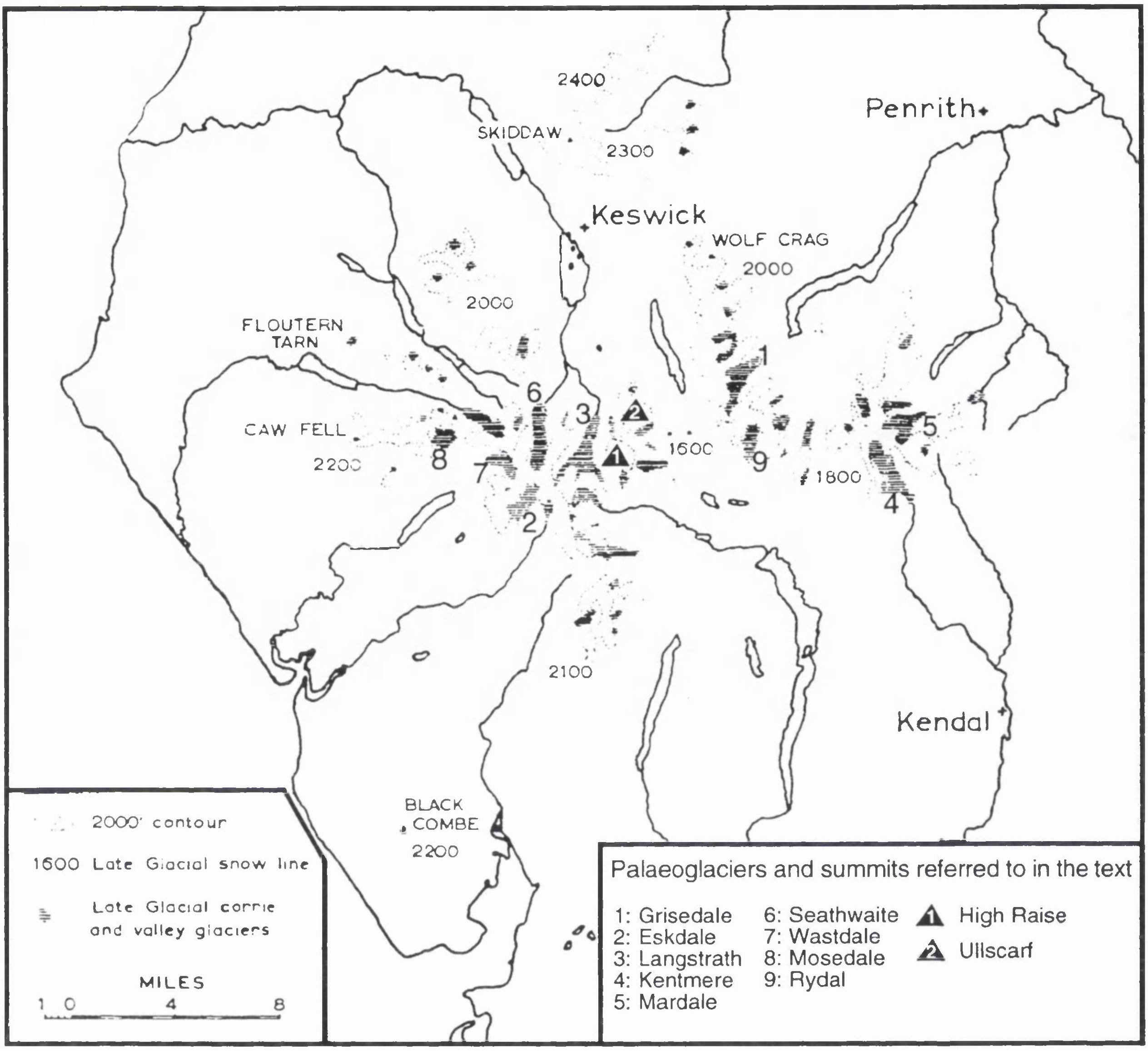


Figure 2.13 The Loch Lomond Readvance in the Lake District according to Manley (1959)
Source: Manley (1959, p. 201)

each palaeoglacier (p. 194). However, Manley considered the whole process of firn line estimation very approximate principally given the uncertainties regarding the limits.

To illustrate his approach to firn line estimation, Manley (1959, p. 194) described the case of the Langdale Combe palaeoglacier, located above Mickleden at the head of Langdale in the central fells. Here, Manley noted the existence of a large number of what he interpreted as 'dead-ice' hummocks, which he inferred to have been produced by a shallow glacier which formed on the slopes north-west of Pike o' Stickle, and which spread into Langdale Combe, just south of the summit of Stake Pass (Figure 4.1). He did not consider that the Langdale Combe ice was ever confluent with the Mickleden glacier in the valley below, although he believed that it did descend the north side of the pass into Langstrath, merging with the palaeoglacier which occupied that valley. Given such a configuration for the Langdale Combe palaeoglacier, Manley estimated that the local snowline did not descend below approximately 490 m.

As a second example, Manley considered the case of the Greenup Gill palaeoglacier, located a short distance to the north (Figure 4.1). A 'conspicuous' end moraine is described as lying at 400 m, with the small crags above at 610–640 m. This led Manley to propose a local firn line altitude of 520 m. Its similarity to the firn line altitude calculated for the Langdale Combe palaeoglacier was considered by him to constitute support for his reconstructions as well as his methods.

Manley considered firn line estimation to be less straightforward in the case of certain corries which would, by virtue of their orientation, be unusually favoured by wind-blown snow. He cites a paper by Enquist (1917) who demonstrated that there is a strong tendency for corries to be located on the lee side of a mountain range with regard to the prevailing snow-bearing wind. Snow accumulates in them preferentially as a result of drifting over the summit, accompanied by back-eddies under the lee of the headwall. Manley reasoned that given the prevailing snow-bearing winds during the Loch Lomond Stadial would have been from the south and south-west, and that most of the corries face between north and east, most of them would have gained appreciably from snow drifting. Particularly favoured would be those corries lying leeward of a long stretch of upland plateau as opposed to beneath a sharp ridge with negligible area. The significance of this, Manley explained, is that total accumulation in a corrie lying leeward of a broad summit would have been greater than elsewhere, resulting in a lower firn line than was generally characteristic of the region. During the Loch Lomond Readvance, therefore, certain corrie glaciers would have survived at lower levels than the larger valley-glaciers.

Manley was interested in estimating the quantitative relationship between the local snowline (i.e. firn line) associated with a sheltered corrie glacier lying in the lee of an extensive plateau, and the prevailing regional snowline. To this end, he made reference to the Skautho glacier in Jotunheimen. This small (< 1 km^2), active glacier, with a then firn line of 1705 m, is described as occupying a sheltered, north-easterly facing corrie to the lee of a 'considerable' stretch of plateau-summit. This is somewhat lower than nearby glaciers occupying more open conditions, all of whose firn lines lie above 1830 m. Manley concluded from this that the effects of additional accumulation in the form of snow blown off the plateau in conjunction with a sheltered location away from the sun served to lower the local firn line by about 150 m.

The firn lines calculated by Manley using the above criteria for the Loch Lomond Readvance palaeoglaciers of the Lake District are shown in Table 2.2. It can be seen that in some locations he distinguished between 'local' and 'regional' firn lines. This follows from his view that certain favourably located palaeoglaciers, principally those occupying north and north-easterly facing corries in the lees of upland plateaux, would gain considerable accumulation from the effects of snow-blow. The addition of 150 m to the local snowline to obtain the regional firn line is only applied, it seems, in the most favoured situations. This includes the Bowscale Tarn palaeoglacier snowline which, at 550 m, had an additional 150 m added to it to obtain a regional firn line estimate of 700 m. Not all firn lines indicated as being 'local', however, were 150 m below the regional firn line (where this was given). Presumably, Manley intended the designation of 'local' in these cases to flag that they had firn lines which departed from the regional value on account of certain favourable or unfavourable circumstances.

Table 2.2 Loch Lomond Stadial firn lines in the Lake District after Manley (1959, Table 1, p. 199)

Group	Valley glaciers	Altitude (m)	Notes
Eskdale	glaciers	600	South-facing
Borrowdale:	Seathwaite glacier	450	Crags 730 m; snout near farm 140 m
	Gillercombe	520	Little ice on adjacent Styhead
	Langstrath	470	Snout probably near 180 m
	Greenup	520	
	Combe Gill	490	Narrow; ice descent to about 320 m
Wastdale:	Piers Gill	490	
	Mosedale	520	Large gathering ground, thin tongue
Ennerdale:	Liza glacier	550	
	Lower Ennerdale	640	"Regional snowline". Based on:-
	Pillar N. side	610	Small ice masses in 'partial corries'
	Seatallan. Local	610	Large snowbed on N. side
	Caw Fell crags (Local)	550	Large drift-source above
	Floutern Tarn (Local)	490	Large plateau drift-source above
Buttermere:	(High Stile)	610	3 corries agree; small drift sources
Langdale:	Mickleden	425	Wide gathering ground
	Stickle Tarn	550	Faces SE
Easedale:		425	Wide basin, much material
	Blind Tarn (Local)	365	Snowline attributable to much snowdrift from plateau
	Wythburn	490	
Langdale to Easedale		490	'Regional snowline' based on the above

continued/...

Table 2.2 – continued

FAIRFIELD GROUP		
Greenhead Gill (above Grasmere)	–	Apparently no ice; faces SW
Rydal	550	Faces South, but high gathering ground
Scandale	?550	Uncertain; gathering ground smaller
Hartsop, Deepdale	520	
Kirkstone (N.)	?490	Nearly due north from Red Screes; ice descended to 180 m
Kirkstone (SE)	?580	Steep short fall off Red Screes; ice descended to 350 m
HELVELLYN RANGE		
Grisedale (NE)	490	
Tongue Gill (S.)	?550	Small local accumulation at head
Glenridding	550	
Glencoyndale	550	
Sticks Pass	610	Thick 'dead ice' hummocky moraine above reservoir
Heads of Matterdale	610	
Wolf Crags (Local)	490	Very large drift source. 'Regional snowline' here >610 m
SKIDDAW-SADDLEBACK		
Scales Tarn	610	Faces E.; small drift-source, but 'dead-ice' traces well below tarn
Bannerdale (Local)	520	Large drift-source; facing NE
Bowscale Tarn (Local)	550	Very large drift-source
[Regional snowline N. and E. of Saddleback	700]	
Southerndale (facing N.)	>670	
Dead Crags (facing NE)	>700	
[Regional snowline N. of Skiddaw	730	
Lonscale Fell (NE side)	–	Probably merely a small snowbed

continued/...

Table 2.2 – continued

CONISTON FELLS		
'Regional snowline'	640	based on:-
Goats Water	>610	Faces SE Tarn dammed by screes. Probably a snowbed.
Low Tarn	580	Some moraine hummocks down to 350 m. Faces NE
Red Dell	>610	Faces SE No sign of recent glaciation, local snowline relatively high
Wetherlam (E.)		
Greenburn	550	
Seathwaite Tarn		
Wrynose	490	Heavier precipitation; active glaciers from Red Tarn and Pike o'Blisco gathering ground, descending in a narrow tongue nearly to Fell Foot
BLACK COMBE		
'Regional snowline'	670	
Corrie snowline	520	Accumulation on NE
HIGH STREET - KENTMERE - MARDALE		
'Regional snowline' about	550	Based on:-
Kentmere	520	Very wide gathering ground facing E.
Mardale	520	
Riggindale	520	
Hayeswater	<550	Pasture Beck probably below 520 m
Swindale (Local)	520	Faces NE with wide area of potential drifting to SW 'Regional snowline' probably 610 m
Long Sleddale	550	Probably shallow accumulation of ice at head.

Manley used his inferred firn lines to address the possibility that plateau ice-caps may have existed over High Raise and Ullscarf in the central Lake District. In an earlier paper, Manley (1955) considered the conditions under which isolated summits of relatively small extent became covered under accumulations of firn or ice. He states (p. 453):

> *"The lowest limit at which a permanently covered summit might exist is evidently but little above the firn line or climatic snowline, provided that its area is sufficiently extensive. Broadly speaking, the narrower the summit, the greater must be its height above the firn line in order to retain a snow cover. In addition to the area, the exact form of the summit no doubt plays some part..."*

For operational purposes, Manley arbitrarily defined plateau-width, measured in the direction of the prevailing wind, as being delimited by the contour approximately 30 m lower than the summit altitude (the value of 30 m appears to derive from most maps at that time having at least 30 m/100 ft. contour intervals.). By applying this criterion to a variety of present-day examples, he concluded that in a disturbed temperate climate, a summit 1000 m broad is likely to retain a snow cap and form a 'dome' if it rises 200 m above the local firn line, as defined by adjacent glaciers. A summit 300 m broad would have to rise 400 m above this level, and a summit with a breadth of only 100 m would need to attain an altitude as much as 600–700 m above the firn line (Section 2.3; Figure 2.6).

Applying this to the Lake District, Manley reasoned that if Ullscarf is 550–820 m wide within 30 m of its summit, and his reconstructed Greenup Gill palaeoglacier suggests a palaeo-snowline of 520–550 m, then by analogy with present day examples this plateau should just fail to have become permanently covered during the Loch Lomond Stadial. High Raise, though of similar width, is slightly higher and apparently wetter so Manley considered it quite possible that it would have become a snow dome. Manley raised the interesting suggestion that only a small lowering of the snowline would have resulted in a quite substantial increase in glacier coverage.

2.4.3.3 Palaeoclimatic inferences

The distribution of Loch Lomond Readvance palaeoglaciers led Manley to suggest that prevailing snow-bearing winds were from the south and south-west. He noted a tendency for preferential glacier development in north-facing corries, particularly those in the lee of upland plateaux. However, he did not discuss the extent to which protection from direct insolation played a part in their orientations. Palaeotemperature estimates were derived by substituting figures relating to snowline altitudes into Ahlmann's curve, which relates accumulation at the ELA with mean summer temperature for southern Norway today.

The palaeo-snowlines reconstructed by Manley from the field evidence increase from about 485 m in the central Lake District to 730 m in the very north (Table 2.2). This increase of 245 m would have witnessed a concomitant decrease in temperature at the firn line, calculated by Manley as being 1.6°C (assuming a lapse rate of 0.6°C/100 m). Manley considered that the best fit for the Lake District data would be provided by the upper part of Ahlmann's curve. The lower part of the curve is associated with a drier climate, and reference to it suggests that a temperature decrease of 1.6°C is too large for the interval of 245 m. In order to derive palaeotemperature estimates from the curve, it is necessary to substitute the water equivalent of accumulation at the firn line. Manley used today's annual precipitation figures for the Lake District. Above the firn line, much of this would fall as snow. In allowing for summer rainfall, Manley estimated that four-fifths of the annual precipitation figures would represent the water equivalent of accumulation. On this basis, Manley discovered that the curve suggested:

> *"...a decrease in the annual precipitation of 55 inches (138 cm) and a rise in the snowline of 245 m. corresponding with a fall of 1.7°C go together. The accordance is such as to lead to the opinion that, until better evidence is forthcoming, in NW England the average annual precipitation during the post-Allerod recession, in years when the glaciers remained in balance, was very nearly as great as to-day."*

In Manley's opinion, this implied a more unsettled, stormier climate since colder air is capable of holding less moisture.

If precipitation totals during the Loch Lomond Stadial were similar to those of today, Manley estimated from Ahlmann's curve that the mean summer temperature at the firn line in the central Lake District (485 m) would have been 3.7°C. Assuming temperatures increased by 0.6°C for every 100 m of descent, then the equivalent sea-level temperature for June–September would have averaged 6.3°C, corresponding to a July mean temperature of 7.5°C.

2.4.4 The Loch Lomond Readvance in the Lake District according to Sissons (1980a)

Sissons (1980a) derived palaeoenvironmental inferences from a study of glacigenic landforms associated with the Loch Lomond Readvance in the Lake District. Although Manley (1959) had already undertaken such a study, it was Sissons' stated objective to provide more detailed evidence for delimiting these former glaciers. Like Manley before him, Sissons cites the work of Pennington (1947, 1978) as evidence for glacier development in the Lake District during the Loch Lomond (Younger Dryas) Stadial. Similarly, he related the glacigenic landform assemblages in the corries and valley heads to the varves in the major lakes on the basis of moraine morphology. Sissons claimed that these glacial landforms were 'fresh' although he did not explain what he meant by this term.

2.4.4.1 Geomorphological evidence for palaeoglacier extents

Sissons' reconstruction of palaeoglacier extents was altogether much more detailed than Manley's (1959) (Figure 2.14). This was facilitated by his use of aerial photographs, at an approximate scale of 1:25,000, onto which he mapped glacial and periglacial landforms prior to checking them in the field. Arguably, the landforms most important to Sissons were hummocky moraine and end moraines, although he also employed boulders and boulder limits, flutings and drift limits.

2.4.4.2 Hummocky moraine

Sissons (1979a) considered hummocky moraine to represent evidence for the *in situ* stagnation of Loch Lomond Readvance palaeoglaciers at or near their maximal extents, an event believed to have been precipitated by very rapid thermal amelioration at the onset of the Flandrian (Coope, 1977a, b). In many cases, the areal distribution of 'fresh' hummocky moraine was used to delimit the extent of these former glaciers. Although Sissons (1980a) did not comment specifically on the genesis and significance of hummocky moraine in the Lake District, he did put forward examples from the area to support his belief that its areal extent could be used for palaeoglacier reconstruction.

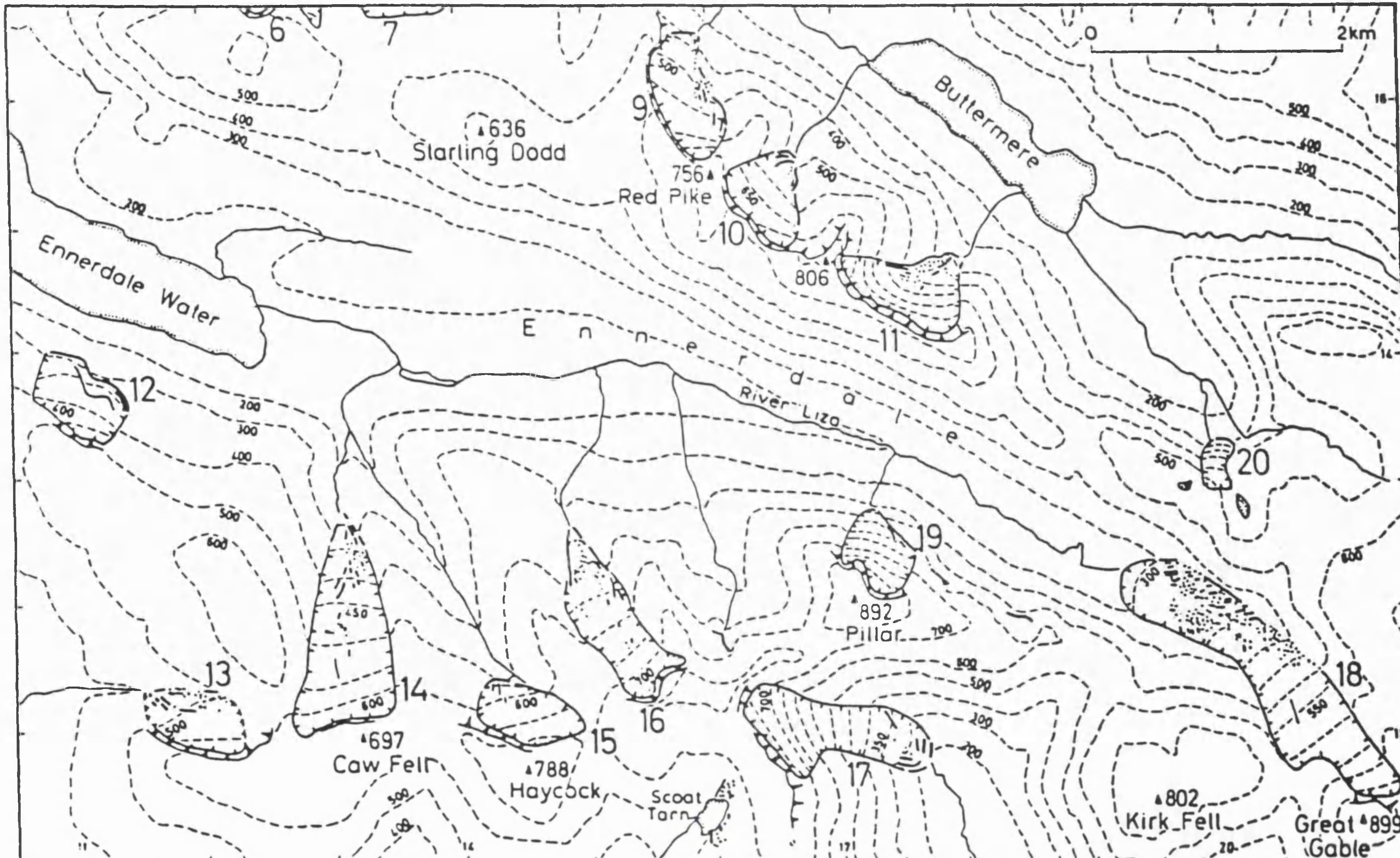


Figure 2.14 Loch Lomond Readvance glaciers in the Ennerdale area, northwest Lake District
Source: Sissons (1980a, p. 16)

Sissons (1980a, p. 18) made the general observation, without quoting any specific examples, that where *"...arcuate moraine and hummocky moraine occur together, the latter always lies upvalley from the former and normally begins immediately behind the end moraine or end moraine belt."* Given this, Sissons reasoned that where end moraines are absent, the down-valley limit of fresh hummocky moraine represented the approximate limit of a Loch Lomond Readvance palaeoglacier. Sissons observed that examples of abrupt down-valley terminations of hummocky moraine can be found in Gillercombe and Langdale Combe in the central fells and Grisedale towards the east. Sissons (1980a, p. 18) stated that in these and some other cases:

> *"...the upper limit of the hummocks rises, often rapidly, up one or both sides of the valley from the inferred glacier limit on the valley floor, thus supporting this inference. Support is also provided by instances where a glacier limit based on hummocks is continued by an end moraine."*

Mickleden and Pavey Ark in the central fells are the examples provided by Sissons to illustrate the occurrence of valley-floor hummocks being continued by an end moraine. In Mickleden, Glacier 30 (location shown in Figure 2.15) has a terminus defined by the extent of hummocky moraine (mapping shown in Figure 4.25). The valley-floor limit is continued for a short distance up the north-eastern side of the valley by a lateral moraine, this also having been noted by Manley (1959). Similarly, the limit of the Pavey Ark palaeoglacier is defined by hummocky moraine which is continued by a boulder limit and a low end moraine. But Sissons also described situations where there is no abrupt limit to hummocky moraine. This includes the case of the palaeoglacier which occupied Honister Pass east of the watershed (Little Gatesgarthdale) (Chapter 5), where the limit was placed in what Sissons terms a 'zone of uncertainty' since the mounds fade out gradually.

2.4.4.3 End moraines

Whereas Manley (1959) considered end moraines to be relatively rare in the Lake District, Sissons (1980a) claimed that clear examples delimit, in part at least, the extent of over forty Loch Lomond Readvance palaeoglaciers. Sissons commented on the association of what he described as 'surprisingly large' end moraines with small Lake

District palaeoglaciers. He considered the most remarkable of these to be the Bowscale Tarn moraine in the north, some 120 m broad and up to 15 m high on both sides, but he also mentioned the Wolf Crags end moraine which stretches for 1 km and is 7 m high on each side. The dimensions of these moraines is interpreted by Sissons as evidence for relatively stable margins. He reasoned that accumulation at both sites would have been considerably supplemented by wind-blown snow from adjacent plateaux, and hypothesised that this would have built up to the top of the headwalls. Once this situation had been attained, any additional snow blown from the plateaux would simply have continued down the smooth glacier surface to be redistributed in the glacier foreland. Thus, the ice-margins would have remained stable.

The major end moraines and arcuate moraines, which occur only in association with the smaller glaciers (maximum volume of 0.05 km^3), are restricted to a terminal zone 100–150 m broad. Where one or more end moraines rise above a terminal drift-belt, Sissons noted that the principal or only ridge corresponds with the very outer margin. He interpreted this as evidence for the glaciers remaining in equilibrium for a time when at maximal extents, and subsequently remaining active while their margins retreated (typically 100-150 m). Sissons considered the almost universal absence of end moraines more than 100-150 m back from these glacier limits to imply *in situ* stagnation followed. He did not discuss his reasons for favouring *in situ* stagnation over uninterrupted retreat.

The absence of clear end moraines demarcating the former termini of the larger glaciers suggested to Sissons that these ice masses remained at their maximal extents only briefly. This differs from Manley's (1959) explanation, which invoked paraglacial reworking by rivers. Some of the larger palaeoglaciers (which occupied Wythburn, Langstrath, Deepdale, Pasture Beck, Hayeswater, and Mardale) are believed by Sissons to have remained active for a time during deglaciation. In these valleys, he identified retreat moraines which he described as trending obliquely down-valley, with a tendency to occur in groups of ridges, the members of which being approximately parallel with each other. His maps clearly show the existence of hummocky moraine between some of these recessional moraines. This may be the reason why Sissons (1980a) did not comment on the genesis of hummocky moraine in this particular paper; the presence of hummocky moraine between recessional moraines is clearly inconsistent with the *in situ* stagnation hypothesis promoted in earlier papers (e.g. Sissons, 1972, 1974, 1977a, b, c).

2.4.4.4 Periglacial evidence

Sissons (1980a) described a 'marked contrast' between the abundant 'fresh' drift mounds deposited by Glacier 34 (the location of which is shown in Figure 2.15, p. 79) and the adjacent ice-free ground, characterised by bedrock knobs and angular boulders. He considered this to represent evidence for minimal modification of Loch Lomond Readvance glacigenic landform assemblages by contemporaneous periglacial activity. In turn, this was used to support the use of moraine morphology ('freshness') as a relative dating technique. Nevertheless, no attempt was made to systematically map the distribution of relict periglacial phenomena in the Lake District. The only periglacial landforms to be considered in any detail were protalus ramparts, of which Sissons identified twenty-six examples. Their proximity to adjacent steep slopes implied that snow could not have accumulated to sufficient depths to have become glacier ice, although he noted the difficulty in distinguishing between very small glaciers and snow beds in the geomorphological record. He did not rule out the possibility that some of his small palaeoglaciers were actually snowbeds, and vice versa. The significance of protalus ramparts to Sissons' environmental reconstruction appears primarily to have been in providing positive evidence for the unsuitability of a site for glacier development. Much more significant was the existence of what Sissons considered to be fossil rock glaciers north of Wastwater. This led him to infer the presence of permafrost and a mean annual temperature at sea-level of 1°C. Unfortunately, no evidence was presented to support his rock glacier interpretations, which have subsequently been challenged (Ballantyne and Harris, 1994, p.242; Whalley, 1997).

2.4.4.5 Palaeoclimatic inferences

Unlike the Western Highlands, where a large ice mass developed (Sissons, 1979a), the Loch Lomond Readvance in the Lake District is thought to have been quite restricted (Sissons, 1980a) (Figure 2.15). The 64 glaciers mapped by Sissons had a total ice volume of approximately 3.3 km^3. Variations in their distribution, size and altitudes led him to derive palaeoclimatic inferences relating to mean summer temperatures (May to September), precipitation patterns, prevailing wind directions, and average atmospheric synoptic conditions. Palaeotemperature inferences were also derived from the periglacial evidence. The palaeoclimatic inferences are summarised as follows:

- **The east–west mountain axis, especially the central fells around Langdale, experienced the heaviest snowfalls at the time of the Loch Lomond Readvance maximum.**

This inference is based on the dimensional and locational attributes of these former glaciers. Of the twelve palaeoglaciers whose volumes exceeded 0.1 km^3, Sissons noted that seven of these, including the four largest, were located in the central fells (Table 2.3). The remaining five were located elsewhere within the east-west mountain axis. Moreover, this axis also apparently contained all those palaeoglaciers whose accumulation areas he considered to be highly adverse to glacier development (Table 2.4). West and south-west of High Raise, for example, glaciers are believed to have developed in Greenup Gill and Langdale Combe, both of which Sissons described as rather shallow open valleys. South of High Raise in Pavey Ark, Sissons noted the former existence of a glacier which was adversely located on account of its southerly aspect. In the east of the district, Sissons commented on a glacier which developed in a shallow valley head in Longsleddale.

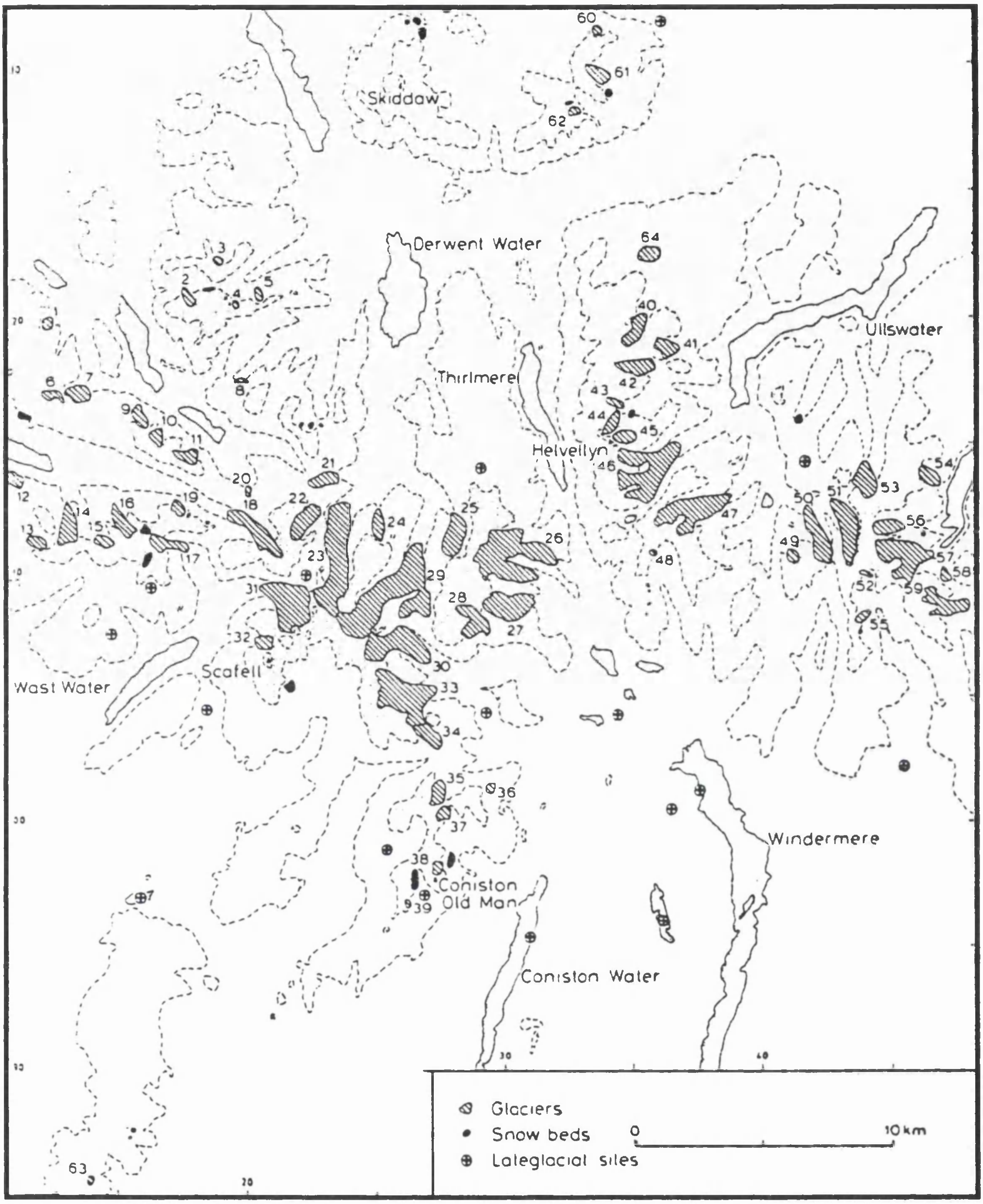


Figure 2.15 The Loch Lomond Readvance in the Lake District according to Sissons (1980)
Source: Sissons (1980, p. 15).

Glacier Number	*Volume*	*firn line altitude*	*Location*	*Aspect*
23	0.347	452	Seathwaite, Borrowdale	north
26	0.240	484	Wythburn	east
27	0.102	458	Easedale	east
29	0.363	483	Langstrath, including Langdale Combe	north
30	0.157	407	Mickleden	south
31	0.187	586	Wasdale	west
33	0.222	511	Oxendale	east
46	0.204	513	Grisedale	north-east
47	0.171	508	Deepdale	north-east
50	0.115	479	Pasture Beck	north
51	0.114	574	Hayeswater	north
57	0.133	542	Mardale	east

Table 2.3 Loch Lomond Readvance palaeoglaciers in the Lake District with volumes in excess of 0.1 km^3 according to Sissons (1980a)

Glacier No.	***Location***	***Comments***
24	Greenup Gill	shallow open valley
28	Pavey Ark	accumulation area faced south
29	Langdale Combe	shallow open valley
59	Longsleddale	accumulated in shallow valley heads

Table 2.4 Adversely located palaeoglaciers, according to Sissons (1980a)

The contribution of wind-blown snow to these palaeoglaciers was believed by Sissons to have been minimal. He thus argued that heavy precipitation in the east-west axis is the only explanation for this pattern.

- **The north-western fells experienced significantly less snowfall than the central fells.**

This inference is based on a comparison of calculated insolation and snow-blow factors, with glacier volumes and ELAs. The highest insolation factor calculated by Sissons for the north-western group (which he defined as comprising glaciers 1-16, 19 and 20) is 10.9, a figure far lower than for any other 'group' of glaciers in the Lake District. Sissons considered the contrast to be particularly marked in the case of the adjacent central group (17, 21-34), in which the lowest insolation factor is 10.3, with several exceeding 14.0. Therefore, despite being more favourably located than the central group with regards to direct insolation, Sissons reasoned the small size of the glaciers in the north west could be explained by lower precipitation. Indeed, Sissons considered some of the north-western palaeoglaciers would not have existed at all if it had not been for the additional contribution of wind-blown snow. This was on the basis of a number of glaciers which Sissons regarded as existing at relatively low altitudes (see Table 2.5) but having high snow-blow factors.

Glacier Nos.	*ELA (m.)*	*Location*
1	284	Black Crags
6	428	Floutern Tarn
7	357	Mosedale (Crummock Water)
8	343	Keskadale
12	384	South of Ennerdale Water
20	408	head of Buttermere

Table 2.5 North-western palaeoglaciers regarded by Sissons(1980a) as existing at low altitudes

- **Snowfall in the south of the district was higher than in the north-west, but not as high as in the central fells.**

Like the north-western group, the glaciers mapped by Sissons in the south were relatively small. Four out of the five palaeoglaciers had firn line altitudes in excess of 600 m and, in this respect he noted similarities with several in the north-western group. But unlike the north-western group, insolation factors were much higher in the southern group. Despite being less favourably located with respect to direct insolation, the glaciers in the south were of similar extents to those which developed in the northwest. Sissons concluded that snowfall in the south must have been higher in order to compensate.

- **Mean annual sea-level temperature in the Lake District was no higher than 1°C.**

This is the only palaeoclimatic inference not directly linked to the glacigenic evidence. Sissons identified what he interpreted as fossil rock glaciers north of Wastwater. Their presence, he argued, indicated the former existence of permafrost which in turn was taken to imply a mean annual temperature no higher than –1°C. The lowest fossil rock glacier identified by Sissons occurs at an altitude of 300 m OD and, assuming a 0.6°C increase in temperature for each 100m descent, he calculated this translated into a mean annual sea-level temperature no higher than 1°C. Despite the significance of fossil rock glaciers in his palaeoclimatic reconstruction, no evidence was presented in support of their existence and there was no discussion on the various modes (including non-permafrost) of rock glacier formation.

- **Mean summer temperature (May to September) at sea-level would have been approximately 6.5°C, equivalent to a July mean temperature of 8°C.**

Using a method described by Sissons and Sutherland (1976), Sissons (1980a) derived palaeotemperature inferences from reconstructed ELAs. This employs a graph provided in a personal communication by Liestøl which relates accumulation at the ELA with mean summer temperature for southern Norway today. Sissons considered the central Lake District palaeoglaciers to be the most suitable for such calculations given their minimal susceptibility to snow-blow effects and other local factors. Today, precipitation in the former glacier accumulation areas is about 3500 mm per year. Sissons was of the opinion that during the Loch Lomond Stadial, the lower moisture-holding capacity of the colder air would have been offset to some extent by more vigorous circulation:

> *"In view of these opposing tendencies a Stadial precipitation of 2000 mm (corresponding to 2°C) seems inadmissible, and even 2700 mm (3°C) seems rather low for it represents a 20-25% decrease compared with the present. It is therefore suggested that May to September average temperature was 3.5 ± 0.5°C at the firn line in the central Lake District, which covers an annual precipitation range of 2700 - 4000 mm. For glaciers in this area not excessively affected by local factors, the firn line is about 500 m, thus suggesting a mean summer temperature at sea-level of 6.5°C and a July mean of 8°C."*
>
> Sissons (1980a, p. 25-26)

- **snowfall was associated with winds from southerly directions**

This is based on the calculation of snow-blow ratios. Manley (1959) made the same inference based on what he perceived to be a tendency for palaeoglaciers to be located beneath north-facing crags. Sissons provided various examples to support the existence of an inverse relationship between snow blowing areas and firn line altitudes. He tested this for each of the sixty-four palaeoglaciers by regressing firn line altitudes with the snow blowing factors, i.e. the square roots of the ratios for each of the quadrants in turn. A statistically significant relationship was found. He interpreted this as demonstrating that winds from a southerly direction had a major influence on glacier altitudes.

- **the average synoptic situation was one in which warm or occluded fronts moved across the area from the west or south-west**

Sissons argued that south to south-easterly winds would normally precede such fronts, when the bulk of the snow would have fallen. These would then have been succeeded by south to south-westerly winds. He believed this to be in accord with the existence of only very small glaciers in the north-west Lake District, and then only in favoured locations since this area would have been in the precipitation shadow of the central mountain mass. Such a situation was also held to account for the locations of glaciers 60–62; they accumulated in the isolated northern mountain massif where the south to south-easterly air streams would have been subjected to strong uplift.

2.5 Summary

The Lake District is assumed to have been characterised by an alpine style of glaciation during the Loch Lomond Stadial (Manley, 1959; Sissons, 1980a). Investigations have placed heavy reliance on the morphological evidence in delineating these former ice margins, with maximal downvalley extents being defined almost entirely on the basis of contrasts in moraine 'freshness.' By way of contrast, no geomorphological evidence has been presented in support of ice-marginal reconstructions in former accumulation zones. Extrapolated ice margins have been based on the assumption of an alpine style of glaciation, with glaciers emanating from cirques and valley heads but with summits remaining ice-free. Although Manley (1959) hypothesised that a plateau icefield may have developed on High Raise in the central Lake District at this time, he presented no geomorphological evidence in support of this and did not show its existence on his small-scale map of palaeoglacier extents.

Recession of some contemporary plateau icefields in north Norway has revealed areas which have experienced little or no subglacial erosion, a situation which has been attributed to low basal shear stresses and, in places, cold based ice (e.g. Gellatly *et al.*, 1988). Where subglacial erosion has occurred, the impacts are generally limited to the evacuation of blockfield and minor scouring of bedrock. Moraine development is usually debris-limited on the summits, and pre-existing sediments may be vital for their formation. However, prominent moraine systems may be produced by plateau icefield outlet glaciers which drain into surrounding valleys, where their margins become traps for debris (e.g. D.J.A. Evans, 1988, 1990).

The implications of research in contemporary glacial environments is that the identification of plateau icefields in the Lake District (and other areas) may be problematical, although faint evidence may nevertheless exist. It is possible that such evidence may have gone unnoticed in previous investigations. The development of Loch Lomond Stadial plateau icefields on some of the more rounded summits in the Lake District is suggested by the existence of what Sissons (1980a) considered to be anomalously located palaeoglaciers. Furthermore, *a priori* reasoning suggests that the failure to account for former plateau icefields will result in an overestimation of ELA lowering. Summits associated with these sites may therefore be worthy of initial investigation. Such a research focus benefits from the recent reassessment of the genesis and significance of hummocky moraine in Scotland by Benn (1990) and Bennett (1991); information on the pattern of decay and decay centres may be provided where these summits are surrounded by hummocky moraine which contains a mappable recessional component.

3

Methods

3. METHODS

3.1 Introduction

This chapter outlines the methods employed in establishing whether there is any geomorphological evidence for the development of Loch Lomond Stadial plateau icefields on the relatively rounded summits investigated. The former existence of hitherto unrecognised plateau icefields may account for anomalously located palaeoglaciers and those with low ELAs as reconstructed by Sissons (1980a) (Section 2.3). It was unrealistic to attempt to map the entire Lake District within the time available and so these locations represent the sampling framework for this investigation. Due regard was also given to summit attributes and the conclusions of Manley (1955) regarding topoclimatic controls on plateau icefield development (Section 3.2). In each area, geomorphological evidence for plateau icefields was searched for on both the summits and in the surrounding valleys, where the emphasis was on reconstructing ice-marginal positions in the late stages of deglaciation (Section 3.3). Finally, these glaciers were reconstructed and firn line altitudes calculated, as described in Section 3.4.

3.2 Site selection

It was not realistic to map the entire Lake District within the time available. One solution could have been to restrict the coverage to just one sector of the Lake District. In his investigation of relict periglacial phenomena, for example, Boardman (1981) considered only the northeast of the district. Such an approach, however, would have necessitated a decision at the outset concerning those areas to be excluded from consideration. Whilst a case can be made that it would have been possible to identify certain areas as being potentially more favourable than others for plateau icefield development [on the basis of summit attributes, assuming Manley (1955) correctly identified the salient topoclimatic controls – see Section 2.3], it does not follow that the failure to identify positive evidence in such areas necessarily means that no evidence will exist elsewhere. Clearly, it would be unwise to conclude that no evidence for Loch Lomond Readvance plateau icefields exists in the Lake District on the basis of a study which considers just one part of it.

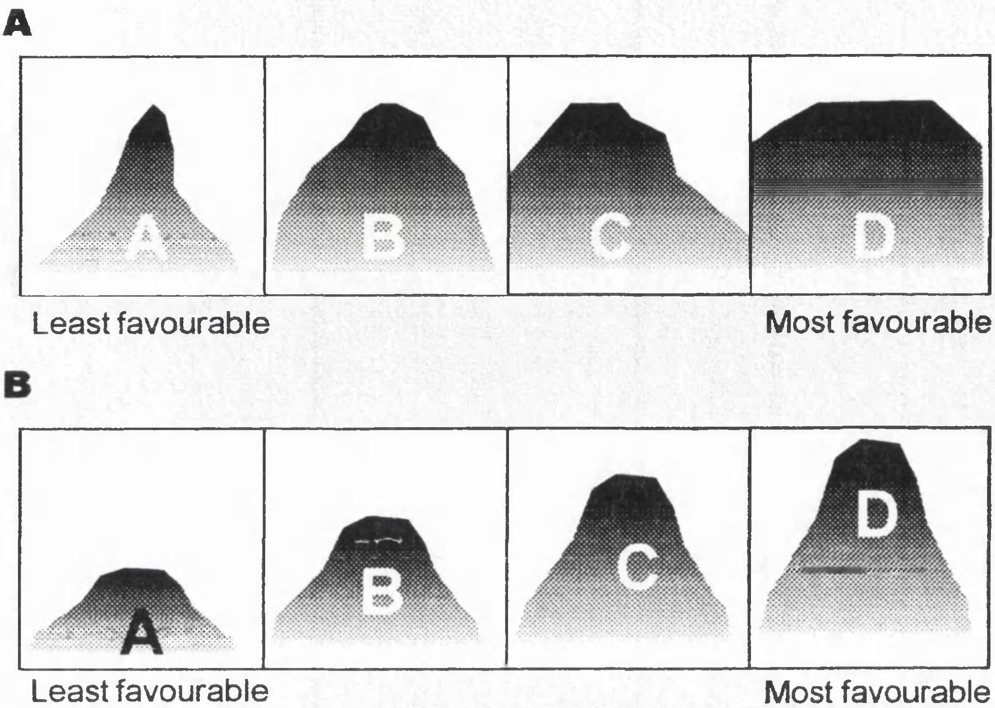
Rather than attempt to delineate the most likely area in which Loch Lomond Readvance plateau icefields may have developed, an alternative approach would be to concentrate on the most favourable summits, regardless of their location within the Lake District. To this end, summit evaluation ideally requires an appreciation of both the salient topoclimatic controls on plateau icefield development and a knowledge of the prevailing (average) boundary conditions. In reality, however, neither of these are sufficiently well understood. However, Manley (1955) considered that the susceptibility of a summit to plateau icefield development would reflect both its altitude and breadth. He stated (p.159):

> *"... the lowest limit at which a permanently covered summit might exist is evidently but little above the firn-line or climatic snowline, provided that its area is sufficiently extensive. Broadly speaking, the narrower the summit, the greater must be its height above the firn-line in order to retain a snow dome."*

By itself, this would not permit the identification of Lake District summits which hosted Loch Lomond Readvance plateau icefields. It is, however, useful insofar as summits can be, to some extent, differentiated on the basis of their relative suitabilities for the development of small plateau icefields. To illustrate this point, Figure 3.1(a) shows a series of summits, each attaining the same altitude but having different breadths. Manley's (1955) observations suggest, all other things being equal, the broadest summit [D in Figure 3.1(a)] would have been the first to have become glacierized. As such, it follows that this site would be chosen for initial investigations. The relationship between altitude and breadth would also be useful in the case whereby, for whatever reason, the first summit to be investigated was summit B. If positive evidence for a small plateau icefield was discovered at Summit B, then it would seem reasonable to suggest that summits C and D would also have supported icefields. Figure 3.1(b) shows a series of summits which have identical breadths but vary in altitude. Using the same reasoning as above, Manley's observations are useful since they suggest summit D would have been the first to have become glacierized, and (probably) the last to become deglaciated, and therefore represents a logical starting point for further investigation. Where a range of altitudes and breadths exists, as is shown in Figure 3.2, the same approach to identifying the most likely site can be adopted; in this case, it is summit D4. And if, for example, positive evidence for a small plateau icefield was recognised on summit B3, then it follows that summits B4, C3, C4, D3, D4 should also have supported plateau icefields, all other things being equal.

In reality, however, the highest peak in the Lake District (Scafell Pike at 978 m) is relatively narrow. This can be clearly seen on Figure 3.3 which shows a plot of altitude against breadth for all summits in the Lake District attaining heights in excess of 600 m OD. These data were extracted from Ordnance Survey 1:25,000 maps. The assessment of summit breadth is inevitably arbitrary, and there appeared to be no compelling reason to deviate from Manley's (*ibid.*) suggestion that plateau breadth should be taken as the horizontal distance between the contours 30 m below the summit in the direction of the prevailing wind (south-westerly). It should be noted that this graph does not take into account continentality, but the Lake District is a relatively compact

A



B

Figure 3.1 Plateau icefield development and summit attributes
A. The four summits shown attain the same altitude but vary in their breadths. Manley's (1955) observations would suggest that, all other things being equal, the broadest summit (D) would be the first to become glacierized in a marginal glaciation.

B. These four summits have the same breadth, but vary in altitude. Again, Manley's (*ibid*) observations suggest that summit D, by virtue of its height, is the most favourable summit for the development of a plateau icefield, all other things being equal.

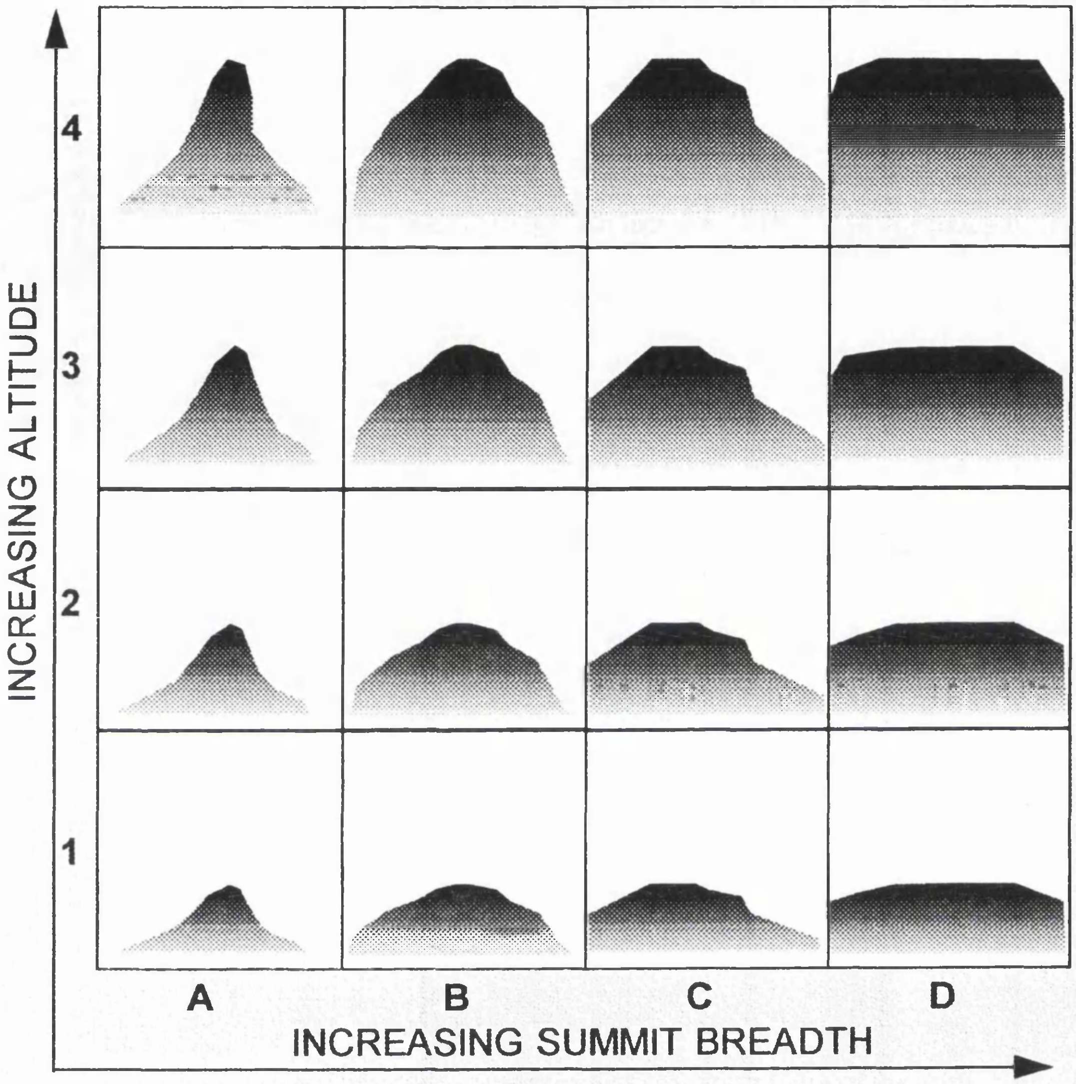


Figure 3.2 **Sampling based on summit attributes.**
See text for explanation.

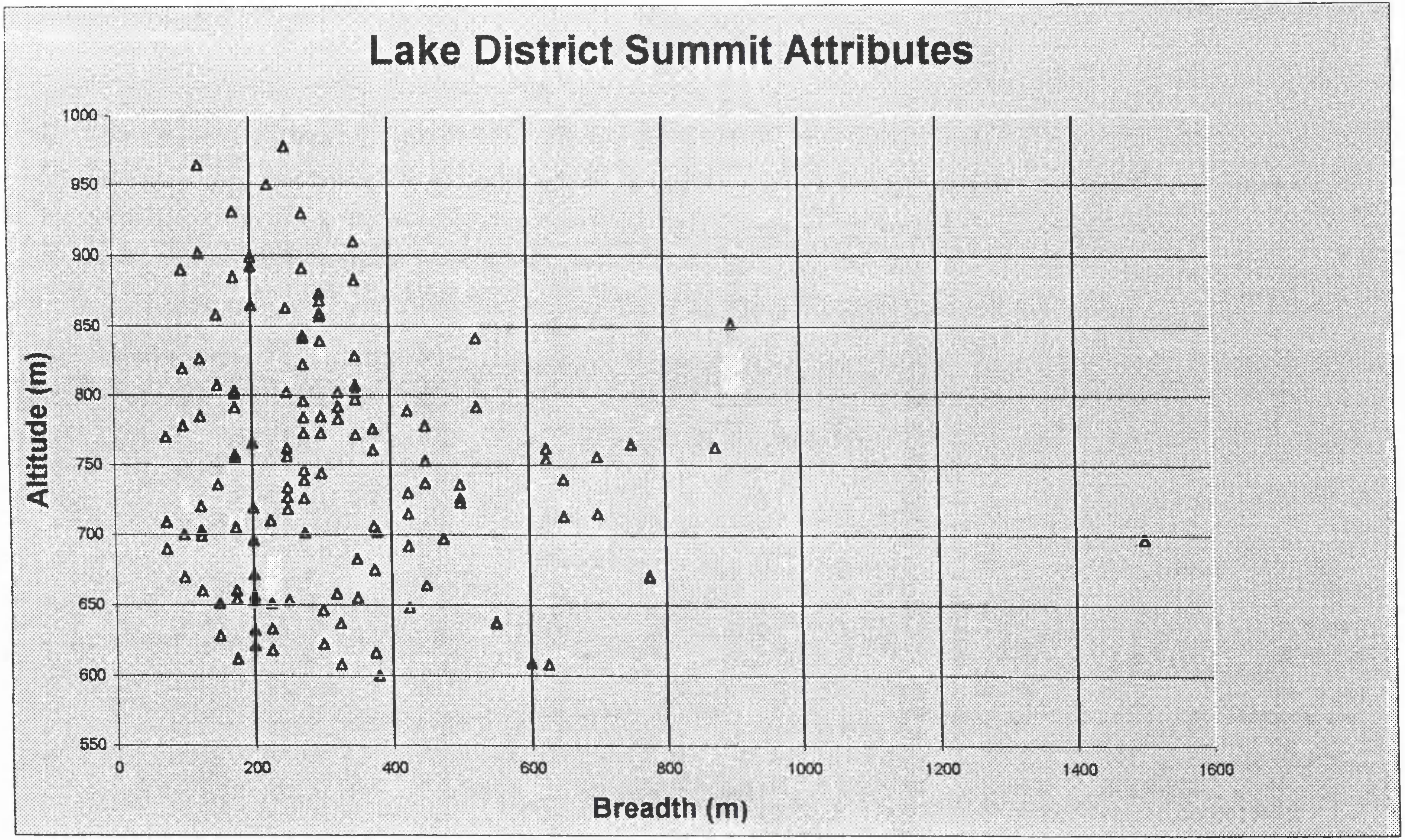


Figure 3.3 A plot of altitude against breadth for all summits in the Lake DIstrict over 600m OD. See text for explanation.

area and so its effects are unlikely to have been marked. In this respect, Sissons (1980a) did not record any statistically significant palaeo-ELA trend within the Lake District.

Whilst it may be fairly obvious in Figure 3.2 as to which summit has the most favourable attributes for plateau icefield development (using Manley's criteria), the same cannot be said for the Lake District summit data shown in Figure 3.3. The question thus arises as to what constitutes the most favourable combination of summit altitude and breadth? This is an important question because the number of summits which can be investigated within the timescale of this research is limited.

If reasonable assumptions are made regarding palaeo-snowline altitudes, then Manley's (1955) curve (Figure 2.6) can be employed in identifying those summits which may have hosted Loch Lomond Stadial plateau icefields. To this end, both Manley (1959) and Sissons (1980a) have derived estimates of firn line altitudes from reconstructed glaciers. Sissons believed a figure of 500 m to be representative of glaciers in the central fells. On this basis, reference to Figure 2.6 suggests that summits 500 m broad would need to attain altitudes of approximately 800 m in order to support a plateau icefield in this part of the Lake District. Similarly, a breadth of 600 m would require a summit altitude of 750 m for plateau icefield development. Nevertheless, given that the firn line altitudes calculated by Sissons (1980a) vary so considerably (the difference between the highest and lowest is 500 m), a more pragmatic approach to sampling has been adopted.

In this investigation, the sampling framework comprised those summits associated with anomalously located palaeoglaciers and those with low ELAs, as reconstructed by Sissons (1980a). The rationale behind this approach is that these 'anomalies' (although Sissons explained some of these by invoking the effects of 'snowblow' from adjacent plateaux) may be due to the failure to account for former plateau icefields. The summits which fall into this category are shown in Table 3.1, from which High Raise, Thunacar Knott, Ullscarf, Grey Knotts and Dale Head were investigated. The proximity of these summits to one another has the advantage that climatic variables are unlikely to have varied markedly between them. It was not possible to investigate all of the summits shown in Table 3.1 within the time available. Indeed, not all of them would have been likely sites for plateau icefield development. For example, although the glacier reconstructed by Sissons (1980) north of Gavel Fell (Glacier 1 in Figure 2.13) is notable for its very low ELA (284 m), the summit is intuitively too low and narrow to have supported a plateau icefield. A preliminary inspection of the area using aerial photographs has revealed the existence of backwall scars and evidence of minor but ongoing mass wasting activity. It is suggested that the mounds and ridges at this site could be the product of a landslide. A detailed investigation of this site is required.

Location	Summit	Altitude	Breadth	Comments
NY281095	High Raise	762	625	Above adversely located glaciers in Greenup (N) and Stake Pass/Langdale Combe (SW), described by Sissons (1980a) as shallow, open valleys. Sampled
NY291122	Ullscarf	726	500	Above adversely located glacier in Greenup (SW), described by Sissons (1980a) as a shallow open valley. Sampled
NY279080	Thunacar Knott	723	500	Above adversely located glacier in Pavey Ark (S), which has a south-facing aspect. Sampled
NY223153	Dale Head	753	450	Above the Honister Pass glacier (S), with significant contribution of wind-blown snow being invoked to explain low ELA Sampled
NY217126	Grey Knotts	697	475	Located to the south of the Honister Pass palaeoglacier (see above). Also east of Glacier 20, considered by Sissons to have an anomalously low ELA (408 m). Significant contribution of wind-blown snow invoked. Sampled
NY202169	Robinson	737	450	Located above the Keskadale palaeoglacier, considered by Sissons (1980a) to have an anomalously low ELA. Significant contribution of wind-blown snow invoked. Not sampled; insufficient time
NY460093	Harter Fell	778	450	Located above Longsleddale palaeoglacier (N), described by Sissons as having developed in shallow valley heads. not sampled; insufficient time
NY117185	Gavel Fell	526	300	Located above Glacier 1, with lowest ELA in Lake District at 284 m. However, summit intuitively too low and narrow to have supported an icefield; ridges and mounds may have a mass wasting origin. not sampled; unlikely location
NY124164	Great Borne	616	375	Located above Floutern palaeoglacier (S), with significant contribution of wind-blown snow invoked by Sissons to account for low ELA (428 m). not sampled; insufficient time
NY142158	Starling Dodd	633	225	Located south of Glacier 7. Significant contribution of wind-blown snow invoked to explain low ELA (357 m).

Table 3.1 Summits associated with Loch Lomond Stadial glaciers which, according to Sissons (1980a), were adversely located or had anomalously low ELAs.

3.3 Field Investigations

There has been an increasing emphasis within geomorphology since the 1960s on relatively small scale surface processes and their associated landforms. As the research focus has changed, so too have the methods; data collection and numerical analysis have tended to replace regional landscape interpretation and geomorphological mapping as the principal approaches to geomorphological inquiry. Nevertheless, there are certain topics for which landscape interpretation and the construction of geomorphological maps remains of fundamental importance. This is certainly true of the present topic, concerned as it is with the landforms and sediments associated with the most recent glaciation to have affected the Lake District. A geomorphological map provides information on the morphology and genesis of landforms within an area and, when the complete succession is established, a chronology of geomorphological development may then be inferred (Mitchell, 1991).

The use of geomorphological maps in Lateglacial environmental reconstructions is perhaps best exemplified by the work of J.B.Sissons and his co-workers who, during the 1970s and early 1980s, delimited the maximal extents of the Loch Lomond Readvance palaeoglaciers on the basis of a range of glacial and periglacial evidence (Section 2.4). Some of the areas mapped at this time have since been re-mapped, and the significance of the geomorphological record reassessed (e.g. Benn, 1990; Bennett, 1991). This serves to illustrate the nature of geomorphological mapping, which is both subjective and dependent upon the skills of the mapper. Subjectivity is inevitable given that decisions need to be made over what to include, and skill (allied to experience) is required both in the accuracy of the plotting and, more fundamentally, in the interpretation of the landscape itself (Cooke and Doornkamp, 1991).

The particular approach adopted will depend on the focus of the research. In this study, the focus is on the summits and the valley heads, where ice-marginal evidence may enable palaeoglacier configurations to be reconstructed in the final stages of deglaciation. Thus, the emphasis is on reconstructing ice-marginal positions during deglaciation rather than moraine genesis. To this end, field investigations were complemented by aerial photograph interpretation.

Aerial photographs have a well-established role within the literature because, quite apart from reducing the amount of time spent mapping in the field, they may reveal the presence of subtle features, such as small meltwater channels, which cannot be discerned on the ground (e.g. Sissons, 1974). However, whether or not such features are visible on aerial photographs depends to a considerable extent on their orientation with respect to the sun. Lineaments perpendicular to the direction of illumination are highlighted whereas those that are parallel may be suppressed to the extent that they may not be visible (Mitchell and Clark, 1994). Therefore, it is clearly desirable to view an area under a range of lighting conditions and, to this end, the author was fortunate in having access to five different sets of aerial photographs (not all of them complete), at approximate scales of 1:10,000, 1:14,000, 1:18,000, 1:20,000, and 1:23,000. These are vertical black and white aerial photographs, apart from the 1:20,000 set which is in colour. The 1:10,000 and 1:23,000 photographs were flown by the Ordnance Survey on a range of dates (see Appendix), with the remainder flown for the Ministry of Agriculture, Fisheries and Food (MAFF) in 1983 (1:14,000) and 1988 (1:18,000 and 1:20,000).

Notwithstanding paraglacial reworking following withdrawal of the ice-margin, landforms on the valley floor tend to be clearest and most distinct, contrasting with summits where glacial geomorphological evidence is, at best, faint. In this respect, the landforms on valley sides can be critical in that they may provide an important link between relatively clear ice marginal evidence on valley floors, and subtle ice-marginal evidence on the summits.

3.3.1 Summits

Investigations focused on the identification and mapping of any geomorphological evidence which could be associated with former plateau icefields. Research at the margins of some contemporary plateau icefields in north Norway have shown that positive evidence may take the form of ice-moulded bedrock (Section 2.2). Thus, the location and orientation of streamlined, abraded bedrock outcrops were noted. At some localities, the vegetation cover was removed in the search for striations although none were found on the summits. This may be due to the rough surfaces of the Borrowdale

Volcanic Group lavas. An attempt was made to identify moraines and meltwater channels, but there is no evidence for their existence on the summits under investigation.

Attention was also paid to the presence (or absence) of relict periglacial landforms (e.g. blockfields, solifluction lobes). On those summits which hosted plateau icefields during the Loch Lomond Stadial, the existence of major relict periglacial phenomena provides valuable information regarding basal thermal regimes in late stages of deglaciation.

A major difficulty encountered in the investigation of summit geomorphology was that in many places the substrate is obscured by blanket peat development. Sometimes in excess of a metre in depth, pits were not a viable option.

3.3.2 Valley evidence

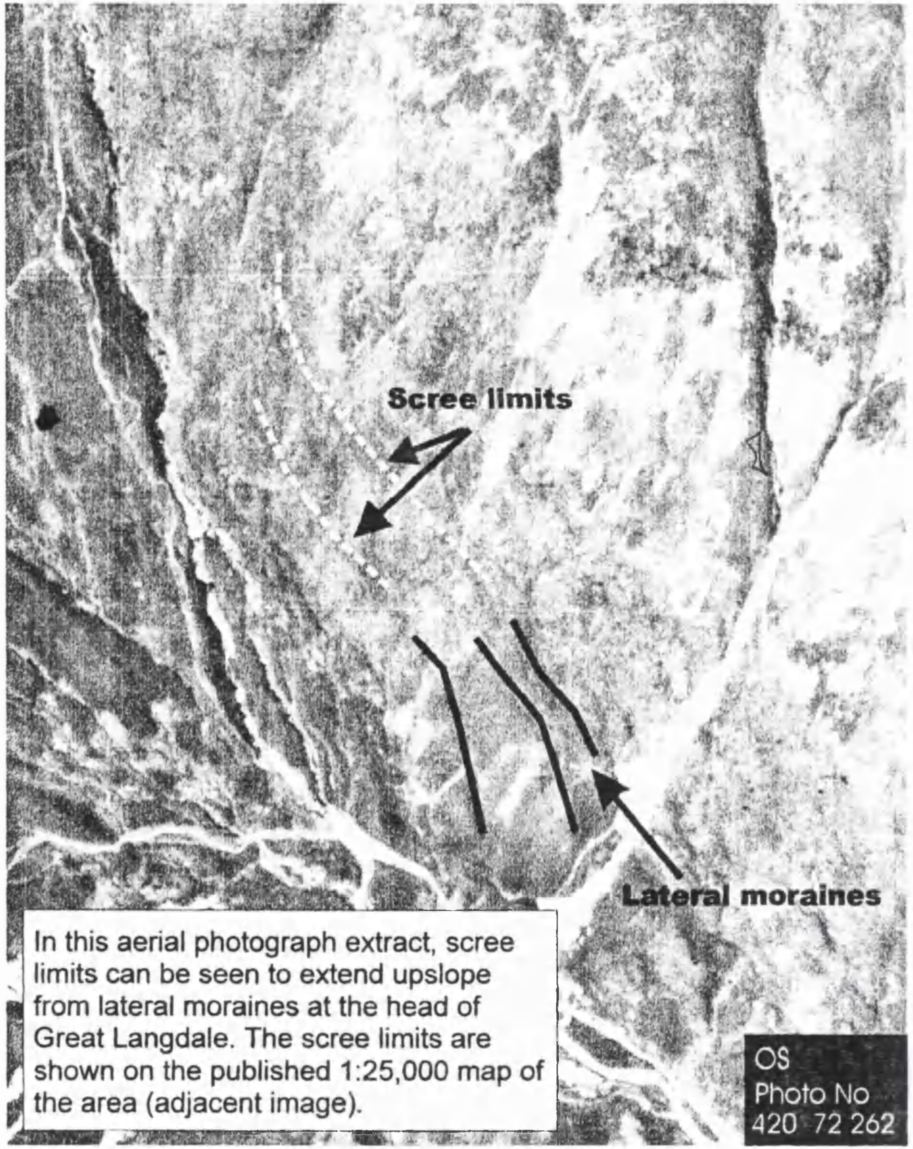
The methods employed in this research, and indeed the focus of the project itself, have evolved over time. That the former existence of plateau icefields in the Lake District could be inferred from glacigenic landform assemblages on the valley floors was first appreciated when mapping the moraines in Stake Pass and Langdale Combe, southwest of the summit of High Raise in the central fells. Previously interpreted as evidence for glacier stagnation (Manley, 1959; Sissons, 1980a), careful mapping using aerial photographs revealed that these hummocky landforms record successive ice-marginal positions of an outlet glacier backwasting towards the high ground of High Raise and Thunacar Knott (Chapter 4). Thus, it was speculated that a similar situation might prevail at other localities, and that former plateau icefields may be recognised by mapping ice-marginal positions in the late stages of deglaciation. The approach employed in this research emphasises the morphological evidence and is therefore similar to that used by Bennett (1991) in the northwest Highlands of Scotland. The lack of suitable exposures in the landforms mapped precluded a detailed sedimentological analysis.

Ice-marginal configurations during deglaciation were reconstructed by mapping ice-marginal landform assemblages (predominantly moraines) on aerial photographs, the results of which were then checked in the field. Because ice marginal moraines can appear hummocky and chaotic in the field, they are most easily and accurately mapped using aerial photographs, assuming suitable images exist (cf. Benn, 1990). They are easily recognised on account of their distinctive planimetric properties, which mirror those of present day ice marginal moraines (e.g. Bennett, 1991). They tend to occur as parallel suites of down-ice orientated chevrons, and are composed of chains of aligned hummocks and ridges (Benn, 1990, 1992; Bennett, 1991, 1994; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b). The occurrence of bifurcating ridges was carefully noted, as their presence lends support to the interpretation that these features were formed in association with an actively retreating ice margin; studies in present day glacial environments reveal that the only cross-valley landforms which bifurcate are ice marginal moraines (e.g. Price, 1973). Bifurcations result from the differential retreat along a section of ice front during a period of moraine formation.

The absence of detailed sedimentological evidence within the study areas (no suitable sections were identified, and the inferences which can be drawn from clast morphology data extracted from shallow pits are limited) meant it was not possible to subdivide these ice marginal moraines according to genesis (e.g. push, dump, ablation). This contrasts with the detailed sedimentological investigations by Benn (1990) on Skye, but identical to the approach of Bennett (1991).

In order to determine whether or not the Loch Lomond Readvance corrie and valley glaciers in the Lake District [as reconstructed by Sissons (1980a)] were actually outlet glaciers draining small plateau icefields, particular attention was paid to delimiting ice-marginal positions in the late stages of decay in the vicinity of the summits. To this end, lateral moraines, scree limits and the limits of drift against drift-free terrain were mapped. Such valley-side evidence is extremely important because it can provide a critical link between relatively clear valley-floor evidence below, and the rather faint evidence on the upper slopes and summit areas.

Lateral moraines on lower slopes may be very distinct and are easily mapped both on the ground and on aerial photographs. This contrasts with the situation on the upper slopes, where they may be difficult to discern on the ground. Nevertheless, aerial photograph interpretation often reveals that these lateral moraines are continued by scree limits



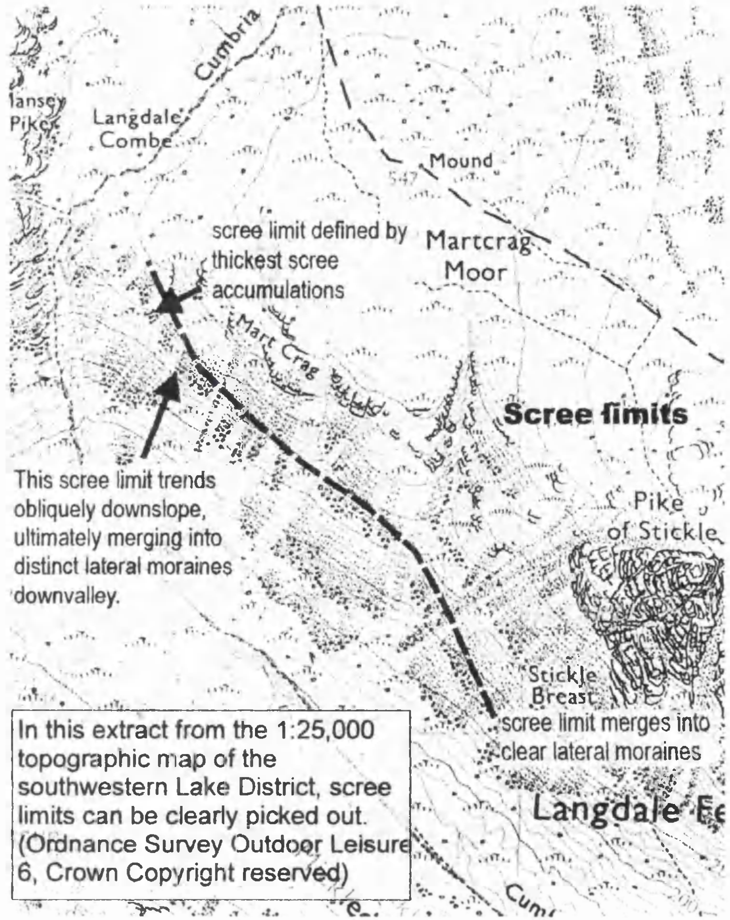


Figure 3.4 Relationship between lateral moraines and scree limits

(Figure 3.4). Scree limits are relatively narrow and discontinuous bands of predominantly sub-angular and angular debris which trend obliquely down valley sides towards former glacier snout positions. These were first described in the New Zealand Alps by Chinn (1979), and are interpreted as lateral moraines which have spilled down steep slopes following removal of the ice margin. Such an interpretation is supported in the Lake District by the fact that in many cases they grade imperceptibly into 'normal' lateral moraines as they approach the valley floor. They are not *in situ* scree accumulations because they lack talus forms and, more to the point, they lack an obvious source (Chinn, 1979). Scree limits, so defined, are often vegetated and can only be identified with confidence on aerial photographs. In the field they are very indistinct. On the geomorphological maps in Chapters 4 and 5, they are shown as lateral moraines for convenience.

The delimitation of ice-marginal positions on valley sides is complicated by the impact of paraglacial reworking by mass wasting processes on valley sides. It is not clear to what extent paraglacial processes on valley sides (and floors) have altered or destroyed landforms associated with the Loch Lomond Stadial in the Lake District. Investigations in contemporary glacial environments demonstrate the efficacy of mass wasting processes on valley sides (e.g. Ballantyne and Benn, 1994a), but the impact and implications for reconstructing Loch Lomond Stadial glaciers in Britain have yet to be investigated in any detail. Such an assessment for the sites investigated is beyond the scope of the present study, although there are a few localities described where paraglacial resedimentation has had an impact and must be considered in ice-marginal reconstruction.

3.3.3 Sedimentological evidence

The dearth of sections within the study area makes it unsuited to detailed sedimentological investigations. Nevertheless, an attempt was made to collect sedimentological data from pits excavated into landforms. Inevitably, this restricted the type of data which could be collected and thus the limited sedimentological analysis within this research is based on clast morphology (shape and roundness). The data were collected primarily to establish whether there is any evidence for active transport (Chapter 4).

The investigation of shape and roundness characteristics of clasts contained within glacigenic sediments can potentially provide valuable information relating to debris transport history, and this can assist in landform interpretation. Following entrainment, two distinct modes of glacial debris transport have been recognised by Boulton (1978). First, debris in passive supraglacial or englacial transport undergoes little modification and retains the characteristics of the parent material, which is typically frost-shattered or rockfall debris. Secondly, active transport occurs at the glacier sole or within deforming subglacial sediment (Boulton, 1982). During active transport, frequent interactions between particles and with rigid obstacles results in progressive debris modification by crushing and abrasion (Boulton, 1978; Ballantyne, 1982). Debris may move from active to passive transport (and vice versa) by a variety of processes . Although the shape and roundness characteristics of actively and passively transported debris is partially dependent on lithology, for many rock types actively transported debris is characterised by abundant edge-rounded clasts, high c/a axial ratios, and a high incidence of facetting and striations, whereas passively transported debris tends to be angular with low c/a axial ratios (Boulton, 1978; Ballantyne, 1982; Matthews and Petch, 1982; Benn and Ballantyne, 1993, 1994). A conceptual model of debris transport pathways and processes in the glacial system is shown in Figure 3.5.

In the field, data were collected from pits excavated along ridge crests to a depth of at least 0.3 m. From each pit, fifty clasts of the same lithology were measured. Sampled clasts were restricted to the size range 35-125 mm (a-axis) because within this range clast form and roundness are believed to be independent of size (Shakesby, 1989). Each clast was assigned to a roundness category on the basis of criteria outlined in Table 3.2. Clast shape was assessed by measuring the long, intermediate and short (a, b and c) axes to the nearest 5 mm. Nevertheless, the utility of clast morphological data collected from relatively shallow pits on the surface of moraines is very limited in this particular exercise. Apart from the inherent difficulties in demonstrating that debris has not been

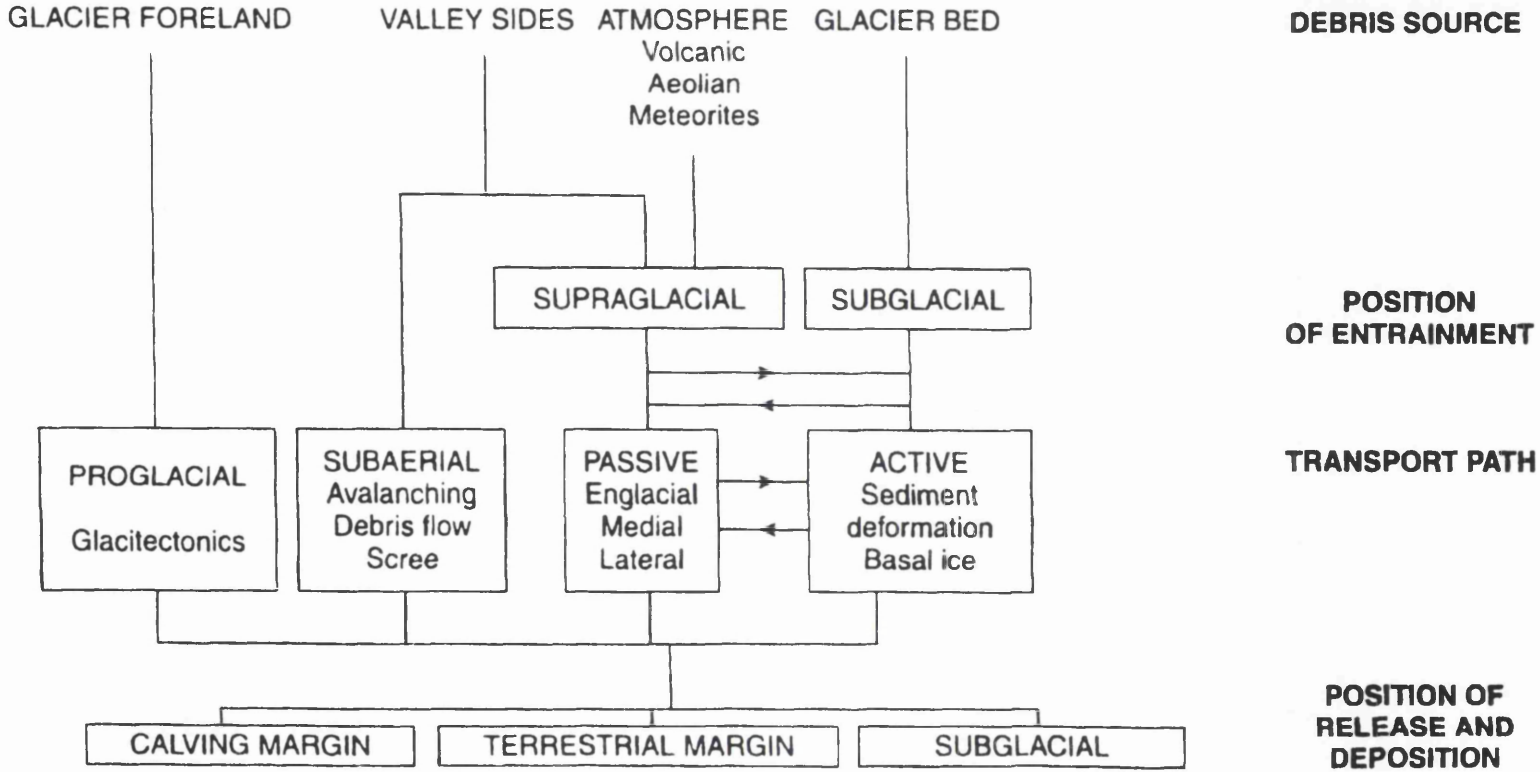


Figure 3.5 Debris cascade system
Source: Benn and Evans (1998, p. 379)

reworked from earlier glaciations, the level of detail provided does not substantially enhance the morphological interpretations.

Roundness class	Description
Very angular	Very acute edges and/or sharp protuberances
Angular	Acute edges with no evidence of rounding
Sub-angular	Rounding confined to edges; faces intact
Sub-rounded	Rounding of edges and faces; often facetted
Rounded	Marked rounding of both edges and faces; merging of edges and faces

Table 3.2 Classification of clast roundness (after Benn and Ballantyne, 1994).

The extent to which actively transported clasts are blocky and passively transported clasts are slabby or rod-like will depend partly on lithology, and so control samples were taken. These were designed to be representative of actively and passively transported clasts of particular lithologies. Passively transported debris control samples were taken from scree samples, which have only experienced subaerial weathering. Unfortunately, the dearth of sections meant that it was not always possible to collect control samples of actively transported debris.

The analysis of the data was relatively straightforward. For each sample, clast shape and roundness is shown graphically, following the approach of Benn and Ballantyne (1994). For each clast, the a, b and c axes define three basic shapes: (1) blocks, with high b/a and c/a axial ratios; (2) slabs with b/a high and c/a low; and (3) rods with b/a low and c/a low. The b/a and c/a axial ratio of each clast was plotted onto a ternary diagram. Actively transported clasts tend to be blockier, and plot towards the top of the diagram, whereas passively transported clasts tend to be more slabby and/or elongate, and therefore plot lower down. A number of authors have attempted to quantify such contrasts by calculating the percentage of clasts in a sample with c/a axial ratios less than or equal to some reference value, usually 0.4 (e.g. Ballantyne, 1982, 1986; Douglas and Harrison, 1987; Benn and Ballantyne, 1993). Benn and Ballantyne (1994) have termed this value the C40 index, and this is employed within this research. The lower the C40 value, the 'blockier' is the sample.

3.3.4 Dating

Direct radiometric dating of glacigenic landforms is rarely possible and so most workers have employed bio–, litho– and morphostratigraphical approaches. In this research, published palynological dates (e.g. Walker, 1966; Pennington, 1978) have been complemented by coring to determine lithostratigraphies. These were extracted using a Russian corer. Complete Lateglacial lithostratigraphies are interpreted as representing evidence for ice-free conditions throughout this period (Gray and Coxon, 1991) (Figure 3.6). However, suitable coring sites rarely coincided with inferred ice marginal positions, and thus the sampling resolution was relatively poor.

Many workers have employed moraine 'freshness' as a relative dating technique, although its use implies the acceptance that Loch Lomond Stadial landforms should everywhere be morphologically similar and distinct from those produced during the decay of the last ice sheet. As this has never been demonstrated (or fully considered) in the literature, and since intuitively such a process is likely to be unreliable (given variations in debris supply, topography, ice-margin behaviour during deglaciation etc.), 'freshness' was not used in this study. Instead, landforms of different ages were differentiated on the basis of style of deglaciation. It is generally accepted that the last ice sheet stagnated within the Lake District (e.g. Boardman, 1981). If the margins of Loch Lomond Stadial glaciers actively backwasted, at least for a short distance, then this provides a means of differentiating between Dimlington Stadial and Loch Lomond Stadial glacigenic landforms which is independent of moraine morphology. Thus, the arcuate end moraines at Rosthwaite (Borrowdale) and those at Wythburn (Clark and Wilson, 1994) are probably not associated with the wastage of the last ice sheet. There are a number of limitations with this approach, the principal one being that the style of deglaciation of the last ice sheet has not been rigorously investigated. Furthermore, it also assumes that there was no advance of glaciers between the wastage of the last ice sheet and the renewed glaciation during the Loch Lomond Stadial. Finally, this approach will not work where localised stagnation of Loch Lomond Stadial glaciers occurred.

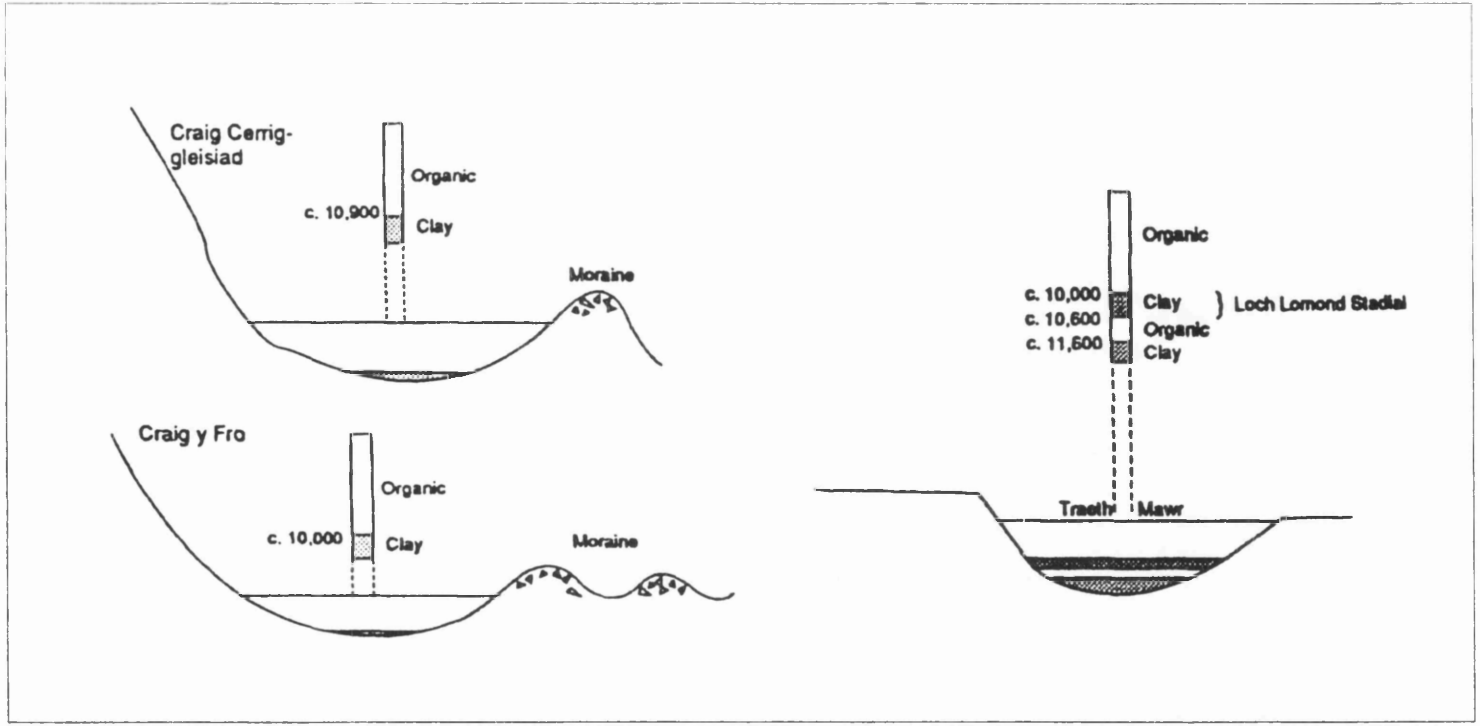


Figure 3.6 Lithostratigraphies and approximate radiocarbon dates from peat bogs in the Brecon Beacons. The Traeth Mawr site shows a full Lateglacial lithostratigraphy; Dimlington Stadial inorganics overlain by Lateglacial Interstadial organics and Loch Lomond Stadial inorganics, with Flandrian organics at the top of the sequence. The absence of Lateglacial Interstadial organics and Dimlington Stadial inorganics at Craig y Fro and Craig Cerrig-gleisiad has been attributed to erosion by Loch Lomond Stadial glaciers
Source: Shakesby (1992, p. 21)

3.4 Reconstructing former glaciers

In many parts of the world, including northern and western Britain, the extents of former glaciers have been used to derive palaeoclimatic inferences (e.g. Porter, 1975; Cornish, 1981; Meierding, 1982; Leonard, 1984). This commonly entails the delimitation of these former glaciers at maximal extents using geomorphological evidence, and contouring of the glacier surface to allow the overall dimensions of the former ice masses to be determined. This in turn permits the calculation of their palaeo equilibrium line altitudes from which climatic inferences are drawn. Studies of the mass balance of modern glaciers have indicated that the ELA is the most critical parameter in the link between glaciers and climate (e.g. Sutherland, 1984).

For valley/cirque glaciers in 'steady-state' conditions, it is commonly observed that the equilibrium line divides the surface of the glacier such that the accumulation area proportionately constitutes 0.6 ± 0.05 of the overall surface area (Meier and Post, 1962; Porter, 1975). This relation has been used to calculate the ELAs of former glaciers reconstructed from the geomorphological record, and is termed the Accumulation Area Ratio (AAR) method. This method has been used to estimate modern ELAs in the Canadian Arctic islands (Andrews and Miller, 1972; Miller *et al.*,1975; Hawkins, 1985), and palaeo-ELAs in West Pakistan (Porter, 1970), the Cascades (Scott, 1977), and the Southern Alps of New Zealand (Porter, 1975). This is widely considered to be one of the most accurate methods for estimating palaeo-ELAs (e.g. Torsnes *et al.*,1993). However, it should be stressed that the ratio of 0.6 ± 0.05 has been derived empirically, and that AARs for glaciers in equilibrium with climate may range from 0.4 to 0.8, the exact figure depending to a large extent on glacier morphology (particularly hypsometry, which is the distribution of glacier area over its altitudinal range) and debris cover. It is certainly the case that this ratio does not hold for glaciers which deviate from a uniform area/altitude distribution, such as icefields (Meier and Post, 1962; Pierce, 1979; Furbish and Andrews, 1984).

In Britain, palaeo-ELAs of Loch Lomond Stadial glaciers have been calculated from the area-weighted mean altitude (AWMA) method (e.g. Sissons, 1974; Sissons and Sutherland, 1976; Sutherland, 1984; Ballantyne, 1989; Mitchell, 1991). This assumes that the ablation gradient (the rate of decrease of the ablation rate with altitude) and the accumulation gradient are each linearly related to altitude. On this basis, the altitude of the firn line is dependent only on the altitudinal distribution of the glacier surface. Thus, this approach can be employed on plateau icefields (e.g. Sissons, 1974; Ballantyne, 1989) as well as cirque and valley glaciers (e.g. Sissons, 1980a; Mitchell, 1991). The altitude of the firn line can be calculated from the following equation:

$$x = \frac{\sum_{i=0}^{n} A_i h_i}{\sum_{i=0}^{n} A_i}$$

where x = the altitude of the firn line in metres

A_i = the area of the glacier surface at contour interval i expressed in km^2

h_i = the altitude of the mid point of contour interval i, and

$n+1$ = the number of contour intervals

(Sissons, 1974, p. 109)

This approach is employed in this investigation because it takes account of the area/altitude distribution of the glacier. As with other methods, there is an assumption that glaciers have achieved mass balance equilibrium (Sutherland, 1984; Leonard, 1984; Murray and Locke, 1989). It is also important to appreciate the critical nature of the contouring of the former ice surfaces, since very small increases in contour curvature can lead to significant changes in ice thickness, particularly on small ice masses (Mitchell, 1991).

3.5 Summary

This chapter has outlined the methods employed in establishing whether any of the Loch Lomond Stadial valley and cirque glaciers in the Lake District mapped by Sissons (1980a) were, in fact, outlet glaciers draining small plateau icefields. These methods have, together with the remit of the topic, evolved over the duration of the study. Prior to the author establishing that the glacigenic landforms in Stake Pass and Langdale Combe southwest of High Raise actually recorded successive ice-marginal positions of an outlet glacier, thus enabling a plateau icefield on High Raise and Thunacar Knott to be inferred, there was nothing in the literature which suggested that the topic was viable. Although due consideration has been given to the geomorphology of the summits, the emphasis within this research has been on the morphological evidence in valleys surrounding the summits under investigation. The sampled summits are those associated with anomalously located palaeoglaciers (according to Sissons, 1980a) and those with very low ELAs.

4

High Raise, Ullscarf & Thunacar Knott

4.1 INTRODUCTION

High Raise is a relatively broad, rounded summit which attains a maximum altitude of 762 m (Figure 4.1). It is located in the central fells, north of Great Langdale and to the east of Langstrath. To the north, east and southeast of the summit, a radial pattern of valley incision has occurred, but this sense of order is countered to the west by the north-trending Langstrath valley. High Raise is bounded by Ullscarf (NY291122) to the north and Thunacar Knott (NY279080) to the south, summits which attain altitudes of 726 m OD and 723 m OD respectively (Table 4.1).

Evidence is presented in this chapter for the development of a plateau icefield on High Raise, Ullscarf and Thunacar Knott during the Loch Lomond Stadial. The geomorphological impact of this icefield appears to have been minimal on the summits, where the survival of blockfields and other frost-weathered debris (mostly peat-covered) implies predominantly cold-based conditions. A transition to warm-based conditions at summit edges is evidenced by the presence of streamlined bedrock above valley heads. Prominent moraine systems were produced by outlet glaciers which descended into the surrounding valleys, where their margins became sediment traps for supraglacial debris and inwash. In some valleys, ice-marginal moraines record successive positions of outlet glaciers which actively backwasted towards their source.

The remainder of this chapter consists of four sections. The evidence for the plateau icefield in the area is outlined in Section 4.2, where the emphasis is on reconstructing palaeo ice-margins during deglaciation at several key sites. This is followed by a reconstruction of this former plateau icefield system at maximal extent in Section 4.3. Finally, the interpretation of the geomorphological record proposed here is compared with those of previous workers (Section 4.4).

SUMMIT	GRID REFERENCE	ALTITUDE	BREADTH
High Raise	NY281095	762	625
Ullscarf	NY291122	726	500
Thunacar Knott	NY279086	723	500

Table 4.1 Summit attributes of High Raise and adjacent peaks

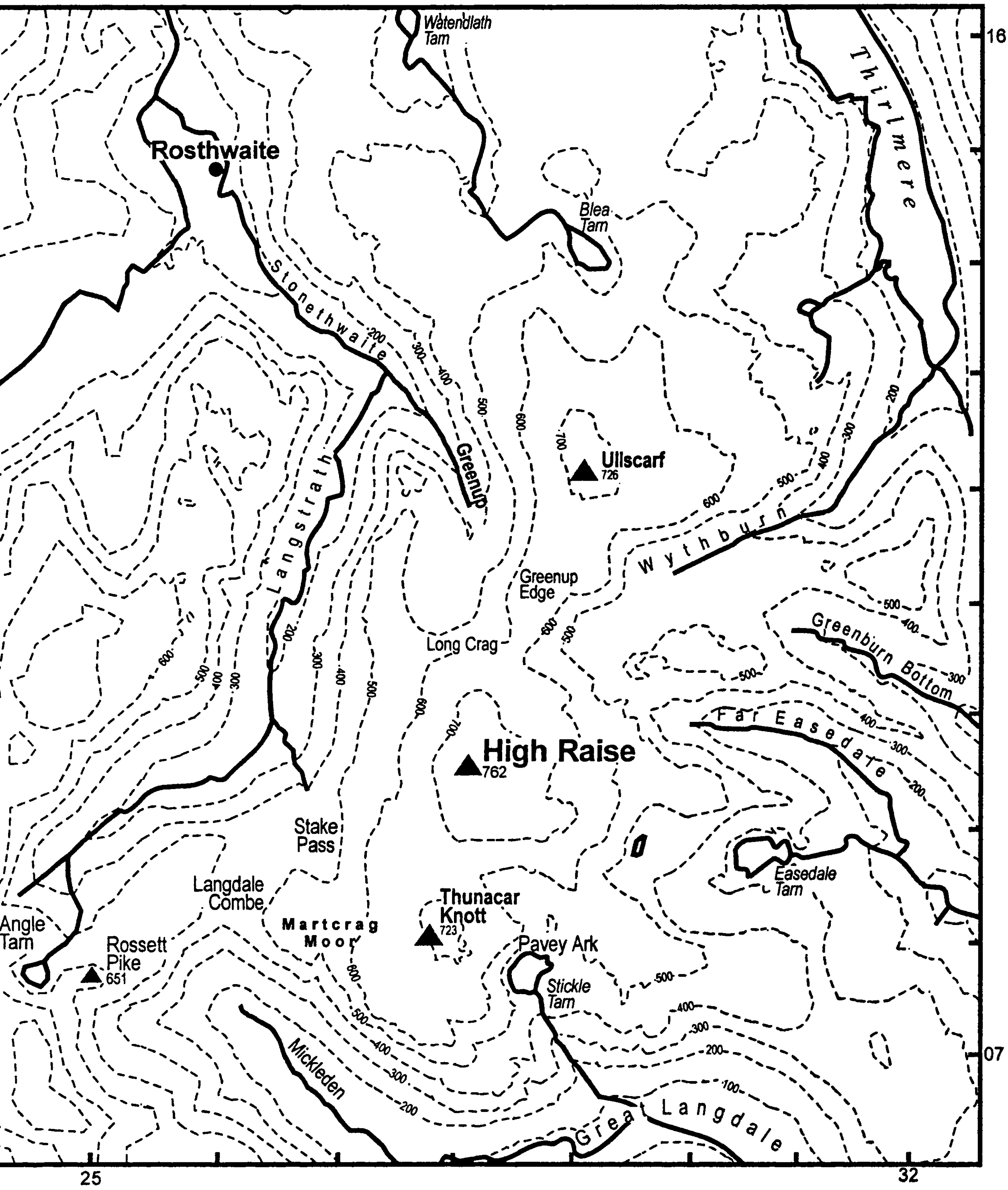


Figure 4.1 Location and topography of High Raise, Ullscarf and Thunacar Knott. Scale and orientation given by National Grid References (NY) at 1 km intervals.

4.2 EVIDENCE FOR A FORMER PLATEAU ICEFIELD

4.2.1 Summit Geomorphology

The broad, peat-covered summits of High Raise, Thunacar Knott and Ullscarf lack any substantial surface relief, with bedrock outcrops more or less restricted to the margins. Moraines and meltwater channels appear to be absent, although small-scale features may be masked by peat. In the vicinity of the summits (*sensu stricto*) of both High Raise and Thunacar Knott, the peat gives way to blockfields (Figure 4.2). These are developed in volcanic sequences which include dacitic lapilli tuffs, volcaniclastic sandstones, and conglomerates (BGS, 1996). Excavations near the summit of High Raise indicate a minimum blockfield depth of 0.4 m.

The boundaries of these blockfield islands on High Raise and Thunacar Knott are sharply defined in some places. Field investigations have shown this to reflect variations in peat development, and that High Raise, Thunacar Knott and Ullscarf probably have extensive blockfield covers which are mostly obscured by peat. Where the peat thins, such as in erosional hollows or on steeper slopes (the eastern slopes of Stake Pass, for example), blockfield either lies at or just below the surface (Figures 4.3 and 4.4).

Towards the margins of these summits, bedrock outcrops exhibit various degrees of streamlining. The most impressive examples occur north of High Raise summit at Long Crag (NY278105), where abrasion and quarrying has produced streamlined bedrock in a direction consistent with ice streaming off the plateau and into the Greenup basin below (Figures 4.5 and 4.6). Elsewhere the extent of ice-moulding is less obvious and is often best appreciated from aerial photographs, such as is the case with the bedrock outcrops above Pavey Ark. In all cases, the direction of streamlining is consistent with an icefield draining into the surrounding valleys.

Reconstructed ice-marginal positions in some of the surrounding valley-heads provides unequivocal evidence for the development of a Loch Lomond Stadial plateau icefield which extended from Thunacar Knott to Ullscarf (Section 4.2.2). Streamlined bedrock outcrops at the margins of High Raise, Thunacar Knott and Ullscarf demonstrate that the ice at these localities was at pressure melting point and sliding over the substrate.

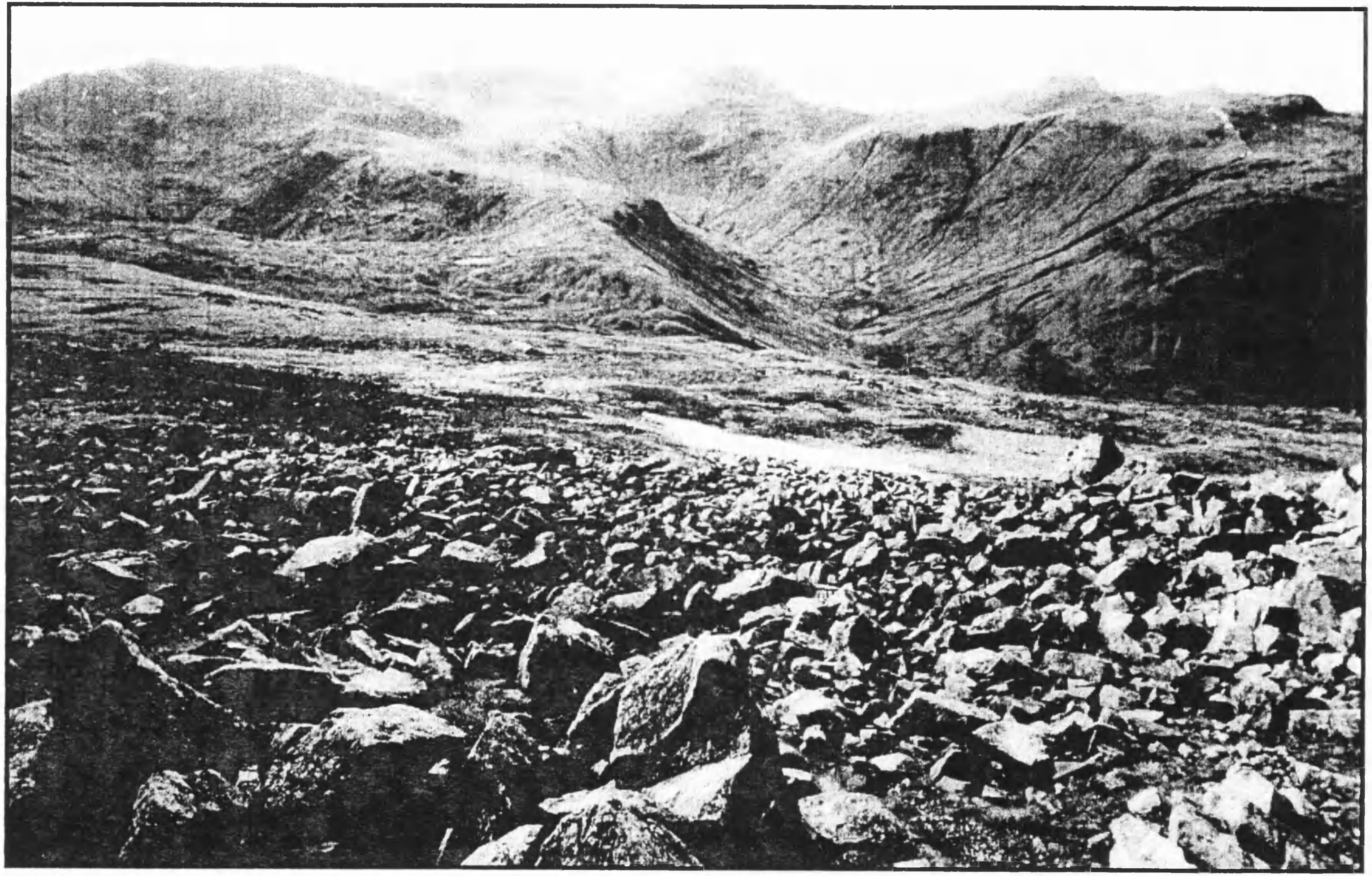


Figure 4.2 Blockfield on High Raise summit, looking west towards Langstrath

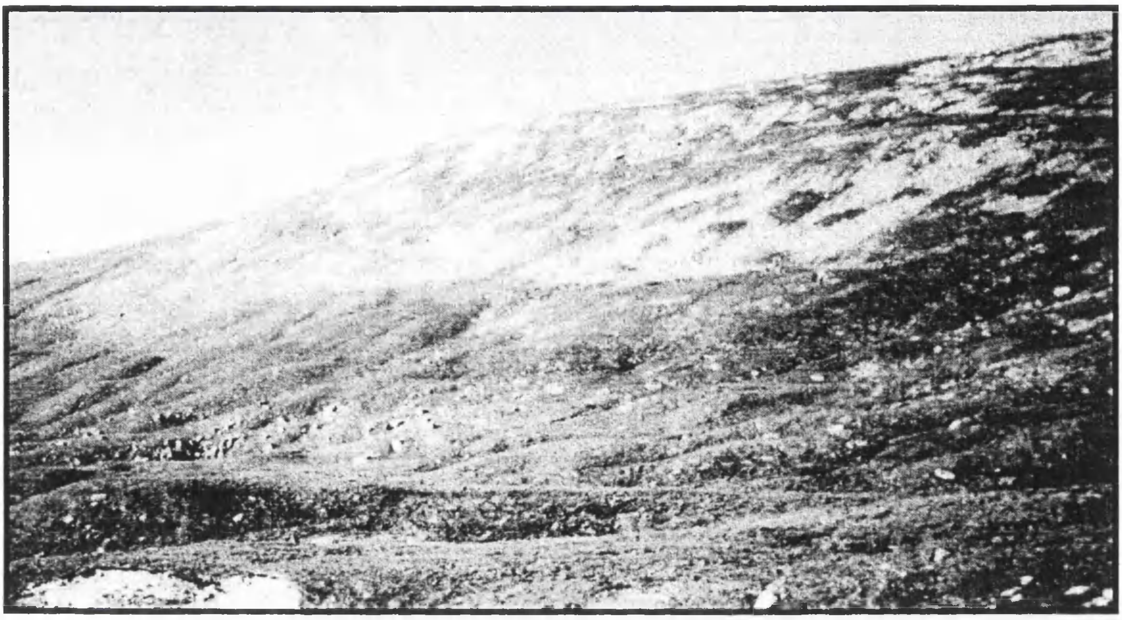


Figure 4.3 Blockslopes on the western slopes of High Raise-Thunacar Knott (Stake Pass)

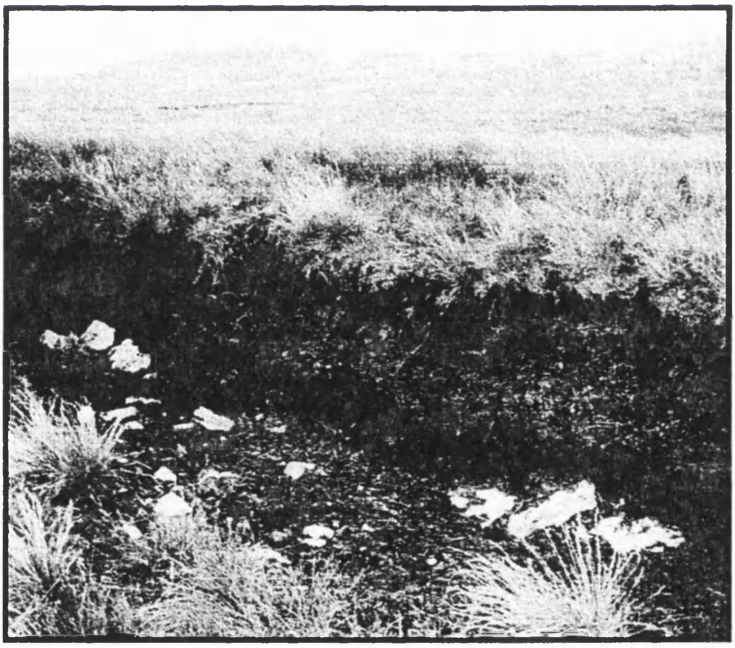


Figure 4.4 Blanket peat overlying blockfield south of Ullscarf.

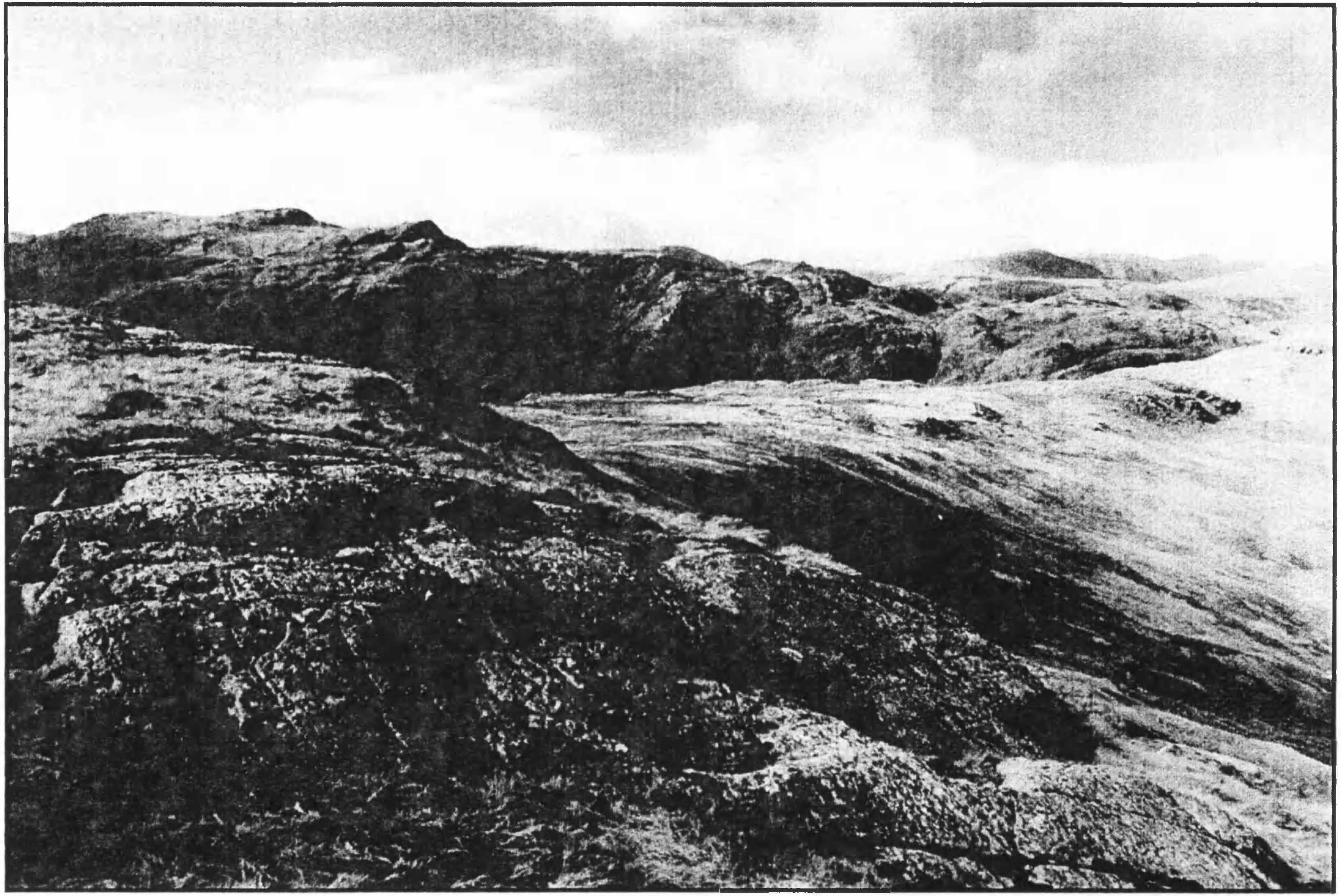


Figure 4.5 Ice-moulded bedrock at Long Crag (NY278105).
Ice flowed from left to right across the photograph, from High Raise into the Greenup basin.

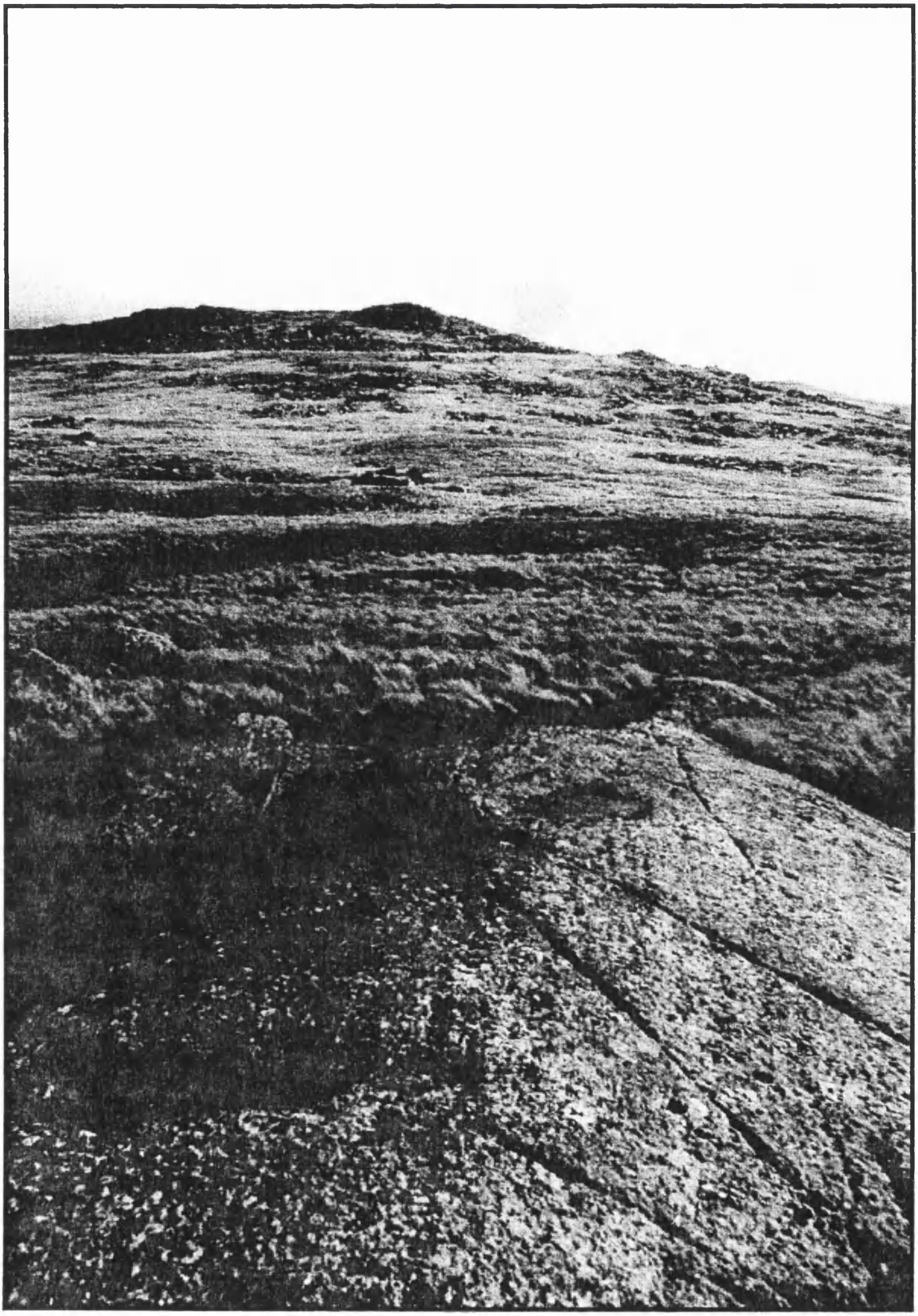


Figure 4.6 View looking south from Long Crag towards the summit of High Raise. Ice flow was towards the camera. Note the blockfield in the distance (Low White Stones) and the ice-moulded bedrock in the foreground (camera case for scale).

Nevertheless, the survival of an extensive blockfield cover implies the maintenance of cold-based, protective ice. The temperate conditions of the present interglacial precludes the possibility that this blockfield post-dates the Loch Lomond Stadial (e.g. Ballantyne and Harris, 1994). Thus, the summit geomorphology indicates that this plateau icefield was polythermal.

4.2.2 Valley-floor evidence

Although the geomorphological impact of this icefield appears to have been minimal on the summits of High Raise, Thunacar Knott and Ullscarf, prominent moraine systems were produced by outlet glaciers which descended into the surrounding valleys where their margins became sediment traps for supraglacial debris and inwash. Due to the absence of periglacial trimlines and the effects of paraglacial resedimentation on valley sides and floors, the overall configuration of the High Raise plateau icefield system at maximum extent cannot be established unequivocally (Section 4.3). Nevertheless, it is clear from the evidence described below that deglaciation was interrupted by stillstands and/or readvances throughout, although some ice-marginal moraines and drift limits are extremely faint and best viewed on aerial photographs. Palaeo ice-margins of these outlet glaciers have been reconstructed which, at three key sites, demonstrate that decay was centred on the plateau itself rather than at the valley heads. This evidence was produced during the final stages of deglaciation.

4.2.2.1 Stake Pass and Langdale Combe

Stake Pass and Langdale Combe (NY2608) are minor topographic depressions located southwest of High Raise summit (Figure 4.1). Langdale Combe is a shallow, south-facing hollow at an altitude of approximately 450 m OD, and has a 'hanging valley' relationship with Mickleden below. Immediately to the east is Stake Pass, a shallow, open valley which descends from approximately 500 m OD in the south to about 450 m OD at the point where it opens out onto the slopes above Langstrath.

Suites of hummocky recessional moraines in Stake Pass and Langdale Combe, palynologically dated to the Loch Lomond Stadial by Walker (1965), define successive margins of an outlet glacier which actively-backwasted towards its source on High Raise, perpendicular to the valley axes (Figures 4.7). Their lobate outlines and well-developed

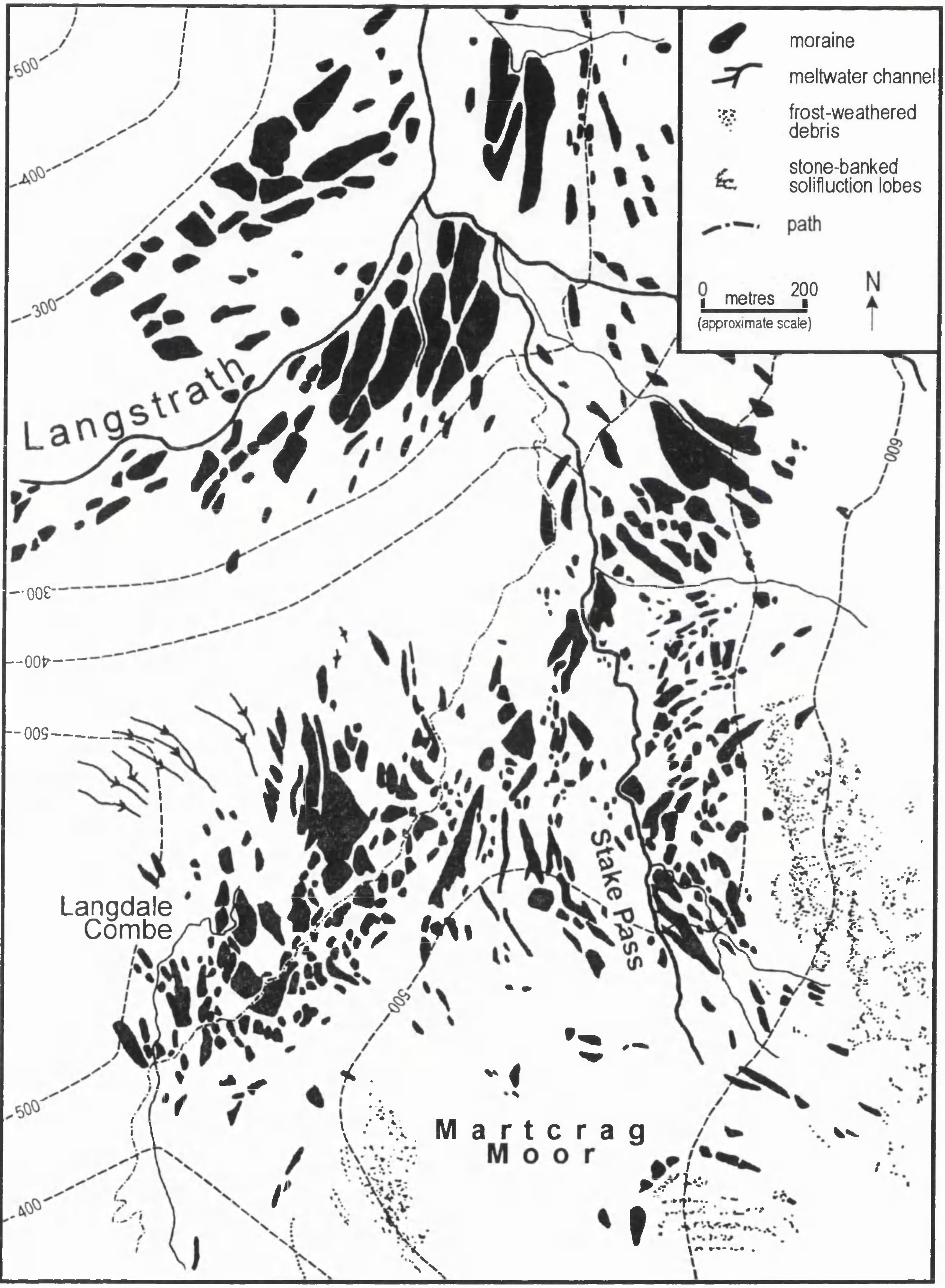


Figure 4.7 Map of Loch Lomond Stadial moraines in Stake Pass and Langdale Combe

bifurcations are characteristic planimetric properties of ice-marginal moraines produced by actively-retreating glaciers in upland areas (e.g. Price, 1973; Bennett and Boulton, 1993a, 1993b). The absence of a regular, 'nested' ice-marginal sequence reflects complex glacier-topographic interactions during deglaciation, the principal effect of which was to induce differential retreat during periods of moraine formation. This is evidenced by the pronounced moraine bifurcations which occur in the area.

The moraines in Stake Pass and Langdale Combe vary considerably in their development, being most prominent on the floor of Langdale Combe where they rise 4-5 m above the surrounding peat infill, and most subtle on Martcrag Moor (Figure 4.8). It is proposed that these variations in moraine morphology primarily reflect variations in debris supply. The ice which occupied Stake Pass and Langdale Combe carried no supraglacial debris from its source area and consequently debris for moraine development had to be derived subglacially and/or proglacially. Clasts extracted from shallow surface pits were predominately sub-angular to sub-rounded, implying active transport and thus wet-based ice (meaningful analysis of clast shape was prevented owing to an absence of control data in the area) (Figure 4.9). It seems likely that more debris would have been available for entrainment on the floors of Langdale Combe and Stake Pass (perhaps due to debris inwash) than on the adjacent higher ground of Martcrag Moor, and this would account for at least some of these variations in moraine development. There are locally high concentrations of boulders on the western slopes of Stake Pass, with some moraines appearing to be entirely composed of boulders (Figure 4.10). The blockslopes on the eastern slopes of Stake Pass represent the most likely source of these boulders.

It is clear from evidence considered later in this chapter (Section 4.3.1) that these moraines were produced at a relatively advanced stage in deglaciation. Palaeo-iceflow directions can be inferred from reconstructed ice-marginal positions for the period of deglaciation represented by these moraines (Figure 4.11). Most of the ice descending from the summit was evacuated to the north to become confluent with the Langstrath palaeoglacier, with excess ice spilling out of this rather shallow valley and advancing westwards over the irregular topography to inundate Martcrag Moor and Langdale Combe. For a time, this was supplemented by ice flowing into Langdale Combe from the

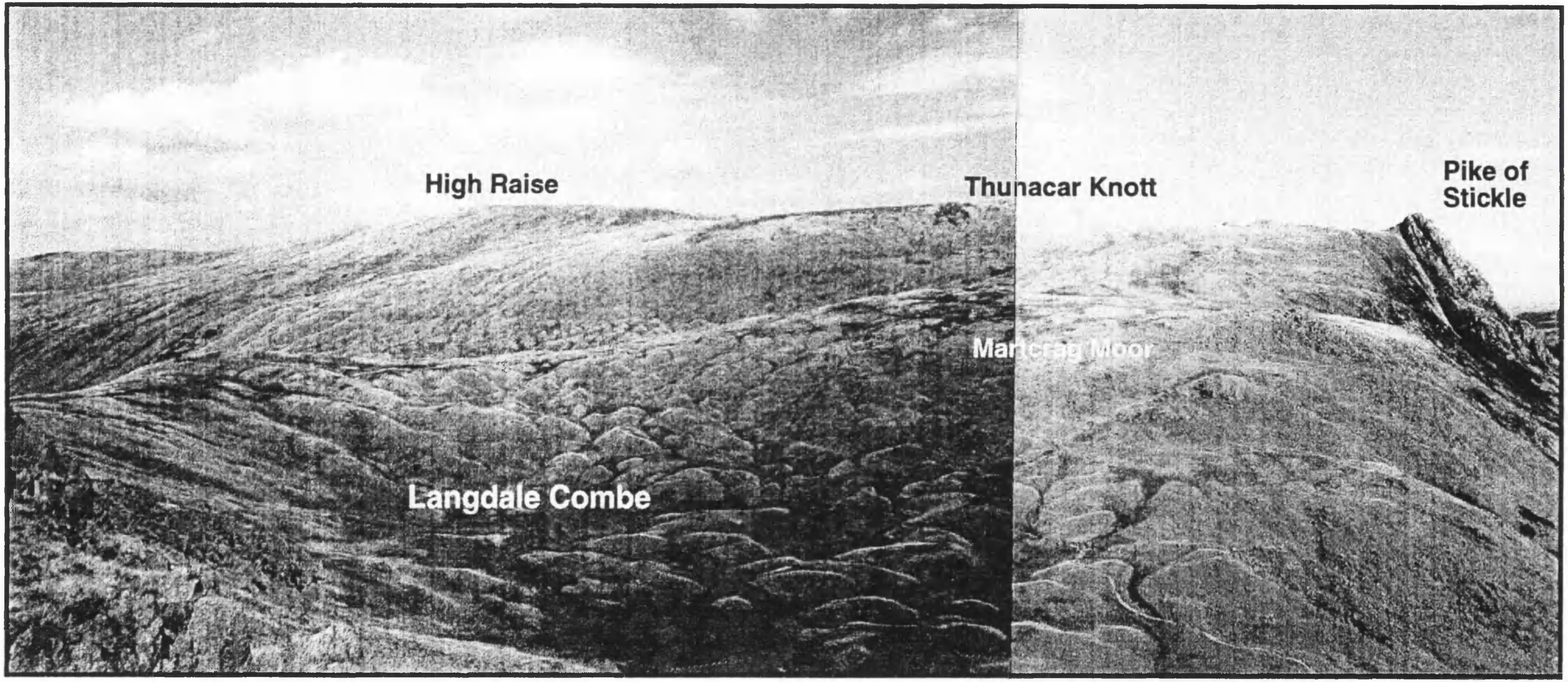


Figure 4.8 Langdale Combe and Stake Pass

View looking northeast from Mansey Pike (NY258084) towards High Raise and Thunacar Knott in the distance.

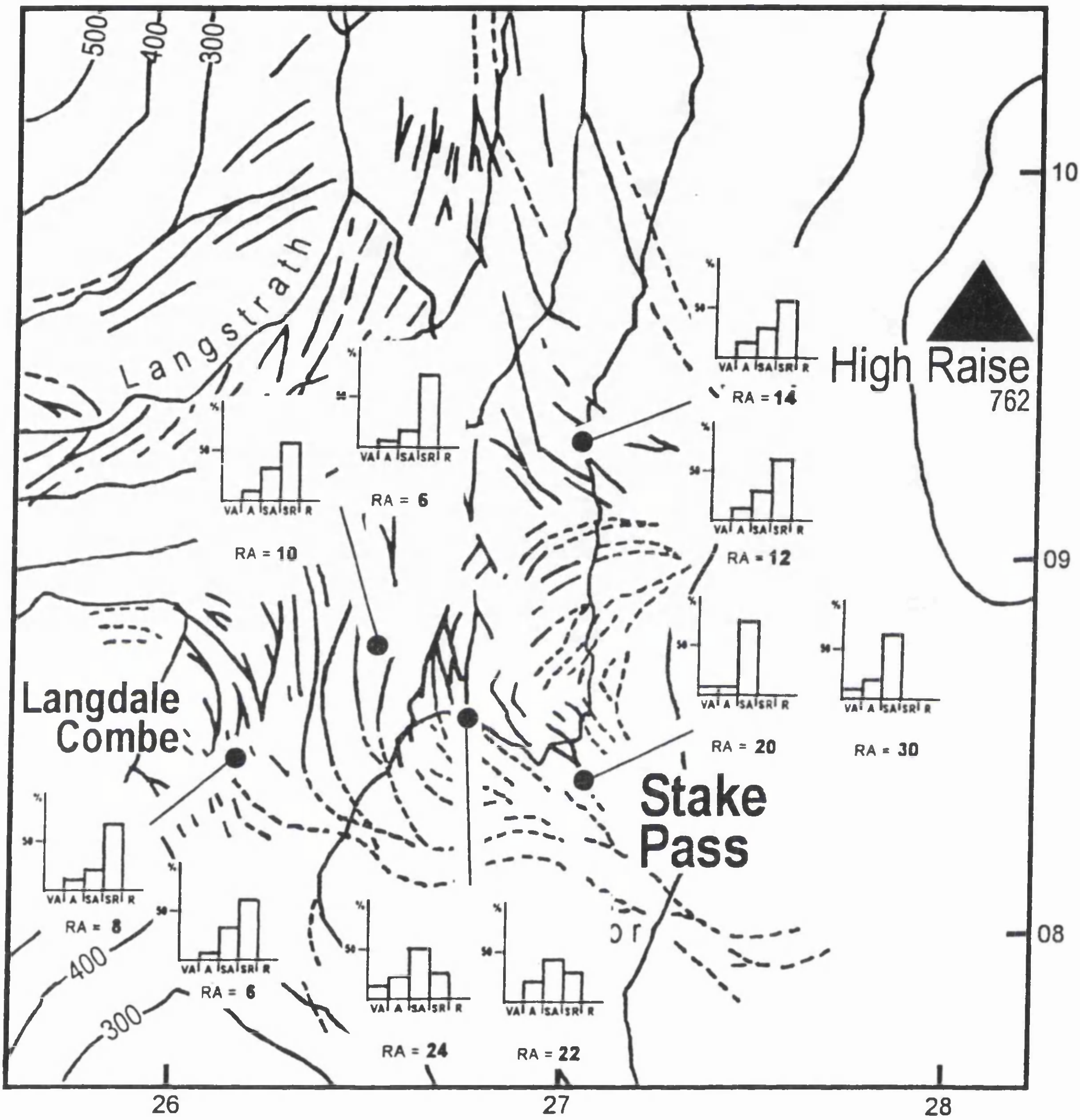


Figure 4.9 Clast roundness characteristics of moraines in Stake Pass and Langdale Combe. Clasts extracted from shallow surface pits with a sample size of thirty clasts per pit. Scale and orientation given by National Grid References (NY) at 1 km intervals.

Figures 4.10 Boulder-strewn moraines on the western slopes of Stake Pass

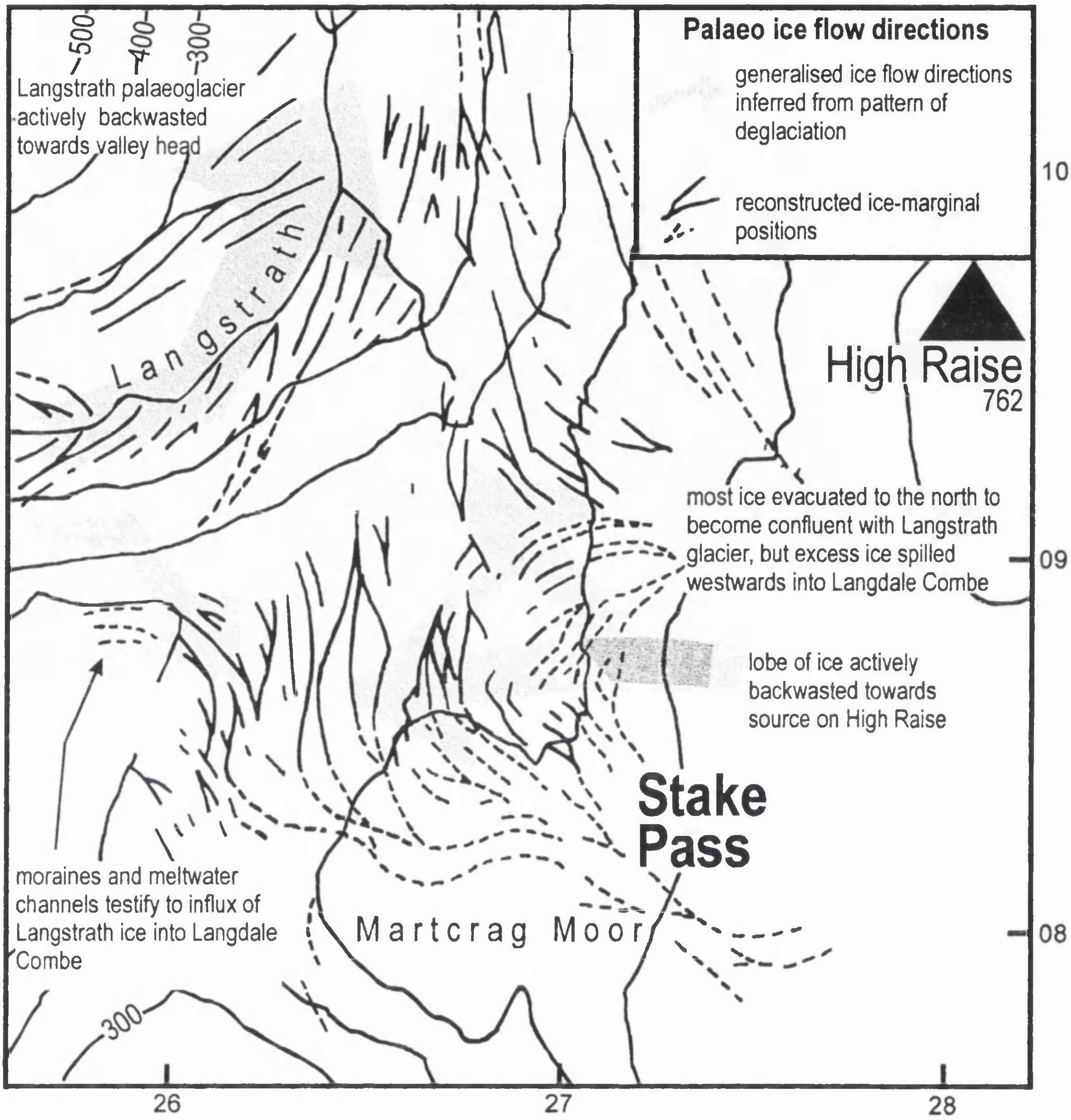


Figure 4.11 Palaeo-iceflow directions and reconstructed ice-marginal positions in Stake Pass and Langdale Combe
Scale and orientation given by British National Grid References at one kilometre intervals

Langstrath palaeoglacier. This is inferred from ice-marginal meltwater channels on the ridge which defines the northern margin of Langdale Combe and implies a substantial volume of ice in Langstrath. Indeed, at maximal extents, there is some evidence to suggest that the Langstrath and Mickleden palaeoglaciers were confluent, leaving Rossett Pike (NY249076, not shown on Figure 4.1) as a nunatak. This takes the form of a blockfield cover on Rossett Pike, which extends downslope on the Langstrath side to a limit defined by lateral moraines, plus ice-scoured and ice-moulded bedrock over a section of ridge which divides Langstrath and Mickleden (Figures 4.12 and 4.13).

The significance of the Stake Pass and Langdale Combe moraines is difficult to appreciate in the field because of their complex arrangement, a situation which to a considerable extent reflects both variations in debris supply and the influence of an uneven topography on ice margin behaviour during deglaciation. This is compounded by a 'drift limit' on the eastern slopes of Stake Pass (separating hummocky, recessional moraines below from blockslopes above) trending downvalley for a short distance in the direction of Langstrath (shown on Figure 4.7 - the downslope boundary of the 'frost weathered' material).

4.2.2.2 Pavey Ark

The Pavey Ark area (NY2807) was completely inundated by ice flowing off the High Raise-Thunacar Knott summit during the Loch Lomond Stadial. Most of the ice descending from the summit would have been channelled down the Bright Beck valley (immediately northeast of Stickle Tarn), although it is likely that, for a time, an icefall developed above Stickle Tarn. The only direct evidence for this latter suggestion takes the form of weakly-streamlined bedrock at the top of Pavey Ark crags, best viewed on aerial photographs.

The channelling of plateau ice down the Bright Beck valley is evidenced by the ice-marginal landforms produced during deglaciation (Figures 4.14 and 4.15). In the lower reaches of the valley, hummocky recessional moraines define the margins of a narrow, lobate ice mass which actively backwasted towards the summit. As deglaciation proceeded, the supraglacial debris load would have decreased. This may account for the very faint nature of the ice-marginal evidence in the upper reaches of this valley,

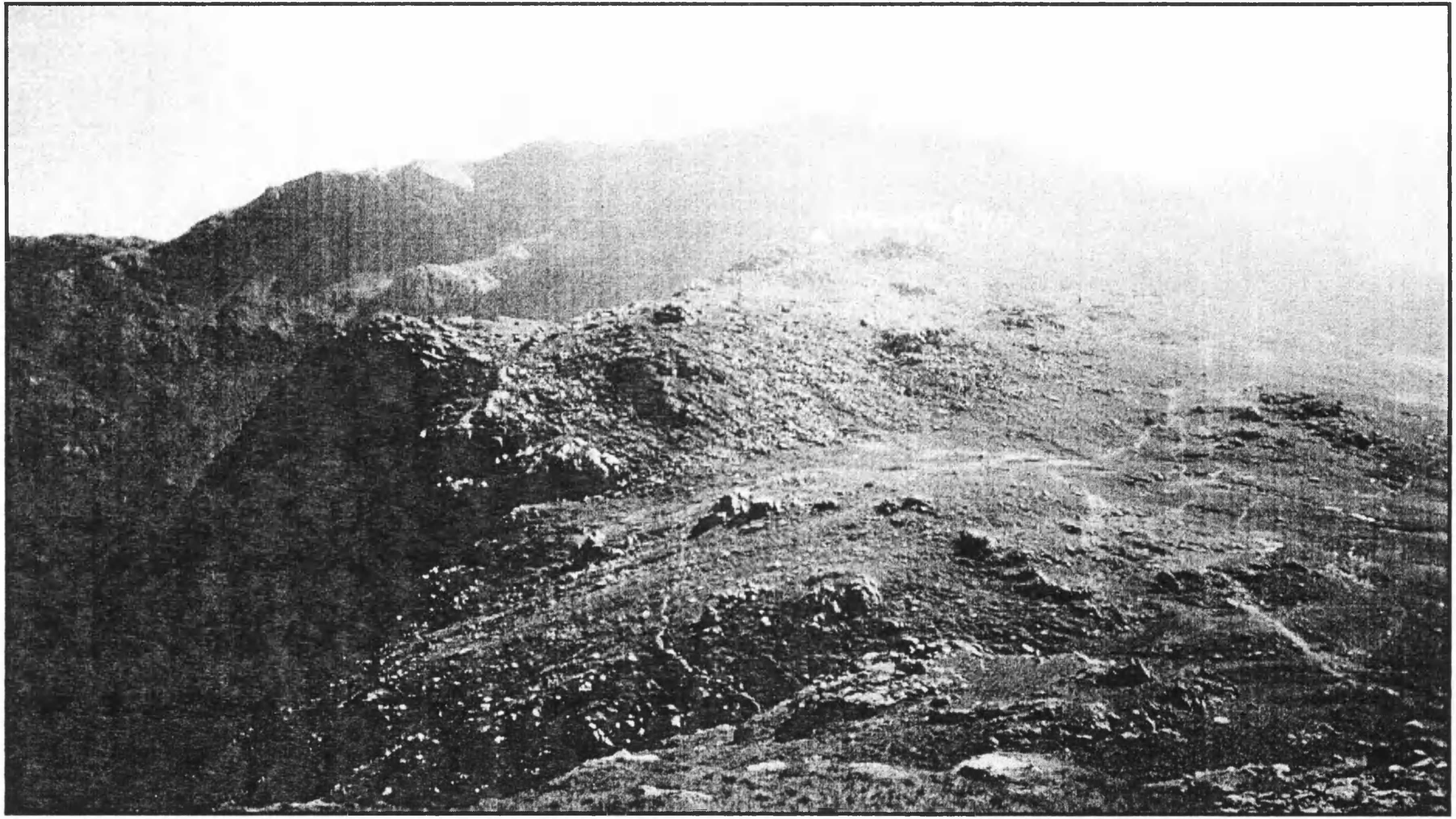


Figure 4.12 View looking towards Rossett Pike blockfield, with evidence of ice scouring in the foreground.

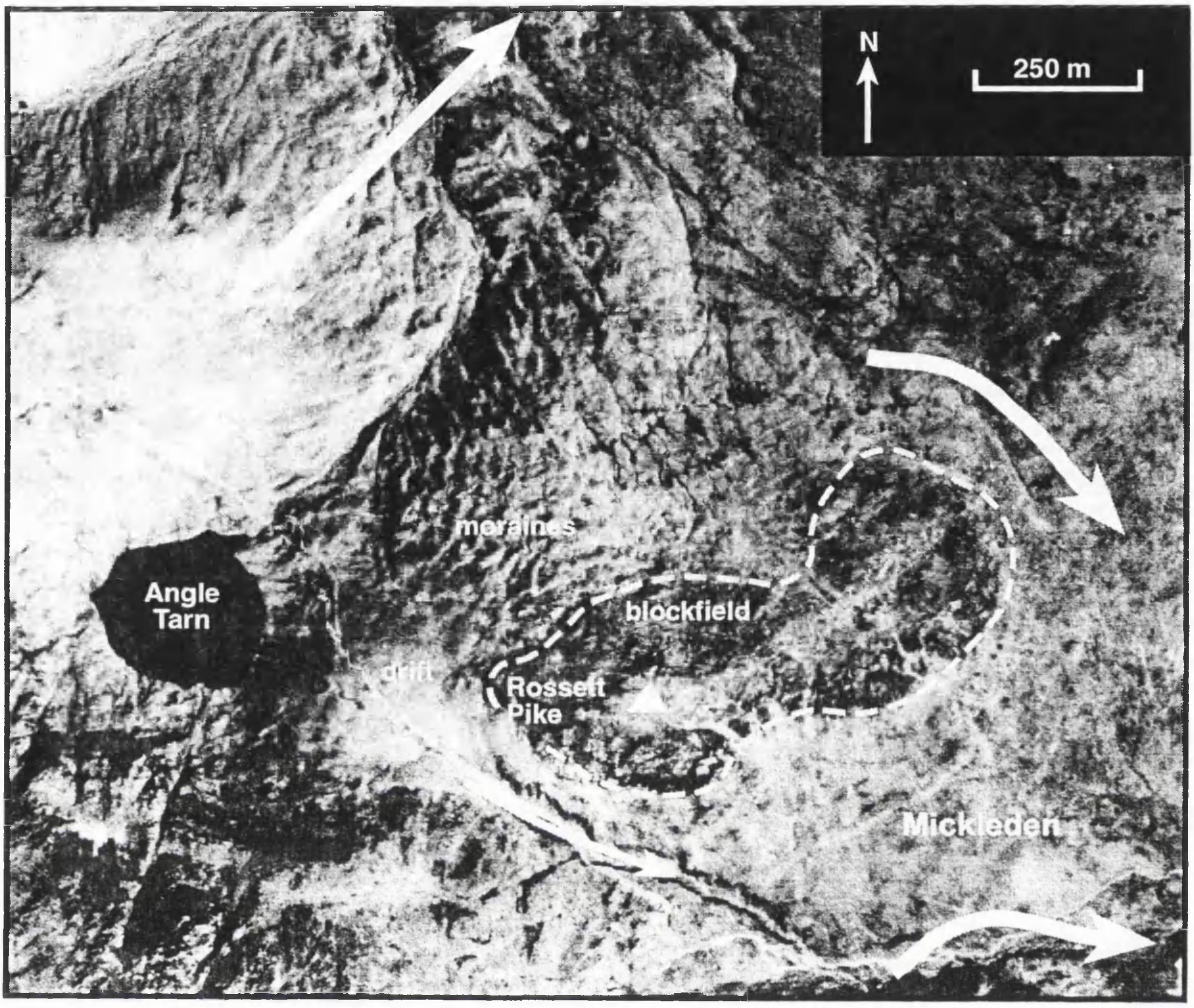


Figure 4.13 Extract from aerial photograph showing Rossett Pike.
White arrows indicate former ice flow direction. See text for explanation.

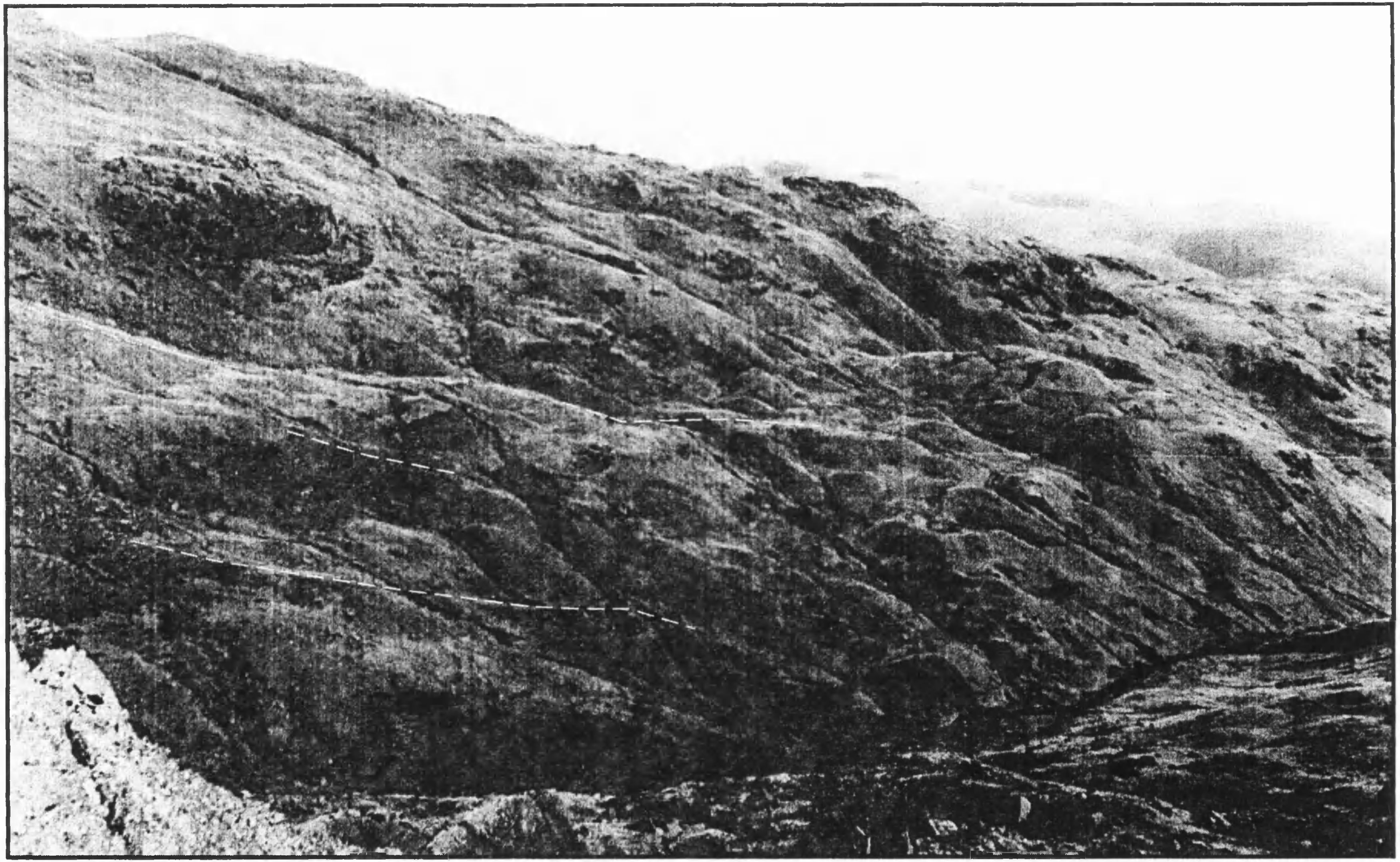


Figure 4.14 Loch Lomond Stadial moraines in Bright Beck, Pavey Ark

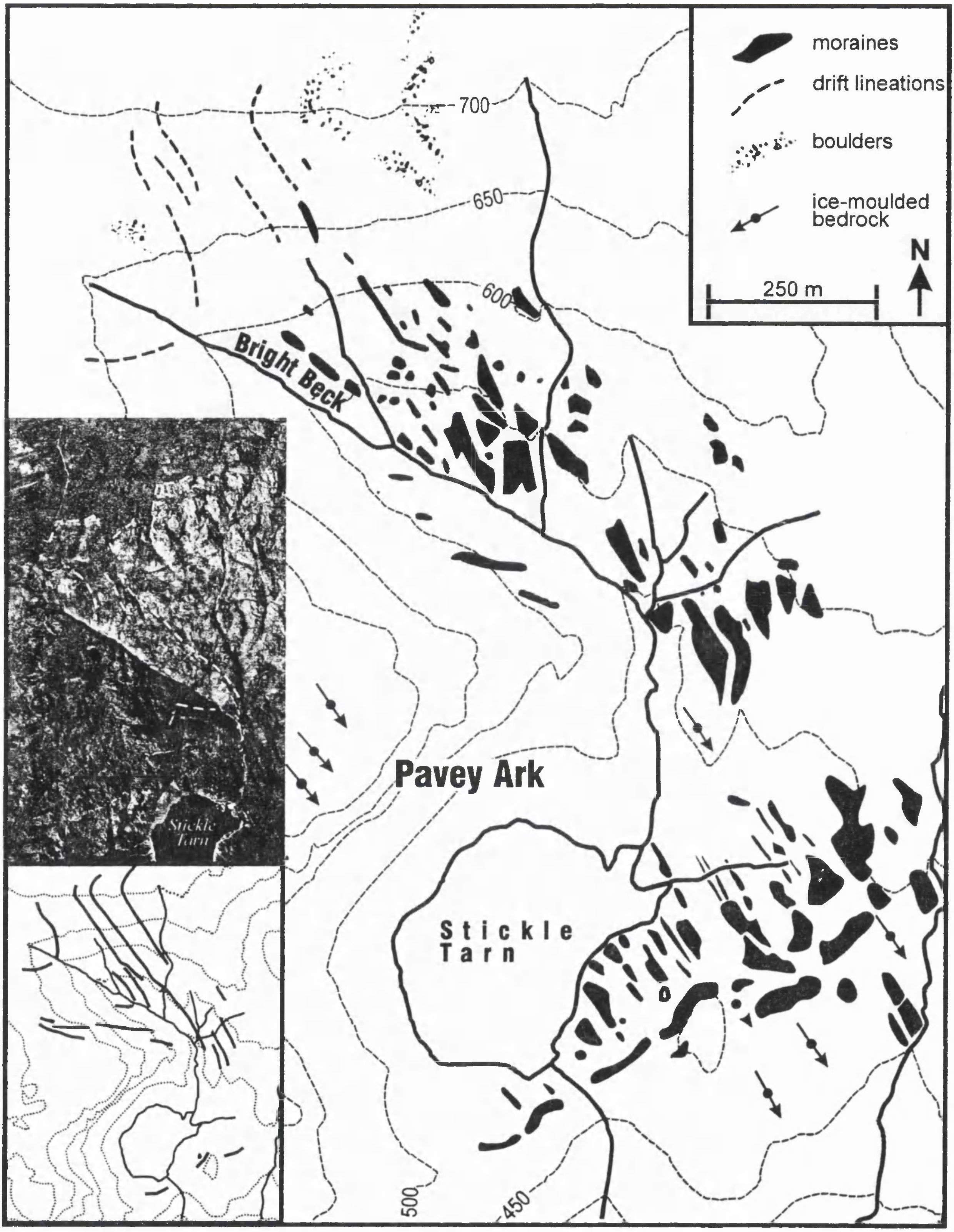


Figure 4.15 Map of Loch Lomond Stadial moraines in Pavey Ark
Ordnance Survey Aerial Photograph (420 72/262) Crown Copyright Reserved.

essentially represented by subtle variations in drift thickness which can only be identified with confidence using aerial photographs. Nevertheless, these can be traced onto the margins of the summit where they define, for a short distance, the margins of the former plateau icefield. East of these subtle drift variations, clearer examples of ice-marginal positions are represented by several discontinuous, curved boundaries between drift/peat and boulder concentrations (Figure 4.16). The restriction of hummocky recessional moraines to the Bright Beck valley may reflect a change in the style of deglaciation. It is equally possible, however, that this distribution reflects the input of supraglacial debris to the ice-margins once these slopes became exposed during deglaciation.

Beyond the mouth of Bright Beck valley is an area of streamlined bedrock and low hummocky moraines, some of which are elongated and parallel with the direction of streamlining (which is downslope towards Great Langdale). These are interpreted on morphological grounds as small flutings (Figure 4.17). Sections are absent in the area but clasts extracted from pits were predominately blocky and sub-angular to sub-rounded (Figure 4.18). Although this differs little from the scree control samples in terms of clast shape, roundness was more marked in the moraine samples. These results imply active transport. The genesis of some of the ridges which share broadly similar orientations on the higher slopes to the east is less certain; they are slightly curved in planform and are probably lateral moraines. There is also an end moraine, composed mainly of boulders, a short distance southeast of the Tarn (Figure 4.19). However, the occurrence of streamlined bedrock and further moraines downslope from this end moraine demonstrates that this does not demarcate the maximal extent of this outlet glacier. There is no evidence in the area which can reasonably be interpreted as representing the maximal extent of Loch Lomond Stadial ice. In fact, it is unlikely that ice-marginal moraines would have formed on the relatively steep slopes which descend to Great Langdale. Possible glacier extents are considered in Section 4.3.

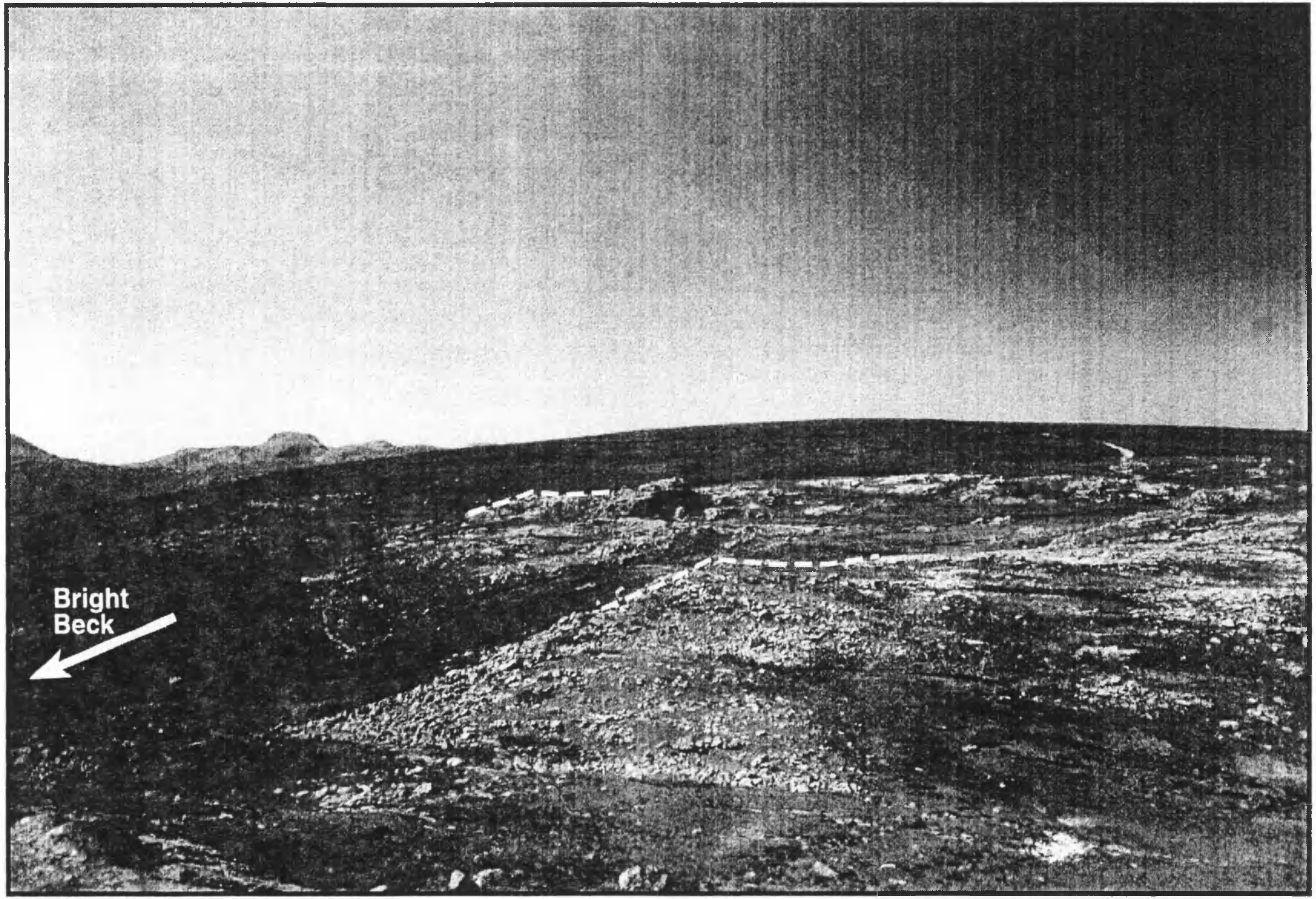


Figure 4.16 Looking north towards High Raise summit from Sergeant Man (NY286089)
The boundaries of the frost-weathered debris appear to define successive palaeo ice-margins associated with the Pavey Ark outlet glacier. These are consistent with reconstructed ice-margins in the Bright Beck valley.

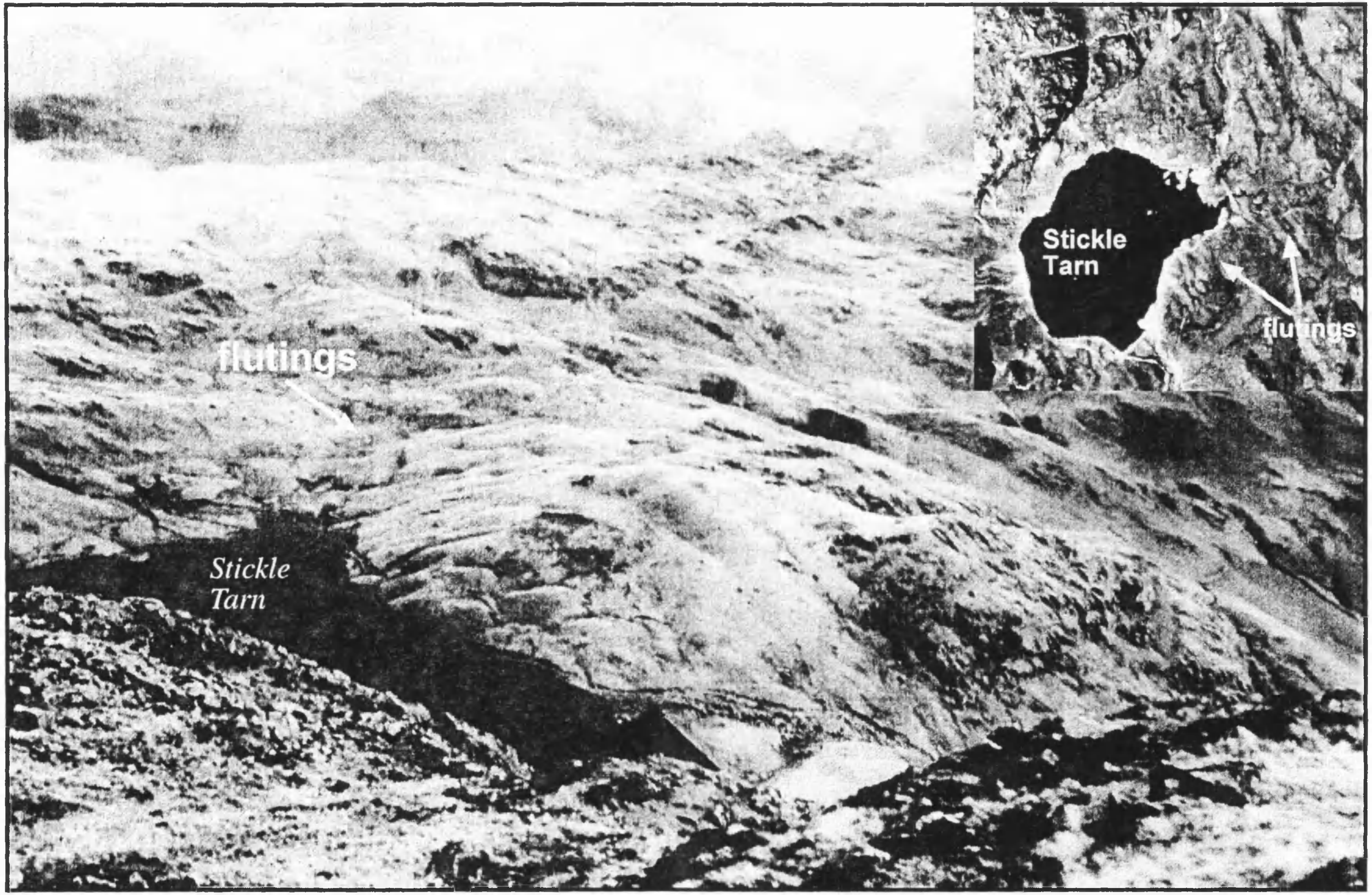


Figure 4.17 Streamlined moraines south of Stickle Tarn, Pavey Ark

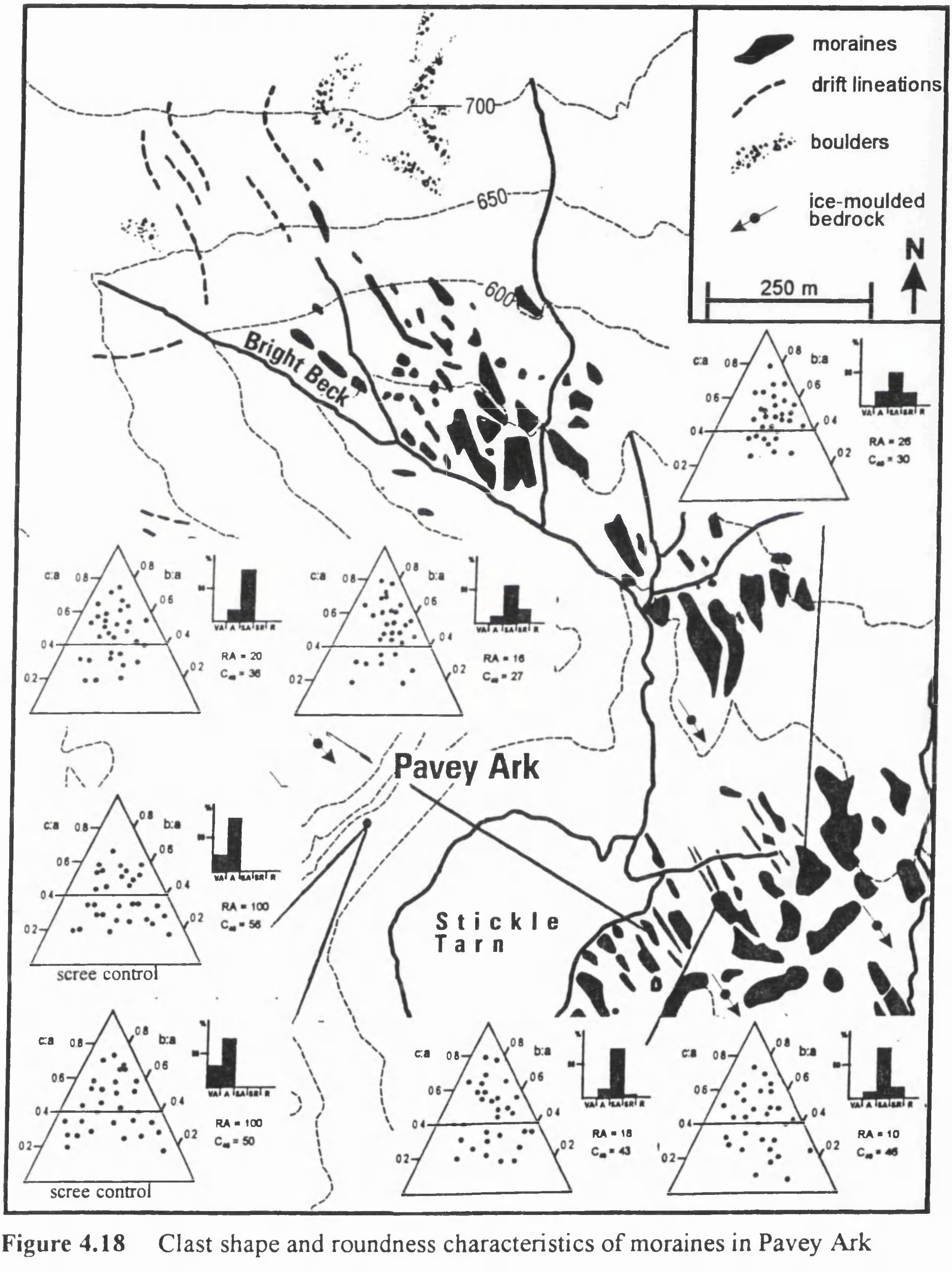


Figure 4.18 Clast shape and roundness characteristics of moraines in Pavey Ark

Figure 4.19 Boulder strewn end moraine southeast of Stickle Tarn, with ice-moulded bedrock beyond.
The presence of ice-moulded bedrock beyond the bouldery end moraine suggests that the latter does not delineate the maximal extent of this former outlet glacier.

4.2.2.3 Greenup

The north-facing Greenup basin (NY2711) is located a short distance north of High Raise summit. It has a relatively broad, gently sloping floor (which mostly lies between 400-450 m OD) on which glaciogenic landforms are well developed. This upland basin is linked to the Stonethwaite valley below by a rather narrow, steep-sided valley which curves around to the northwest on its descent (Figure 4.1).

The glaciogenic landforms on the floor of Greenup comprise a mixture of mounds and ridges, typically rising 3-4 m above the surrounding terrain, and which collectively convey a chaotic appearance (Figure 4.20). Aerial photographs with good tonal contrasts reveal that, although some moraines appear to lack any orientation whatsoever, some of the glaciogenic landforms occur within large scale linear bands of varying width which adopt lobate, bifurcating planforms (Figures 4.21). These features are interpreted as having been produced in association with an actively-backwasting ice-margin, although the overall hummocky and chaotic appearance suggests localised ice core stagnation. As with all the other sites investigated, the assessment of moraine genesis is hampered by the absence of sections.

A decay centre on High Raise summit is suggested by the existence of faint palaeo ice-marginal positions at the head of the basin which trend up the valley sides and onto the edge of the plateau. Distinct ice-marginal moraines on the lower slopes give way to much fainter drift lineations near the top. In the field, these ice-marginal positions cannot be identified with confidence, although the threshold between Greenup basin and the High Raise plateau beyond is characterised by a gullied drift cover. This, together with the streamlined bedrock outcrop of Long Crag, is consistent with a plateau icefield on High Raise (Section 4.2.1).

The geomorphological evidence suggests that the Greenup basin was more or less submerged at the maximal extent of the Loch Lomond Readvance by ice draining from the High Raise and Ullscarf plateau icefields. For example, the prominent lateral moraines north of the river at the mouth of Greenup trend obliquely upslope at a uniformly steep angle. When viewed on aerial photographs, these can be seen to extend

Figure 4.20 Extract from aerial photograph showing Greenup moraines. For interpretation, see Figure 4.21.

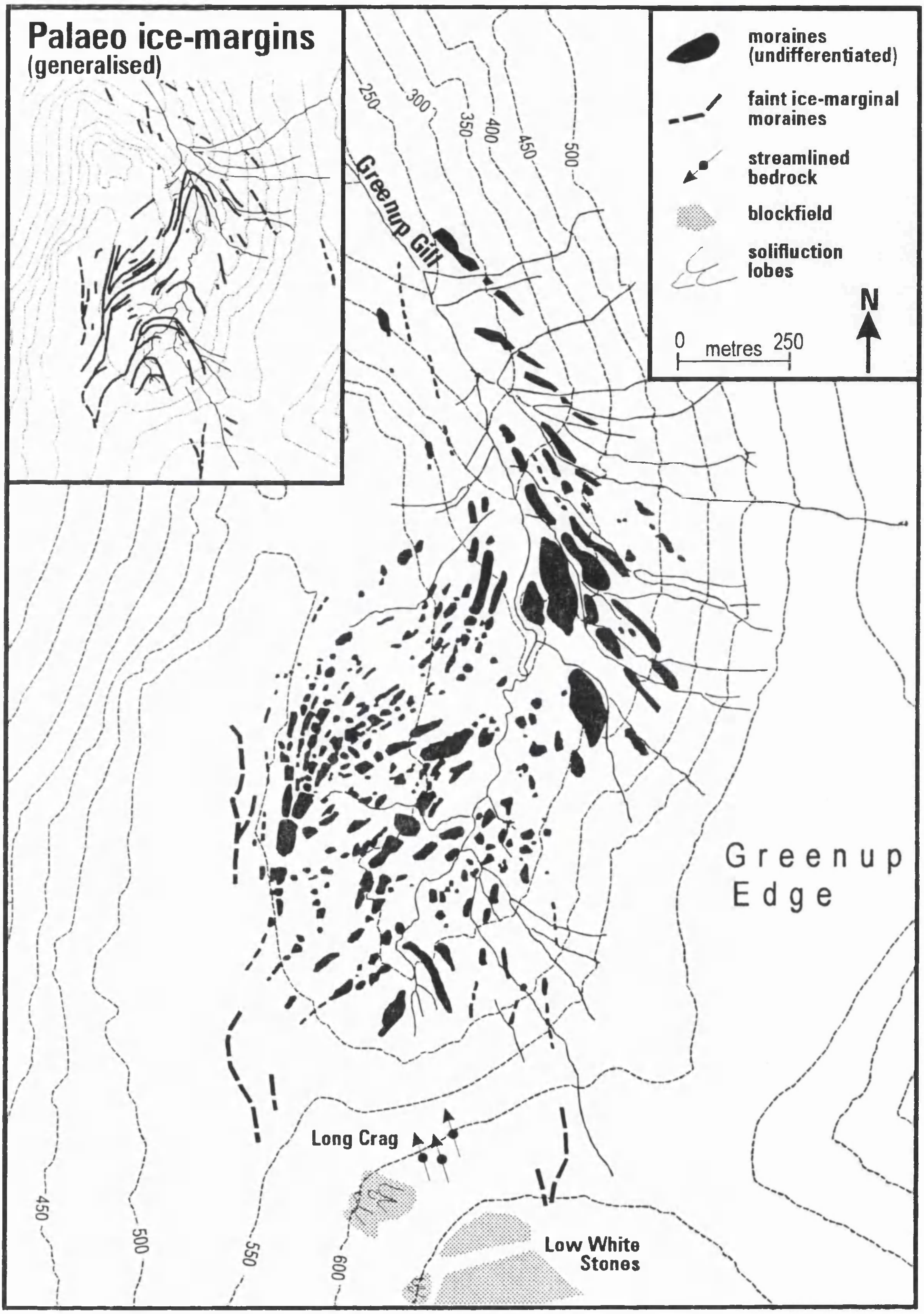


Figure 4.21 Map of Loch Lomond Stadial moraines in the Greenup basin

beyond the confines of the basin and onto the gentler slopes above, on the southern flanks of Ullscarf. Furthermore, these form part of an areally restricted sequence of ice-marginal moraines and ice-marginal meltwater channels on the southwestern flanks of Ullscarf. These are mostly low mounds and ridges which appear to be composed mainly of boulders, although meltwater channels cannot be identified with confidence on the ground (Figure 4.22). They are consistent in terms of orientation with more distinct moraines in the Greenup basin and also with the more subtle ice-marginal moraines on the steep slopes of the narrow valley which links the basin with the Stonethwaite valley below. This sequence of ice-marginal moraines and meltwater channels also constitutes a line of evidence to support the existence of an icefield on Ullscarf during the Loch Lomond Stadial. The uppermost moraines and meltwater channels on the flanks of the mountain branch out towards the Watendlath valley to the northwest. The meltwater channels which branch in to this rather shallow valley can be traced down to Watendlath, where they are continued by an end moraine.

Additional evidence that Loch Lomond Stadial ice was not restricted to the Greenup basin can be found on the valley sides as far as Rosthwaite on the floor of Borrowdale. Short stretches of ice-marginal moraines, trending obliquely down valley, can be found in many places. For example, very clear lateral moraines occur on the slopes to the south of Seathwaite. For the most part, these are best viewed on aerial photographs with good tonal contrasts. This is particularly so for the moraines on the steep slopes just below the Greenup basin (Figure 4.23). The contrast between these and the prominent moraines on the floor of Greenup can be explained, at least in part, by the effects of mass wasting processes on the steep slopes following the withdrawal of the ice-margin. Such reworking is evidenced by thin debris cones which completely mantle these slopes. Thus, there is no sound reason for assigning the more subtle moraines on the steeper slopes to an older event than that represented by the more prominent moraines in the upper basin purely on the basis of their morphology (or apparent 'freshness').

4.2.2.4 Watendlath Fell

As outlined above, evidence for a plateau icefield on Ullscarf is provided by the configuration of the system of ice-marginal moraines and meltwater channels on the

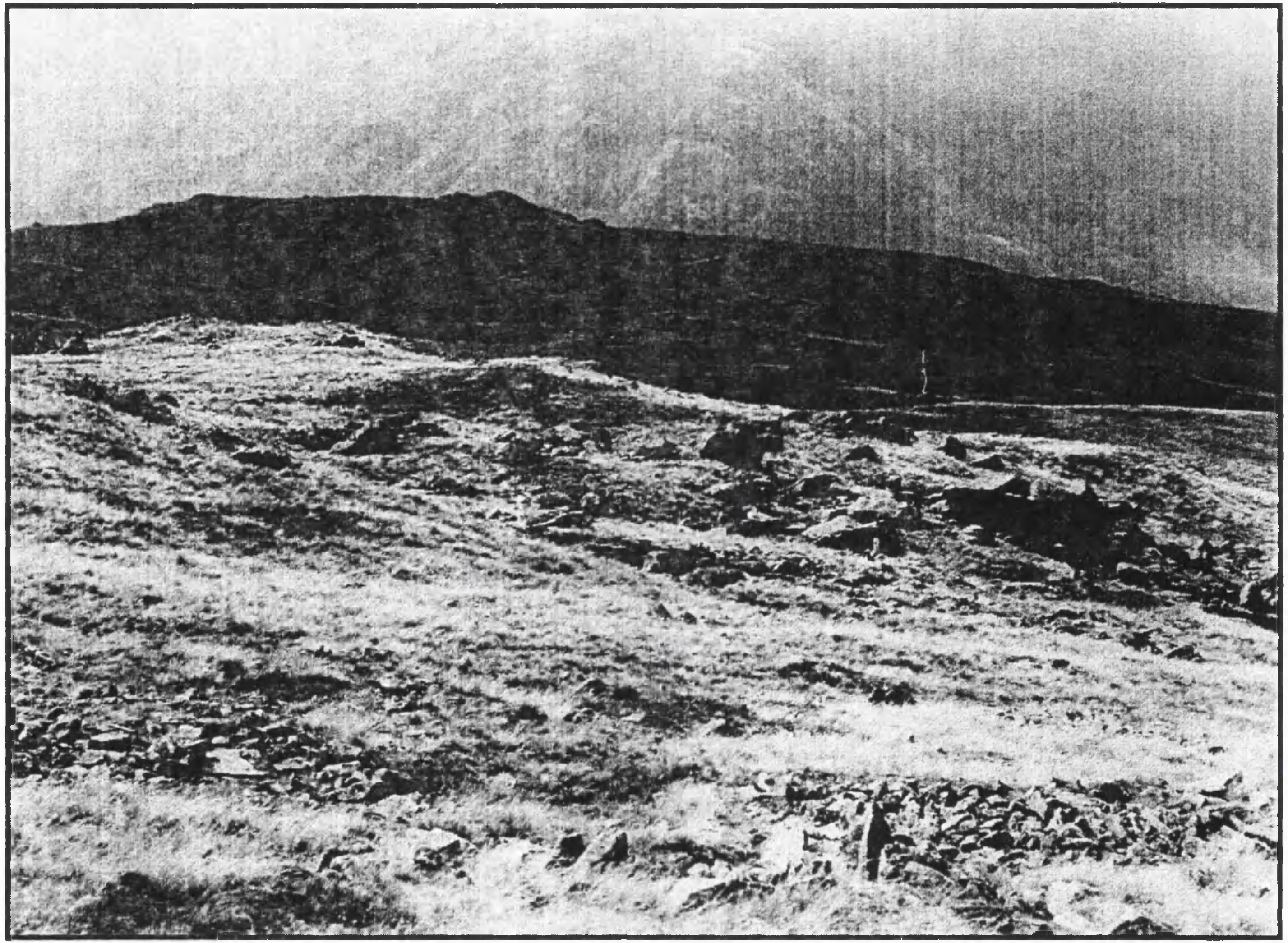


Figure 4.22 Bouldery moraines on the southwestern flanks of Ullscarf

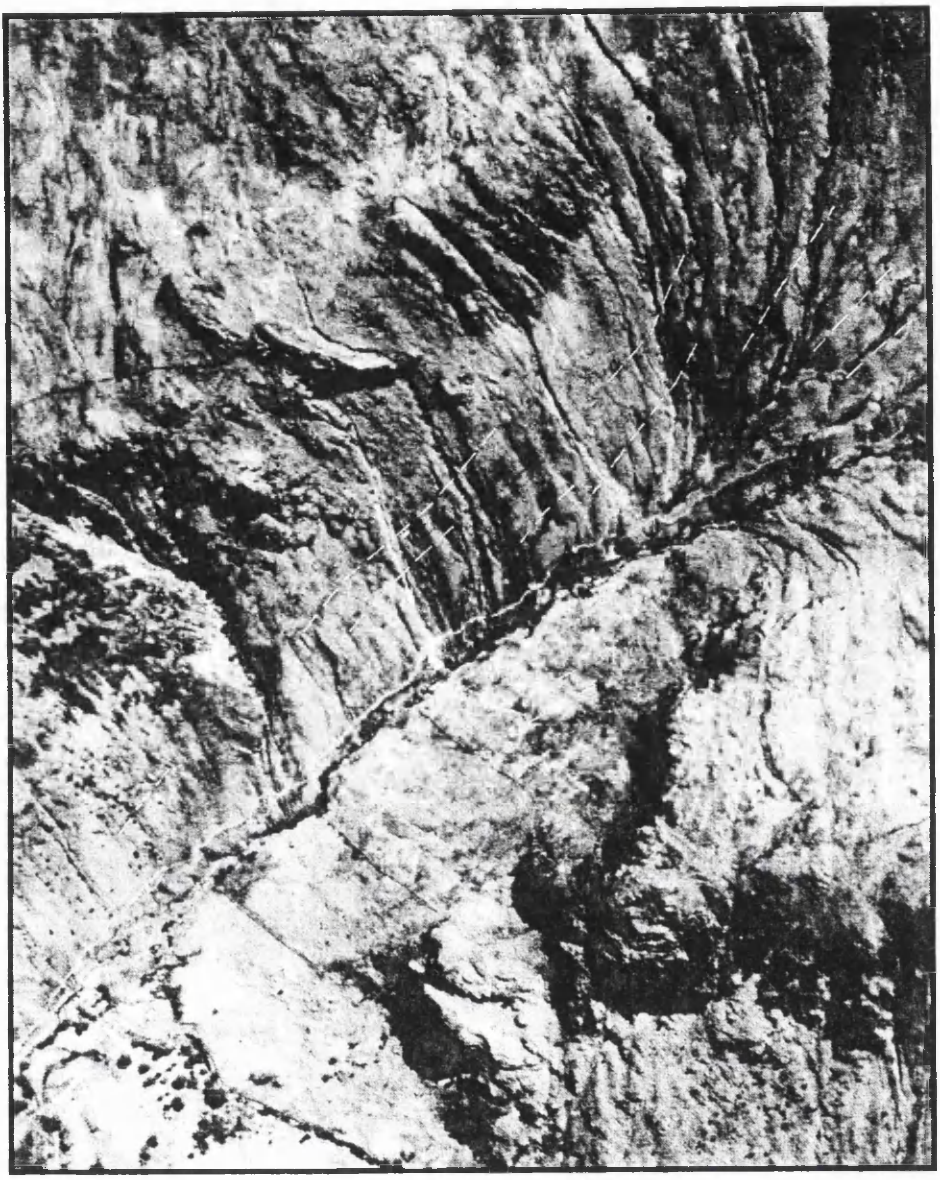


Figure 4.23 Extract from aerial photograph showing subtle lateral moraines below Greenup

slopes west and northwest of the summit, specifically the branching out of meltwater channels across Watendlath Fell in the direction of Watendlath. These meltwater channels are continued by end moraines at Watendlath (Figure 4.24). Contiguity of the ice masses which occupied Greenup and Watendlath Fell thus implies the former existence of an icefield on Ullscarf. Supporting evidence is provided by low moraine ridges on the slopes above Blea Tarn, less than one kilometre north of the summit. However these are fragmentary and do not permit the reconstruction of palaeo ice-margins during deglaciation.

4.2.2.5 Wythburn valley

The Wythburn valley contains no direct evidence for a plateau icefield on High Raise or Ullscarf, although lateral moraines and drift limits in the upper reaches of this valley on south-facing slopes can be traced onto the gentler ground above where they become indistinct. Although they do not provide unambiguous evidence for a plateau icefield on Ullscarf, they do imply that the Greenup and Wythburn ice masses remained contiguous at a late stage in deglaciation, with Ullscarf either being ice-free by this stage or hosting a small, isolated icefield. Streamlined bedrock outcrops on the slopes east of the summit are consistent with a plateau icefield on Ullscarf.

4.2.2.6 Easedale and Far Easedale

Ice moulded bedrock and small flutings in the valley heads, particularly around Codale Tarn, are consistent with a former plateau icefield on High Raise. Nevertheless, an absence of clearly identifiable recessional moraines prevents the reconstruction of palaeo ice-marginal positions in the late stages of deglaciation. Given the steep terrain of these valley heads, it is unlikely that ice-marginal moraines would have formed. Thus, as with Wythburn valley to the north, unequivocal evidence for a former plateau icefield is lacking.

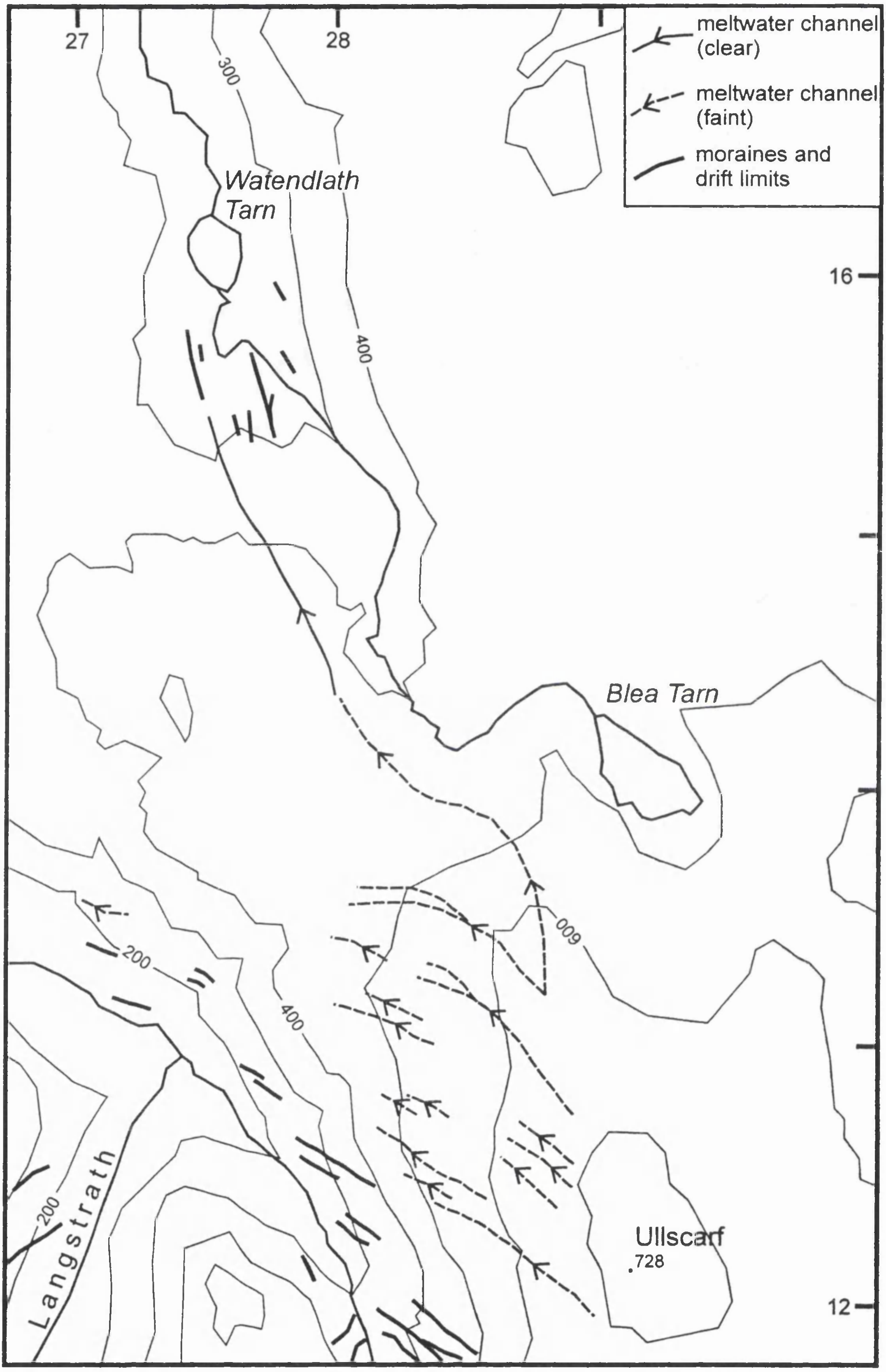


Figure 4.24 Map of Loch Lomond Stadial glaciogenic landform assemblages between Ullscarf and Watendlath
Scale and orientation given by British National Grid References at 1 km intervals

4.3 PALAEOGLACIER RECONSTRUCTIONS

4.3.1 Outlet glacier extents

Whilst unequivocal evidence exists for the development of a plateau icefield on High Raise, Thunacar Knott and Ullscarf during the Loch Lomond Stadial, this primarily takes the form of recessional moraines produced when deglaciation was relatively advanced. Establishing the overall configuration of this icefield system at maximal extent, on the other hand, is fraught with difficulties. With the exception of the locally impressive boundary between the Rossett Pike blockfield and the lateral moraines which define its lower edge in Langstrath, no convincing examples of periglacial trimlines exist in the area. This situation is attributed to the frost-resistant nature of lavas (Thorp, 1981, 1986). Thus, considerable reliance has been placed on depositional evidence, such as lateral moraines and drift limits, with extrapolation necessary in the accumulation zones (Section 4.3.2).

The identification of the maximum downvalley extents of the High Raise outlet glaciers using depositional evidence is complicated by the impacts of paraglacial resedimentation on both the valley sides and valley floors. In considering the reliability of downvalley extents of 'fresh' glaciogenic landform assemblages, it is useful to consider the case of the Oxendale palaeoglacier. Although not part of the plateau icefield system under investigation in this chapter, it produced lateral moraines which are clearly visible from the high ground west of Stickle Tarn (Figure 4.25). These lateral moraines imply a snout at least 500 m more extensive than suggested by Sissons (1980a) on the basis of the downvalley extent of 'fresh', valley-floor moraines. Furthermore, if the debris cone (D in Figure 4.25) comprises reworked glaciogenic debris from the Loch Lomond Readvance, then its altitude suggests a substantially greater volume of ice in Great Langdale than has hitherto been accepted (Section 4.4). It must be stated that, due to different lighting conditions, these lateral moraines are not clearly visible on the aerial photographs used by Sissons (1980a).

The problem of paraglacial resedimentation on valley sides and floors is exacerbated by an absence of suitably located sites for litho- and biostratigraphic investigations. Although moraine morphology ('freshness') has been employed extensively in the past as

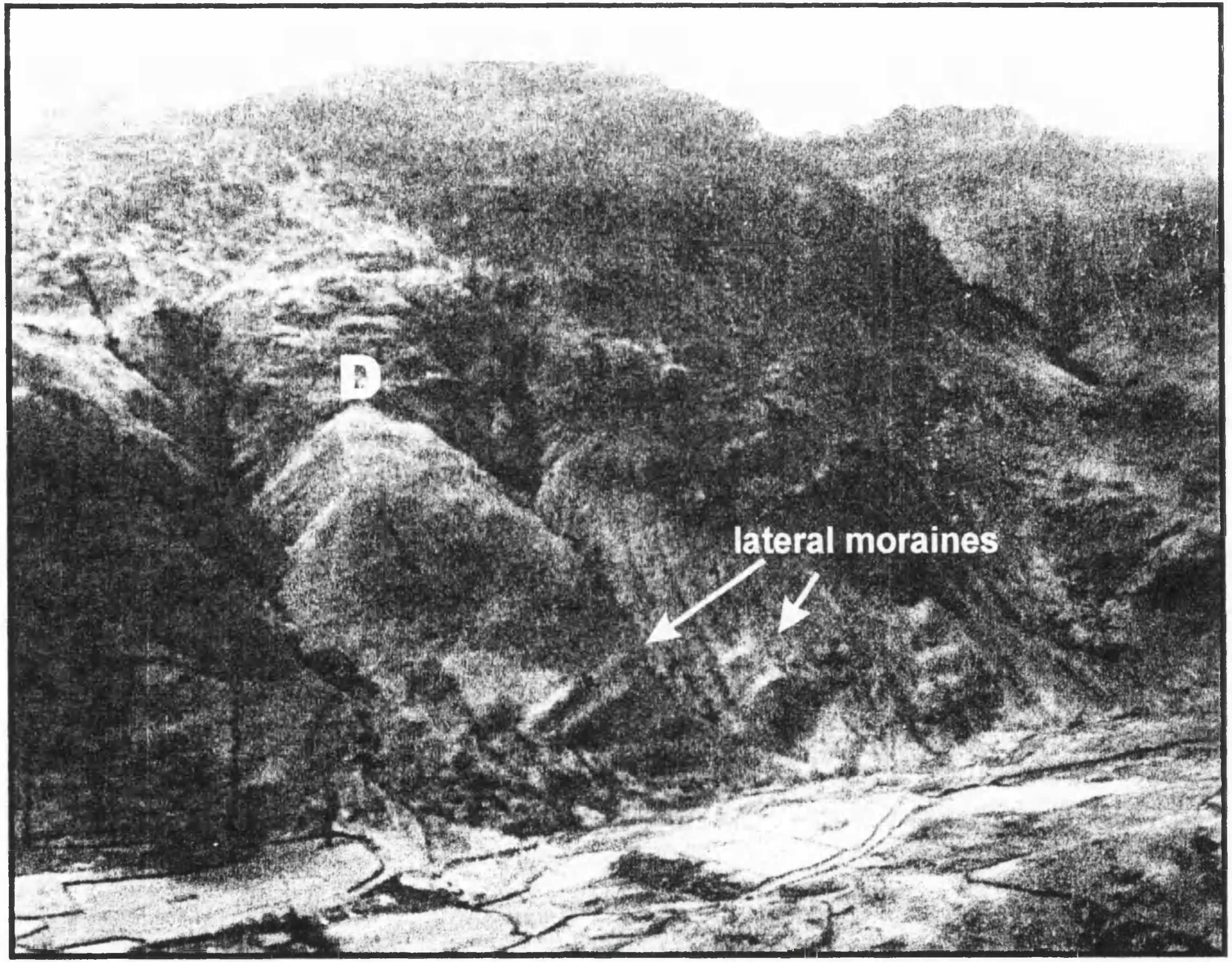


Figure 4.25 View looking towards Oxendale. The lateral moraines visible on the southern slopes imply a maximal glacier position at least 0.5 km more extensive than is suggested by the valley floor moraine limit. See text for explanation.

a relative dating technique, the belief that the downvalley extents of Loch Lomond Stadial palaeoglaciers should be everywhere defined by morphologically similar moraines, regardless of topographic setting or palaeoglacier dynamics, is clearly difficult to sustain.

In this research, Loch Lomond Stadial and Dimlington Stadial glaciogenic landforms are distinguished on the basis of (inferred) styles of deglaciation. It is believed that the Dimlington Stadial ice sheet more or less stagnated *in situ*, at least in the final stages of decay (e.g. Boardman, 1980; Sissons, 1981). In contrast, recent investigations in Scotland have shown that most Loch Lomond Stadial palaeoglaciers actively backwasted from maximal positions over some or all of their lengths (e.g. Benn, 1990; Bennett, 1991; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b). The outlet glaciers of the plateau icefield which developed on High Raise and its contiguous peaks also appear to have experienced active deglaciation (Section 4.2; Figure 4.26). Thus, the areal extents of ice-marginal moraine systems are taken to define the downvalley limits of Loch Lomond Stadial glaciers, regardless of 'freshness'. The limitation of this approach, of course, is that the nature of decay of the last ice sheet has not been comprehensively investigated. Furthermore, the possibility that some of the outermost moraines in these valleys may be associated with a late readvance during ice sheet deglaciation, such as the Wester Ross Readvance (Robinson and Ballantyne, 1979), must be considered (Section 1.4.4).

To the northwest, it is probable that the Stonethwaite palaeoglacier (formed by the confluence of the Greenup and Langstrath palaeoglaciers) extended at least as far as Rosthwaite (NY257147) in the Borrowdale valley (Figure 4.26). Here, evidence for a glacier terminus takes the form of a group of three or four concentric moraine fragments, concave towards the Stonethwaite valley at an altitude of approximately 100 m OD (Wilson, 1977). The occurrence of lateral moraines and drift limits on the valley sides upvalley from the Rosthwaite moraines provides evidence that deglaciation was interrupted by stillstands and/or readvances. Some clear lateral moraines do occur reasonably close to the terminus, with particularly good examples occurring on the slopes immediately south of Stonethwaite at NY263136.

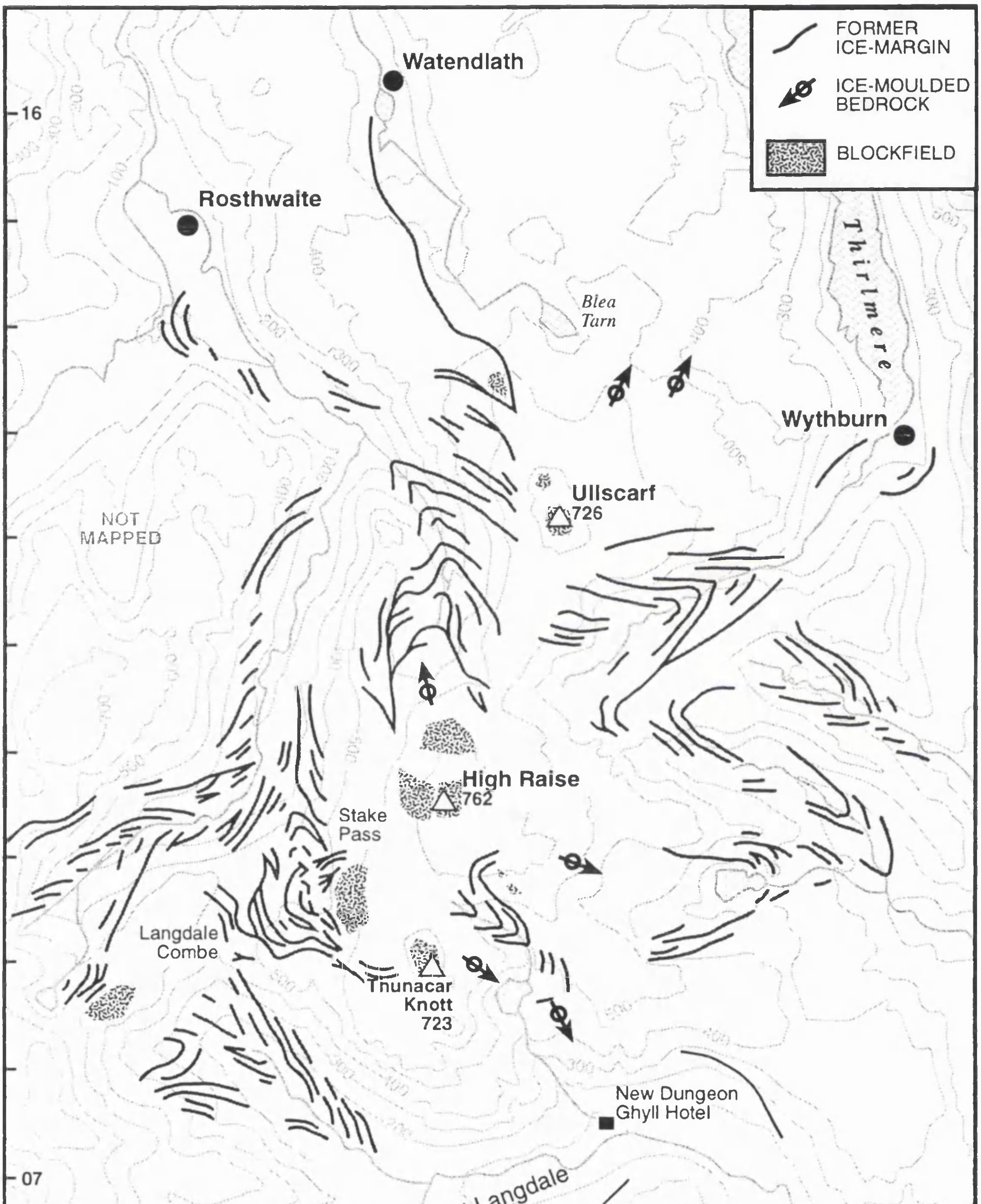


Figure 4.26 Ice-marginal positions associated with the High Raise plateau icefield

To the north, it is suggested that a narrow outlet glacier draining the Ullscarf icefield extended as far as Watendlath Tarn (NY275163), terminating against the rock bar at 350 m OD. This is evidenced by end moraine fragments at Watendlath, the most prominent being situated on the western slopes. An ice-marginal interpretation for the westernmost ridge is confirmed by meltwater channels which continue upvalley to merge with the meltwater channel and ice-marginal moraine system on the western flanks of Ullscarf discussed previously. An apparent absence of ice-marginal moraines upvalley from Watendlath Tarn may reflect uninterrupted retreat of the ice-margin, but it is also the case that a combination of a plateau ice source and an absence of extensive extraglacial slopes above this lobe of ice would have limited supraglacial debris input to these ice margins. In turn, this would have limited the potential for moraine formation. Some moraines occur on the slopes of Ullscarf above Blea Tarn, but these are fragmentary and do not permit palaeo ice-margin reconstructions.

A northwards extension of ice from Ullscarf to Watendlath Tarn is apparently at odds with Pennington's (1978) inference that Blea Tarn remained free of glacier ice during the Loch Lomond Stadial, based on the recognition of Lateglacial Interstadial sediments. It is suggested here, however, that these interstadial sediments may have survived beneath cold-based ice or perhaps escaped erosion due to some other set of favourable subglacial conditions. To this end, it is interesting to note that, in a recent publication, Pennington (1996) states that Blea Tarn is one of a number of 'anomalous' sites where the lithostratigraphy is heavily disturbed, which she attributes to the growth of seasonal lake ice during the Loch Lomond Stadial.

To the northeast, the end moraine at the southern end of Thirlmere (NY324128, ~180 m OD) represents the most likely terminus of the Wythburn outlet glacier. Although incomplete, this end moraine is quite prominent in places, particularly to the west of the A591 where it rises 4–5 m above the surrounding terrain. Following maximal extent, it is inferred that deglaciation was initially characterised by uninterrupted retreat, during which time the ice margin backwasted by approximately one kilometre. This is based on the absence of clear ice-marginal moraines. Thereafter, active deglaciation occurred, with retreat interrupted by stillstands and/or readvances, as evidenced by the moraines and drift limits in the upper Wythburn valley.

It is appreciated that an absence of ice-marginal moraines in the lower Wythburn valley would also result if the debris supply to the ice margin was very low. This is potentially a major problem for palaeoglacier reconstructions in areas, such as the central Lake District, where ice may have originated on the summits and, as such, carried no supraglacial debris from their source areas. This effect is likely to have been most marked at maximal extent, decreasing in importance as deglaciation proceeded and slopes became available for debris supply. Thus, there are at least two potential explanations for the apparent absence of ice-marginal moraines between the northern foot of Dunmail Raise and the upper Wythburn valley. Nevertheless, drift limit altitudes on the upper slopes of the Wythburn valley clearly imply that ice extended beyond the position marked by the prominent ('fresh') moraines on the valley floor in upper Wythburn valley.

Elsewhere, clear evidence for possible maximal extents of these outlet glaciers is lacking; moraines become progressively degraded downvalley. For example, there is no evidence in the Stickle Tarn area which can reasonably be interpreted as representing the maximal extent of Loch Lomond Stadial ice. The absence of such evidence on the steep slopes which descend to the floor of Great Langdale is inconclusive given that moraines are unlikely to have formed there. Nevertheless, abundant drift and ice-moulded bedrock suggest that ice did descend these slopes. Indeed, subdued hummocky depositional landforms on the floor of Great Langdale below Stickle Tarn (around the New Dungeon Ghyll Hotel) may represent the former terminus of this glacier as it debouched onto the valley floor. These fragmentary morainic hummocks only occur north of the river at the foot of the valley sides, a distribution which probably owes much to proglacial and postglacial fluvial erosion. It is suggested here, however, that ice from Stickle Tarn became confluent with ice emanating from Mickleden and Oxendale. A very subtle drift limit on the northern slopes of Great Langdale, if correctly interpreted, suggests that glacier ice extended almost as far as Elterwater, with an equally subtle drift limit on the southern slopes indicating a recessional position (Figure 4.26).

4.3.2 Plateau icefield reconstruction

A tentative reconstruction of the Loch Lomond Stadial plateau icefield system centred on High Raise is shown in Figure 4.27. The general configuration of this reconstructed ice mass is based largely on palaeo-ice-marginal positions produced during deglaciation (Section 4.2). The maximum downvalley extents of the outlet glaciers which drained this plateau icefield are less certain, however, particularly in Great Langdale and on the slopes northeast of Ullscarf (Section 4.3.1).

The general approach to contouring former ice surfaces is discussed in Section 3.4 and need not be repeated here. Nevertheless, it should be noted that considerable uncertainty is introduced by the virtual absence of any ice-marginal control points in the upper reaches of this plateau icefield system. The contours shown in these areas are tentative, drawn with regards to topography and ice-directional indicators. Ice thicknesses on the summits are estimates based on the theoretical relationship between plateau icefield depth and summit breadth shown in Figure 2.2

Firn lines have been calculated for the northern and eastern outlets using Sissons' (1974) AWMA method (Section 3.4) (Table 4.2). The division of a plateau icefield system into sub-systems for the purposes of calculating firn lines is an accepted approach, particularly where these may have varied significantly across the icefield (e.g. Sissons, 1974). In this case, however, the rationale for doing so is that the western and southern outlets merged with ice originating from outwith the system. It is interesting to note that the firn lines calculated for these northern and eastern outlet glaciers are broadly similar (Table 4.2), a consistency which lends support to the general nature of the reconstruction in these areas shown in Figure 4.27. The palaeoclimatic implications will be addressed in Chapter 6.

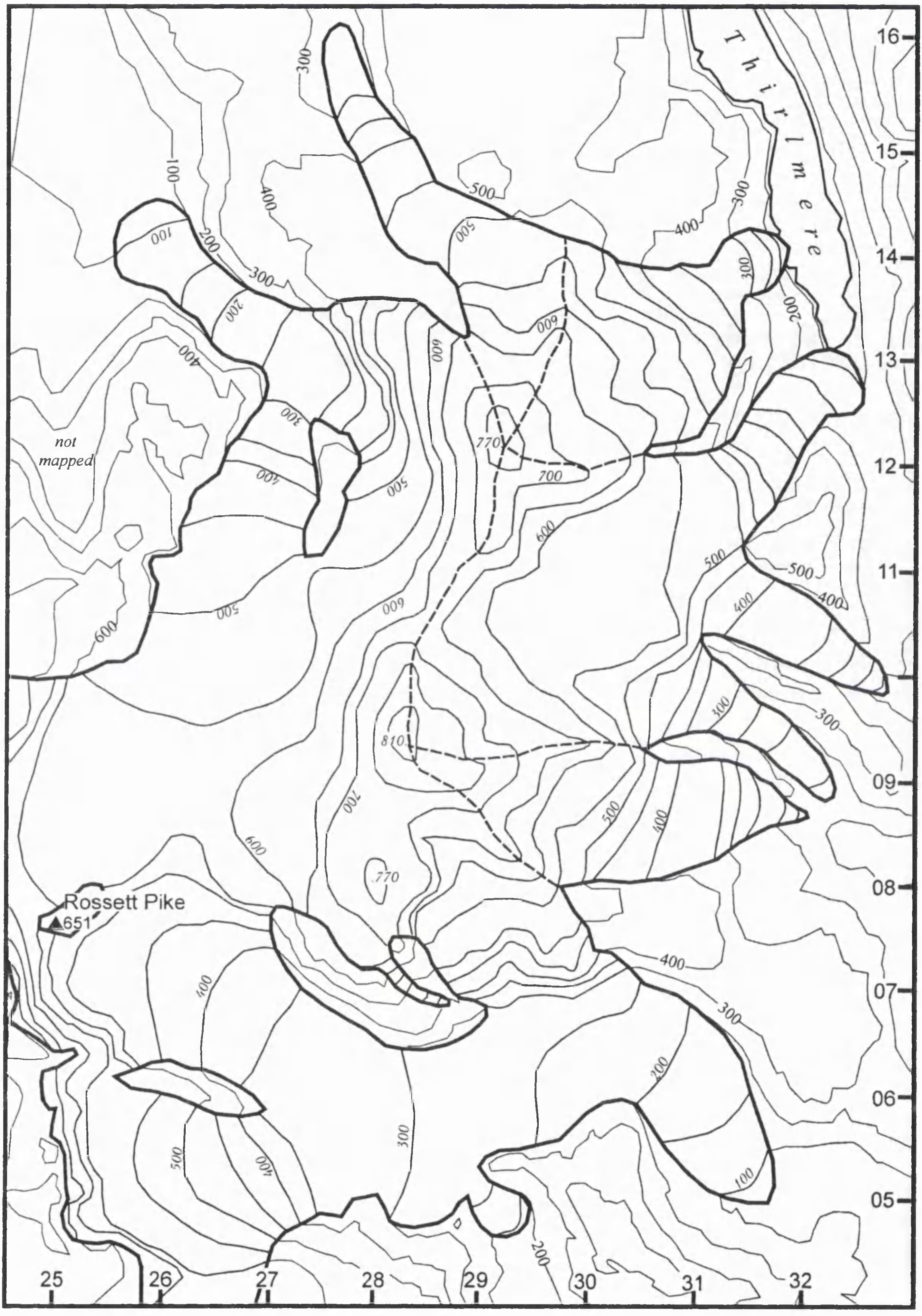


Figure 4.27 Reconstruction of the High Raise plateau icefield system
Scale and orientation given by British National Grid references (NY) at one kilometre intervals

Outlet Glacier	Firn line (m)
Watendlath	507
Wythburn Fells (NE of Ullscarf)	525
Watendlath and Wythburn Fells combined	516
Wythburn, Greenburn and Far Easedale combined	514
Easedale Tarn	498

Table 4.2 Firn line altitudes for the outlet glaciers north and east of the High Raise plateau icefield system

4.4 COMPARISON WITH PREVIOUS WORK

In reconstructing the extents of Loch Lomond Stadial glaciers in the High Raise area, both Manley (1959) and Sissons (1980a) assumed an alpine style of glaciation (see Figure 4.28). Sissons (1980a) viewed the Lake District in general as being very marginal for glaciation at this time, with glacier development largely limited to favourable locations (which would have been both sheltered from the sun and in receipt of considerable quantities of wind-blown snow from extensive plateaux behind). Nevertheless, his maps showed the existence of what he believed to be adversely located palaeoglaciers in the central Lake District. In the High Raise area, Sissons considered the Stickle Tarn palaeoglacier (Glacier 28 in Figure 4.28) to be particularly adversely located on account of its south-facing accumulation area. Also adversely-located, he argued, was the Stake Pass palaeoglacier (the southern margin of Glacier 29) because it developed in a rather shallow open valley. Sissons invoked relatively high levels of snowfall in the central Lake District to account for these and other 'adversely-located' palaeoglaciers.

In the plateau icefield interpretation proposed in this chapter, the problem of Sissons' (1980a) 'adversely-located' palaeoglaciers in the High Raise area is removed, as is the need to invoke relatively high levels of snowfall in comparison with other parts of the Lake District. The existence of ice in the Stickle Tarn area during the Loch Lomond Stadial is to be expected given the evidence for a contemporaneous plateau icefield extending from Thunacar Knott through to Ullscarf. The geomorphological evidence clearly demonstrates that this was drained on all sides by outlet glaciers, including one which occupied the Stickle Tarn area. In the case of the Stake Pass palaeoglacier, the westwards descent of ice from High Raise-Thunacar Knott provides a more convincing explanation for the glacigenic landform assemblages which occur in the Stake Pass and Langdale Combe basins. The valley glacier configuration proposed by Sissons (1980a) is based on the areal extent of 'fresh' moraines, an interpretation identical to that put forward by Manley (1959).

An aspect of Manley's work in this area which deserves attention is his proposal that a small icefield may have developed at this time on High Raise, with ice flowing eastwards to nourish the Easedale Tarn palaeoglacier. Although no geomorphological evidence for

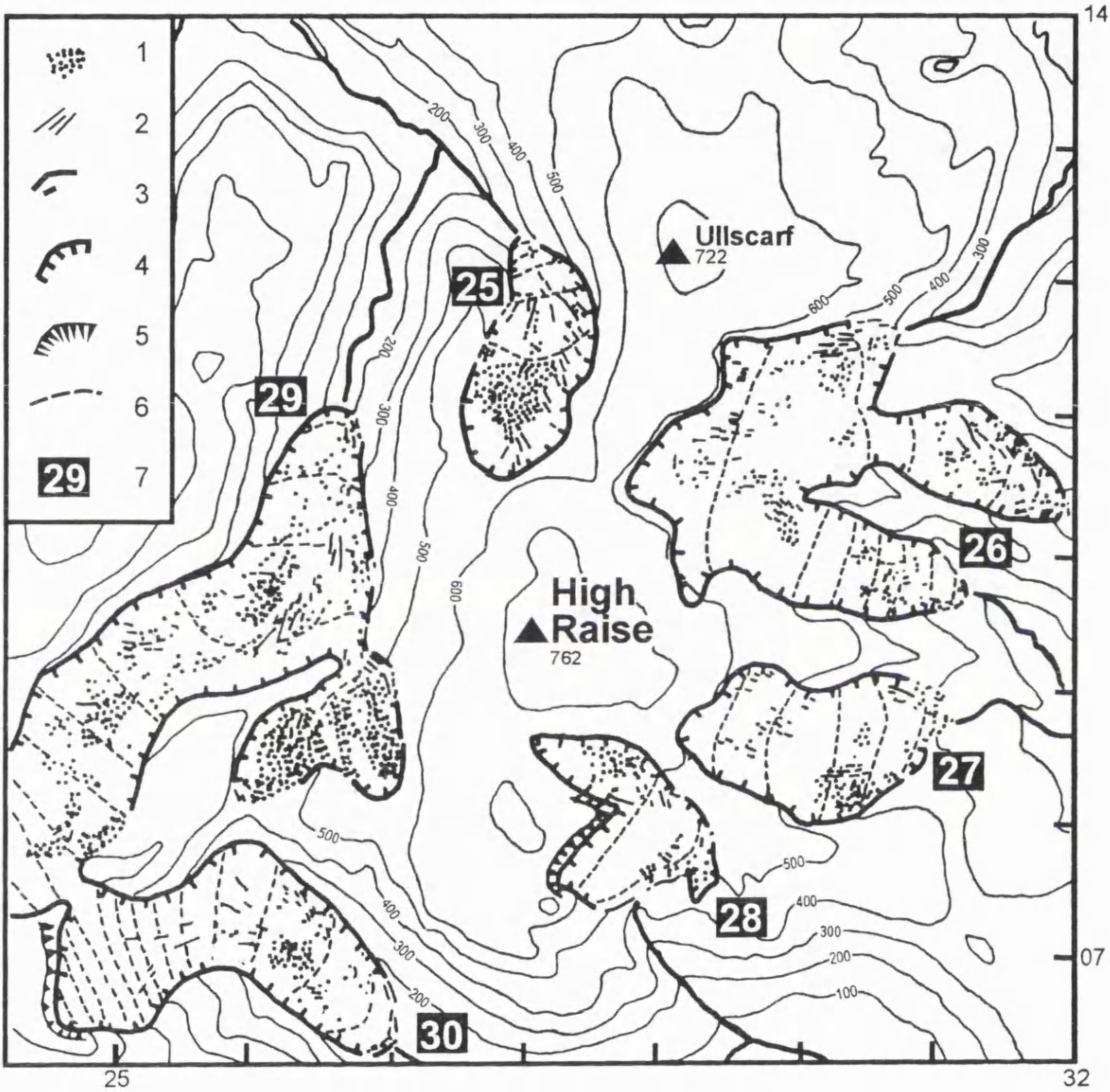


Figure 4.28 Loch Lomond Stadial glaciers in the vicinity of High Raise according to Sissons (1980a).
1. Hummocky moraines
2. Flutings
3. Other linear till ridges, including end moraines
4. Inferred glacier margins
5. Very steep slopes
6. Contours of reconstructed glaciers, 50 m interval
7. Glacier number assigned by Sissons (1980a)

this is provided, he believed the relatively rounded summit of High Raise may have attained a sufficient altitude above the inferred local snowline for the development of a small plateau icefield. His reasoning and figures are partly based on an earlier paper (Manley, 1955) discussed in Chapter 2.

Apart from the existence of a plateau icefield extending from Thunacar Knott to Ullscarf, this research differs from previous reconstructions in terms of maximal downvalley glacier extents, with substantially more extensive positions suggested in some cases here. This difference reflects the heavy reliance placed by previous workers on 'freshness' as a relative dating technique, an approach considered to be unreliable and unnecessarily restrictive by the present author (e.g. Section 4.3.1). Although firn lines have only been calculated for the northern and eastern outlets, their similarity (500–525 m) inspires confidence in the general validity of the reconstruction in these areas. Furthermore, despite being more extensive than the glaciers proposed by previous workers, a plateau ice source means than their firn lines are actually 20–30 m higher than those calculated by Sissons (1980a).

4.5 SUMMARY

Unequivocal evidence has been presented for the development of a plateau icefield on High Raise, Thunacar Knott and Ullscarf during the Loch Lomond Stadial. An impressive sequence of ice-marginal moraines in Stake Pass and Langdale Combe, palynologically dated to the Loch Lomond Readvance by Walker (1965), records in considerable detail the active retreat of a lobate ice margin as it backwasted towards the summit area. The mapping of ice-marginal moraines in Pavey Ark and Greenup reveal a similar pattern.

The geomorphological impact of this icefield appears to have been minimal on the summits, where the survival of blockfield derived from the frost-resistant lavas of the Borrowdale Volcanic Group is interpreted as evidence for at least patchy cold-based conditions during deglaciation. This represents the first recorded example in the literature of a polythermal Loch Lomond Stadial ice mass. It is likely that cold-based conditions were promoted by a combination of thin, slow-moving ice plus the influence of low mean annual air temperatures on the summits. At the plateau edges, where slopes are steeper and ice velocities would have been higher, ice-moulded bedrock imply a transition to wet-based, erosive conditions.

Although the proposed maximal downvalley extents of the plateau icefield outlet glaciers are more extensive than the limits of previous workers, firn lines are approximately 30 m higher because much of the accumulation took place at a higher elevation. The consistency of firn line estimates for the northern and eastern glaciers suggests that the proposed ice margins are contemporaneous, and are broadly in line with those calculated for Loch Lomond Stadial glaciers elsewhere in the central Lake District (Sissons, 1980a).

5

Grey Knotts, Brandreth, & Dale Head

5.1 INTRODUCTION

The summits investigated in this chapter are located between Buttermere and Borrowdale in the west central fells (Figure 5.1). Grey Knotts (NY217126) and Brandreth (NY215119), which attain altitudes of 697 m and 715 m respectively, are contiguous and form a gently undulating summit area which is bounded to the east by Gillercomb, an overdeepened hanging valley which opens out into the Seathwaite valley (Table 5.1). The western slopes of Grey Knotts and Brandreth are, by contrast, gentler and more extensive, terminating above Warnscale at the head of the Buttermere glacial trough. To the south, the slopes of Brandreth are truncated by the Ennerdale trough and Green Gable (NY215107). The northern slopes of Grey Knotts descend steeply into Honister Pass and Little Gatesgarthdale, north of which lies Dale Head (NY223153), a summit which attains an altitude of 753 m.

Evidence is presented in this chapter for the development of plateau icefields on Grey Knotts/Brandreth and Dale Head during the Loch Lomond Stadial. Reconstructed palaeo ice-marginal positions indicate that the glaciogenic landform assemblages east of the Honister Pass watershed were produced by plateau icefield outlet glaciers which debouched from these summits. Similarly, reconstructed ice-marginal positions on the northern slopes of Ennerdale provide additional evidence for a contemporaneous Grey Knotts/Brandreth plateau icefield. Although ice-moulded bedrock occurs at several localities on the western slopes of Grey Knotts/Brandreth, the extent of blockfield and other frost-weathered debris implies minimal glacial erosion on summit areas.

The remainder of this chapter consists of three sections. The evidence for plateau icefields in the area is outlined in Section 5.2, where the emphasis is on reconstructing palaeo-ice-margins during deglaciation (Chapter 3). A reconstruction of these former plateau icefield systems at maximal extents is presented in Section 5.3. The interpretation of the geomorphological record proposed here is then compared with those of previous researchers (Section 5.4). The chapter concludes with a summary in Section 5.5.

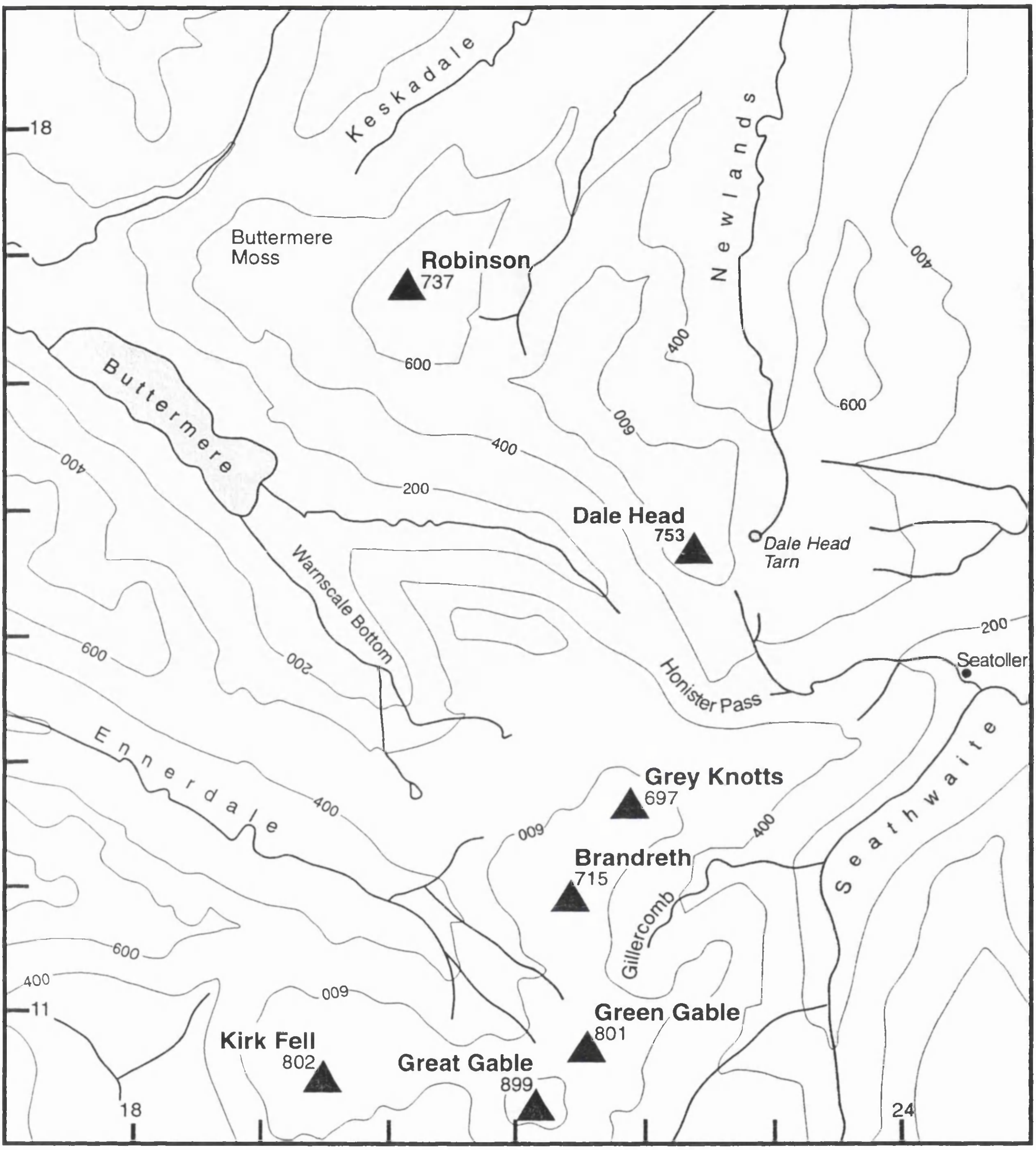


Figure 5.1 Location and topography of Grey Knotts, Brandreth and Dale Head. Scale and orientation given by National Grid References at 1km intervals.

Summit	Grid Reference	Altitude	Breadth
Dale Head	NY223153	753	450
Grey Knotts	NY217126	697	475
Brandreth	NY215119	715	425
Green Gable	NY215107	801	175
Great Gable	NY211103	899	200
Kirk Fell	NY195105	802	325

Table 5.1 Altitudes and breadths of summits mentioned in the text

5.2 EVIDENCE FOR PLATEAU ICEFIELDS ON DALE HEAD, GREY KNOTTS AND BRANDRETH

5.2.1 Summit geomorphology

Frost weathered mountain-top detritus derived from the Lower Borrowdale Volcanic Group mantles the upper slopes and summits of Dale Head, Grey Knotts and Brandreth. This generally takes the form of clasts embedded in a matrix of sand and silt, with a vegetation cover that is in some places almost complete (Figures 5.2 and 5.3). On Grey Knotts and Brandreth, however, this locally grades into patches of true blockfield, with a complete cover of boulders and no visible fines. Two pits excavated into this frost-weathered material in the vicinity of Brandreth summit indicate a minimum thickness of 0.3 m. The adjacent summits of Green Gable, Great Gable and Kirk Fell also possess mantles of frost weathered debris (Figure 5.4).

The distribution of frost-weathered debris in upland Britain and the ground covered by Loch Lomond Stadial glaciers is conventionally regarded as being mutually exclusive (Section 2.4.2). Nevertheless, it has been shown that the plateau icefield which developed on High Raise and its contiguous peaks during the Loch Lomond Stadial was polythermal, and that the frost weathered debris on the summit areas predates this glaciation (Section 4.2.1). A polythermal regime is also proposed here for the plateau icefield which developed on Grey Knotts/Brandreth. Evidence for this plateau icefield can be found east of Grey Knotts on the spur of high ground which separates Honister Pass from Gillercomb. Fragments of low linear till ridges, interpreted as ice-marginal moraines, extend from reconstructed ice-marginal positions in Little Gatesgarthdale (Section 5.2.2) across this high ground, descending into the Gillercomb basin to the south (Figure 5.5). This implies a common source for the Honister Pass and Gillercomb palaeoglaciers, and thus the existence of a contemporaneous plateau icefield on Grey Knotts/Brandreth.

Although the extent of frost weathered mountain-top detritus suggests that glacial erosion was minimal or absent on and around the summit areas, some bedrock outcrops exhibit clear evidence for ice-moulding. The whalebacks which occur on the western

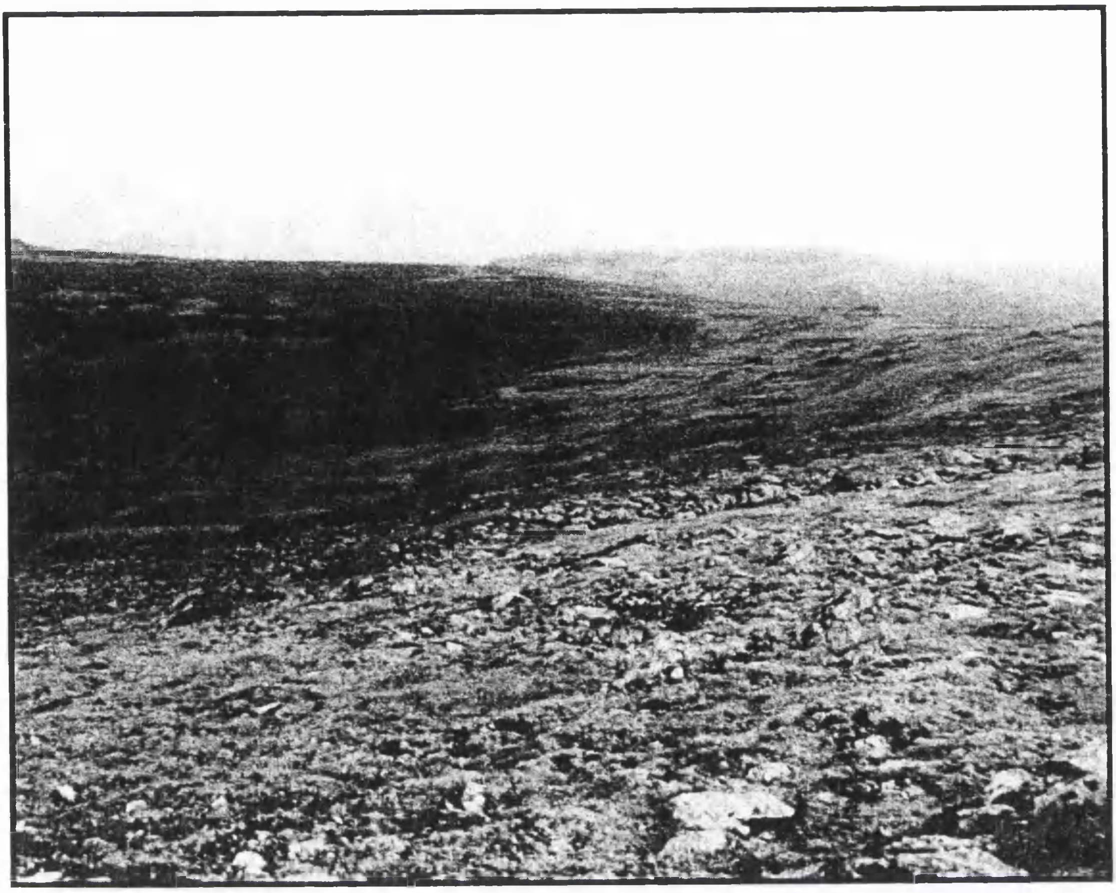


Figure 5.2 Frost weathered debris on Dale Head, east of summit

Figure 5.3 Frost weathered debris on Brandreth. View looking north towards Grey Knotts.

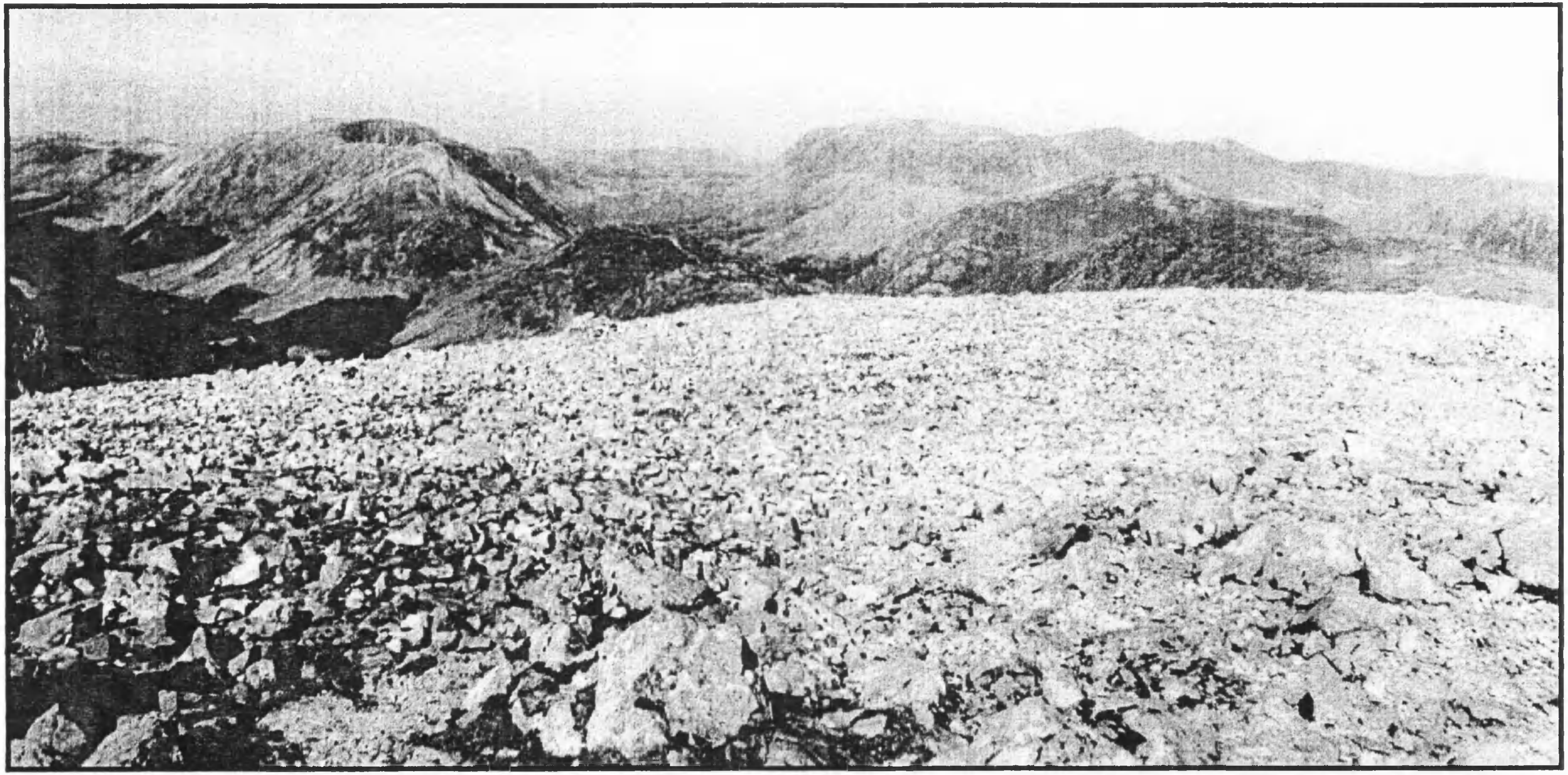


Figure 5.4 Blockfield on Great Gable. View looking west towards Ennerdale and Buttermere.

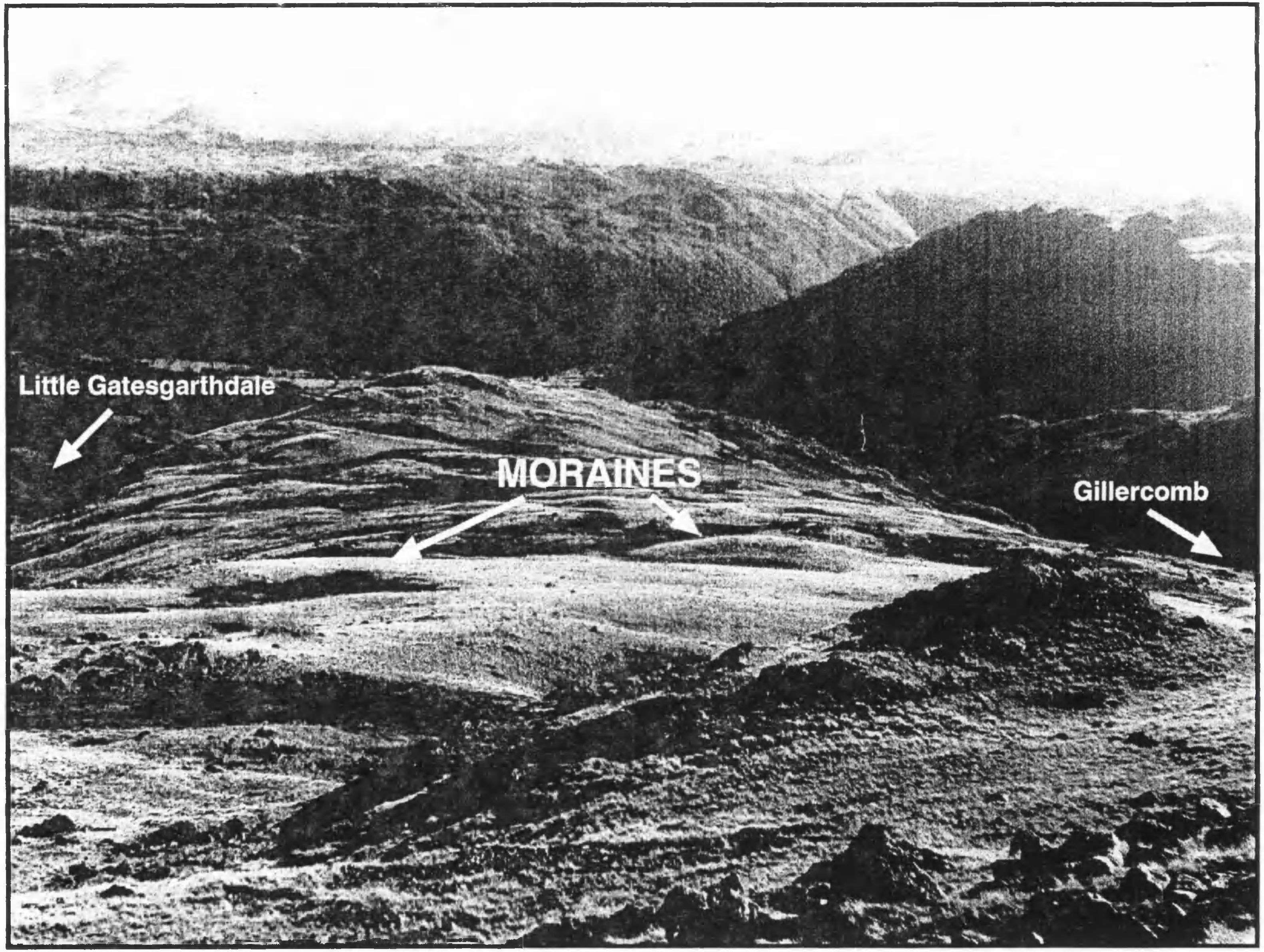


Figure 5.5 Low till ridges interpreted as ice-marginal moraines on high ground separating Little Gatesgarthdale and Gillercomb.

slopes of Brandreth are particularly good examples (Figure 5.6). It is thought the absence of quarried lee faces on whalebacks implies high effective normal pressures and the absence of low-pressure basal cavities during their formation (Bennett and Glasser, 1996). Whalebacks are apparently best developed where ice was 1–2 km thick, but can form below ice a few hundred metres thick (I.S. Evans, 1996). Such thickness' exceed those which could reasonably be associated with a small Loch Lomond Stadial plateau icefield (Section 5.3). It has also been proposed that whalebacks may evolve from roches moutonnées where ice-flow directions change and quarried faces are removed by abrasion (Veillette, 1986; Anundsen, 1990). In such circumstances, whalebacks could develop beneath relatively thin ice typical of a small plateau icefield. In this particular case, however, it is likely that ice-flow directions for both the Late Devensian ice sheet and the proposed Loch Lomond Stadial plateau icefield would have been more or less coincident. Moreover, it is difficult to envisage ice-flow directions within the Loch Lomond Stadial plateau icefield changing significantly during deglaciation.

There is no clear geomorphological evidence in the vicinity of Dale Head summit for the development of a Loch Lomond Stadial plateau icefield, although evidence can be found in Honister Pass to the south (Section 5.2.2). Precipitous slopes define the northern and western margins of Dale Head, but gentler slopes extend to the south and southeast. This asymmetry resulted in the plateau icefield being most extensive to the south. Gullying and thicker drift on the slopes above Honister Pass are interpreted as having been produced in association with a plateau icefield outlet glacier (Figure 5.7). However, the nature and significance of this evidence must be evaluated in the context of reconstructed ice-marginal positions on the valley floor (Section 5.2.2).

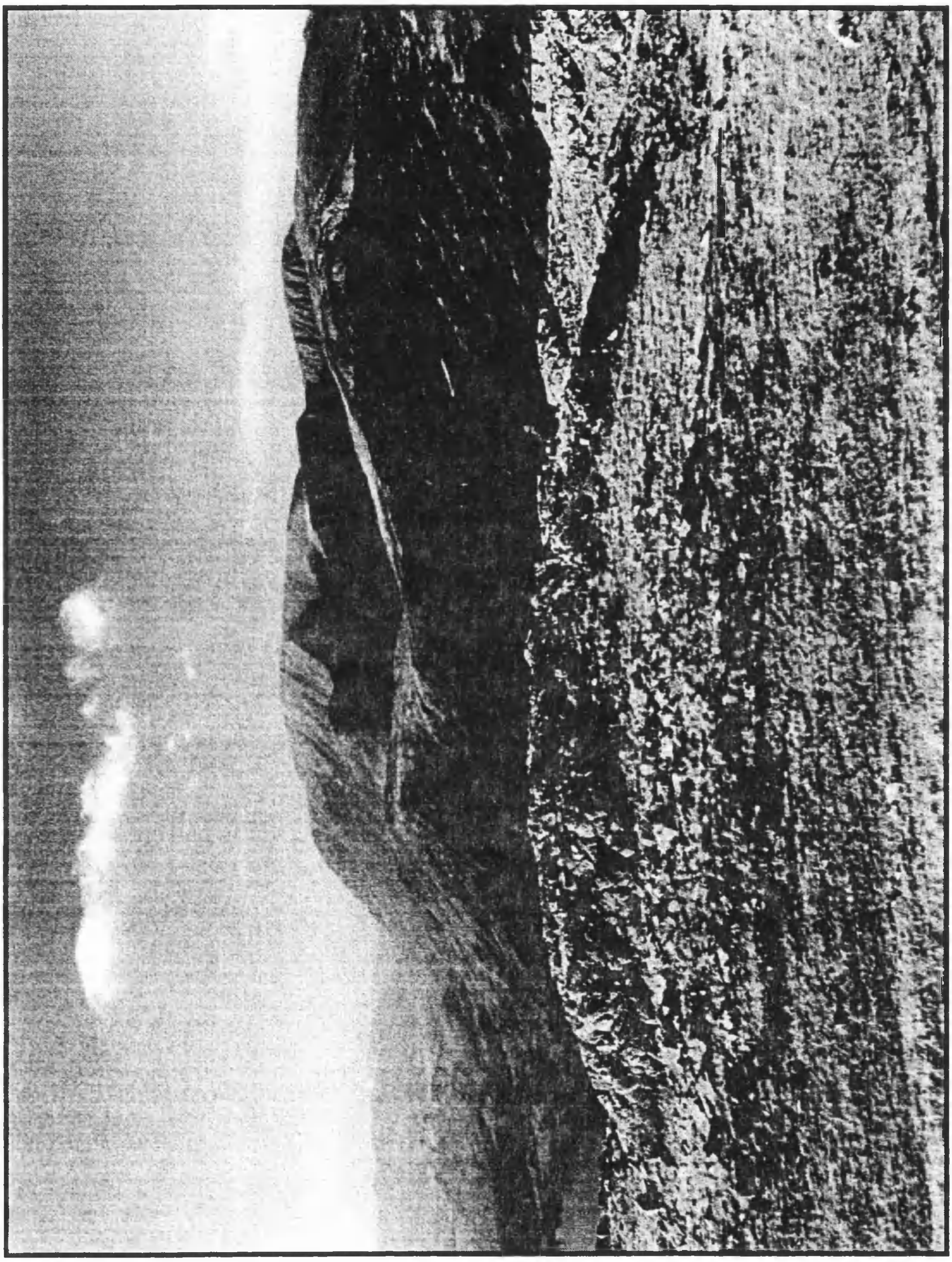


Figure 5.6 Whalebacks on western slopes of Brandreth

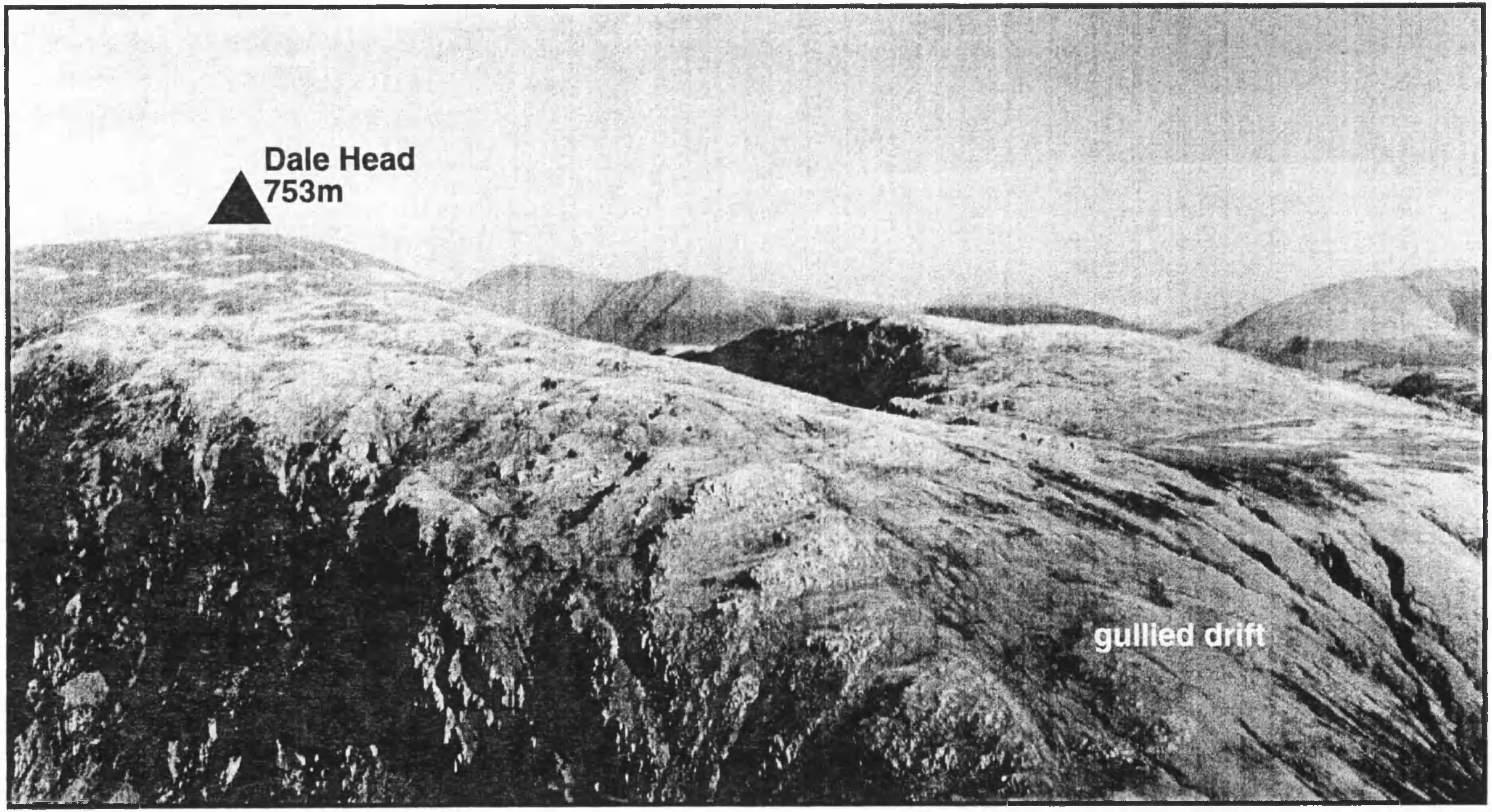


Figure 5.7 Dale Head from Grey Knotts. Note the gullied drift on the southern slopes.

5.2.2 Honister Pass

At an altitude of 360 m OD, the Honister Pass watershed separates the Gatesgarthdale and Little Gatesgarthdale valleys. This is the route taken by the B5289, which links Buttermere with Borrowdale. West of the watershed is Gatesgarthdale, which is steep and relatively narrow in its upper reaches where it is bounded by the precipitous crags of Fleetwith Pike and Dale Head (Figure 5.8). The crags are developed within the resistant rocks of the Lower Borrowdale Volcanic Group, whereas the lower slopes, which are strewn with rockfall and quarry waste, belong to the less resistant Buttermere Formation of the Skiddaw Group. Little Gatesgarthdale, by contrast, is wider and has a gentler relief (Figure 5.9). Throughout much of its length the valley floor gradient is relatively gentle, the notable exception being the final steep descent to Seatoller in Borrowdale.

Although virtually absent west of the watershed, glaciogenic landform assemblages mantle the slopes and valley floor of Little Gatesgarthdale. These consist predominantly of ridges and hummocks arranged in closely-spaced chains orientated obliquely downvalley (Figure 5.10). Some ridges have a beaded long profile, which accentuates the hummocky appearance of these landform assemblages. In general, the most prominent moraines occur in the middle and upper reaches of the valley, where some rise 4–5 m above the surrounding terrain. Inter-moraine peat development has, however, served to subdue local relief in many places. A transition from relatively prominent moraines to more rounded forms occurs in association with a marked change in gradient as the valley falls towards Seatoller.

In the past, variations in moraine morphology in upland Britain have been attributed almost solely to weathering contrasts, with 'freshness' used to differentiate between Loch Lomond Stadial and older glaciogenic landforms (Section 2.4.2). As argued in Section 2.4.2, such an approach fails to address other controls on moraine morphology, including glaciation style, ice-margin behaviour during deglaciation, and paraglacial resedimentation processes (cf. Evans, 1988, 1990; Ballantyne and Benn, 1994). In this case, it is likely that, on these steeper slopes, paraglacial resedimentation processes would have been promoted, thus accounting for the more subdued appearance of these moraines. Without sections, however, the efficacy of paraglacial processes following the

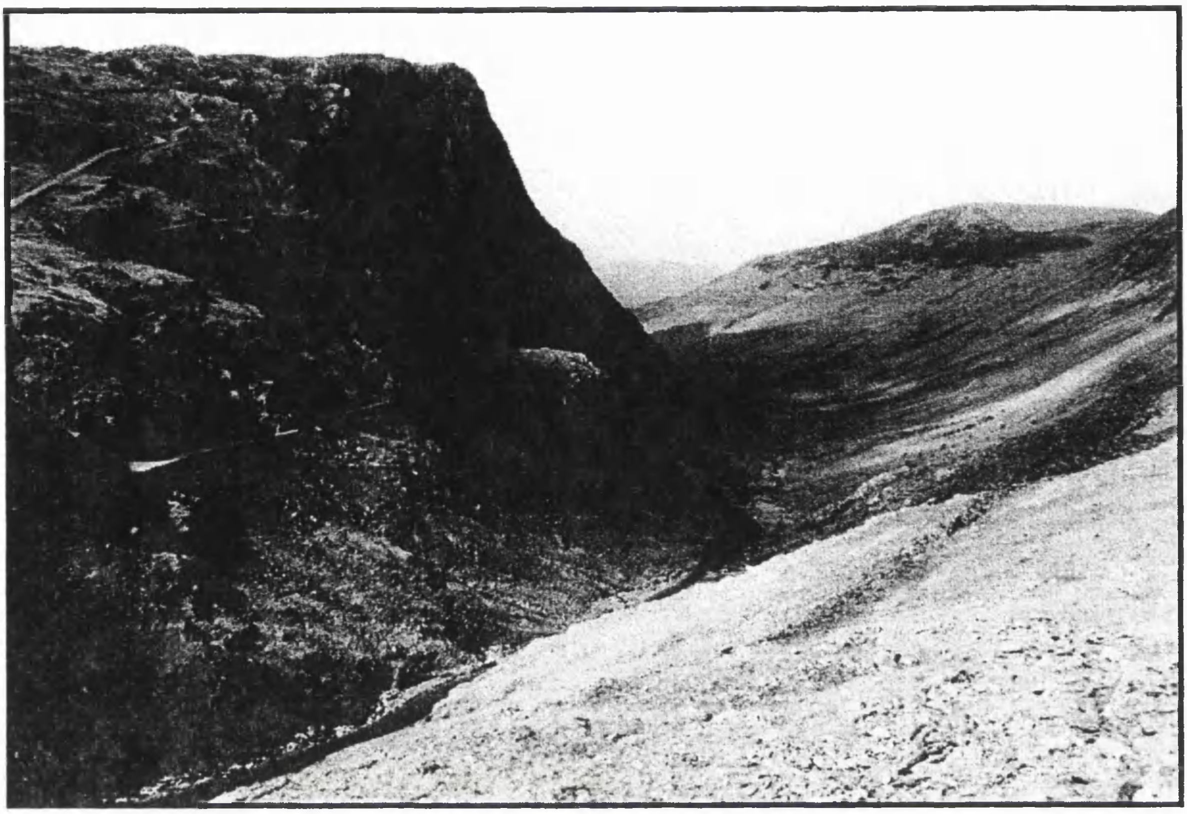


Figure 5.8 Upper Gatesgarthdale, looking west. The crags to the south of the road are those of Fleetwith Pike.

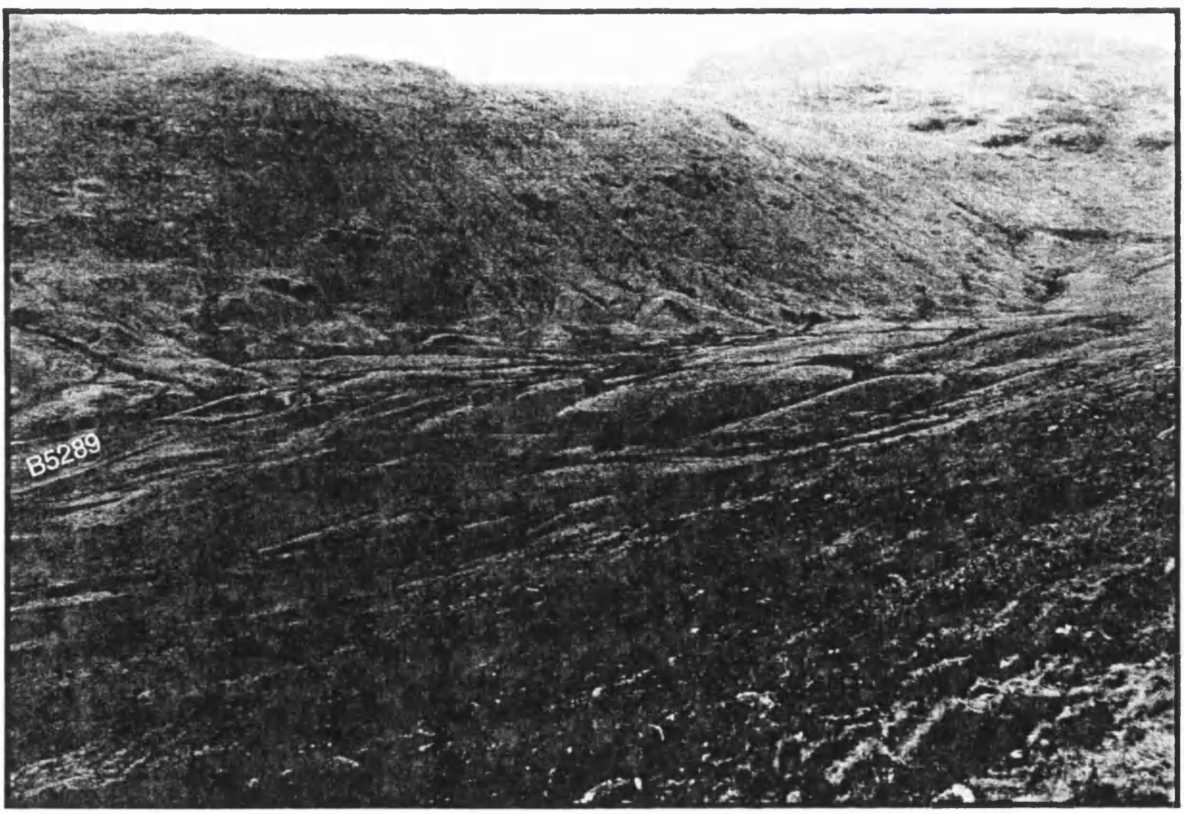


Figure 5.9 Little Gatesgarthdale looking towards Seatoller Fell and Grey Knotts. Glaciogenic landforms are well-developed on valley floor and lower valley sides.

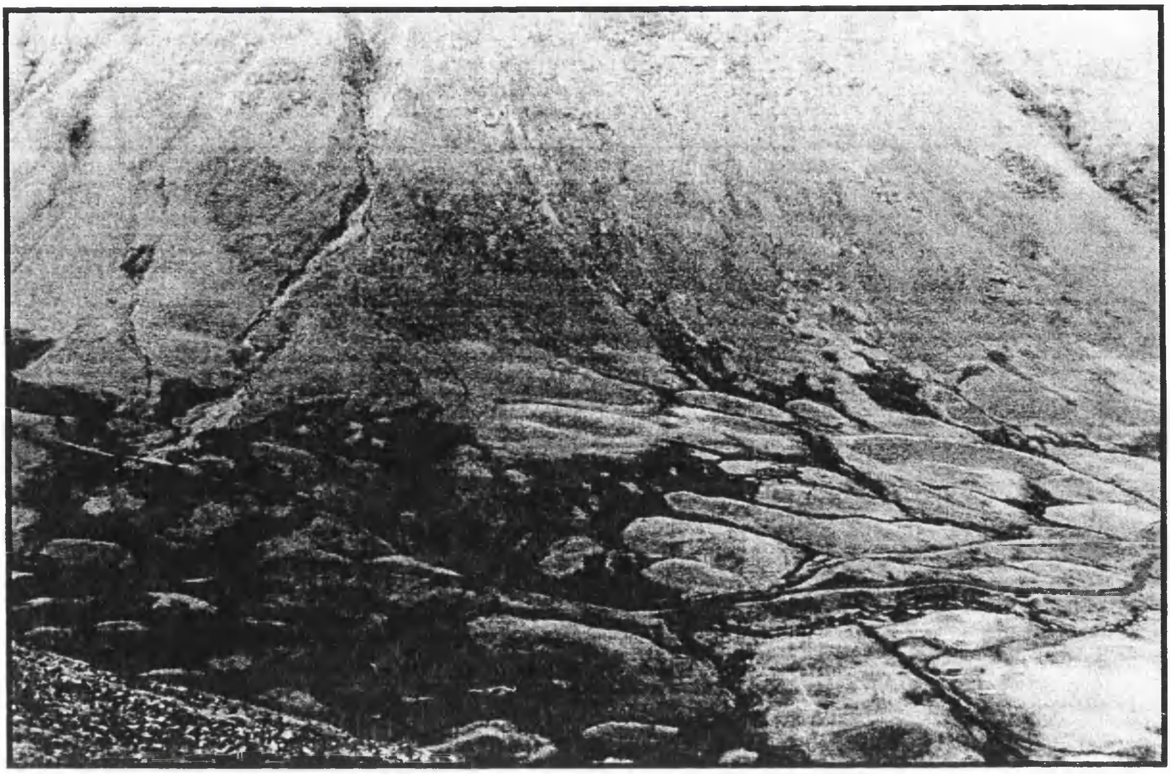


Figure 5.10 Loch Lomond Stadial moraines on the floor of Little Gatesgarthdale. View looking north from Seatoller Fell

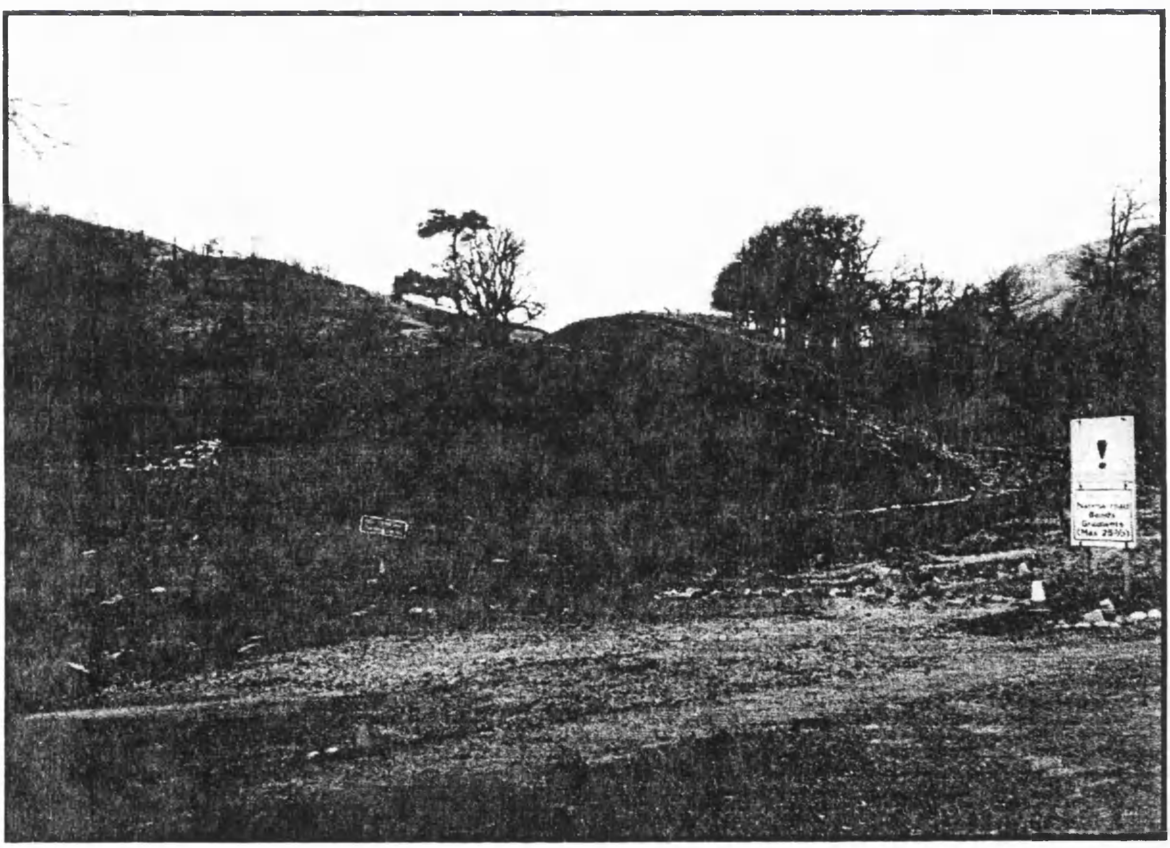


Figure 5.11 Latero-frontal moraine at Seatoller, Borrowdale. This may represent the maximal downvalley extent of the Honister Pass palaeoglacier.

withdrawal of the ice margin at this locality cannot be established. Nevertheless, to argue that variations in moraine development in Little Gatesgarthdale reflect a transition to older moraines requires evidence for a former glacier terminus, of which there is none. Indeed, reconstructed ice-marginal positions can be shown to extend across the boundary from prominent to subdued moraines (see below). Furthermore, lithostratigraphic analyses of inter-moraine peat cores taken throughout Little Gatesgarthdale have revealed only Flandrian sediments, with the characteristic tripartite Lateglacial sequence absent (see Figure 5.12 for coring sites). It is presumed, therefore, that all the glacial depositional landform assemblages, whether sharply defined or more rounded, relate to the Loch Lomond Stadial glaciation and that the palaeoglacier which produced them extended at least as far as Seatoller (NY245137) (Figure 5.11).

The glaciogenic landform assemblages in Little Gatesgarthdale are interpreted as having been produced in association with an actively-backwasting ice margin. Geomorphological mapping reveals that these moraines possess a large-scale organisation similar to that of ice-marginal moraines produced by many present-day actively-retreating glaciers in upland areas (Figure 5.12). The frontal margins of mountain glaciers tend to have a strongly-developed lobate planform which, where conditions permit, is often mirrored by the distribution of ice-marginal landforms (e.g. Bennett and Boulton, 1993a, b). It is clear that, in planform, both the moraines and meltwater channels in Honister Pass have a similar arcuate or lobate arrangement, particularly in the downvalley reaches (e.g. M1-M4 on Figure 5.12). This is complemented by clear evidence for moraine bifurcations at several localities (e.g. B1, B2 on Figure 5.12). These are characteristic of ice-marginal moraines in contemporary glacial environments, resulting from differential retreat of the ice front during a period of moraine formation. Although the geomorphological record clearly indicates that the glaciogenic landform assemblages in Little Gatesgarthdale were produced by an actively-backwasting ice-margin, the absence of sections prevents a detailed assessment of moraine genesis.

The ice-marginal moraines which occur on the southern slopes of Little Gatesgarthdale trend obliquely down the valley sides at relatively steep angles, although their

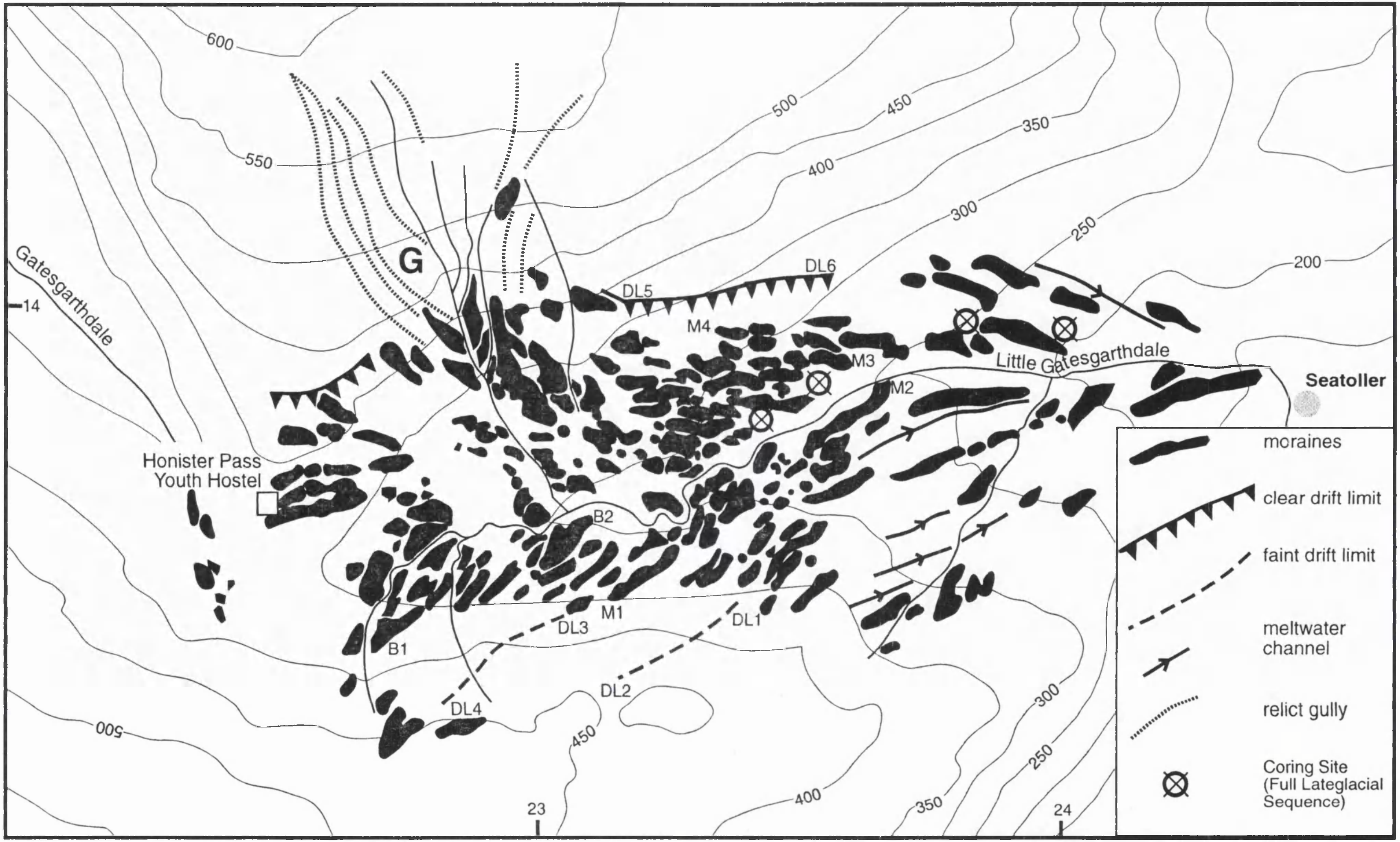


Figure 5.12 Map of Loch Lomond Stadial moraines in Little Gatesgarthdale.
Scale and orientation given by British National Grid References (NY) at 1km intervals.

interpretation is complicated by the effects of paraglacial resedimentation. Many moraines have been dissected by fluvial incision and locally debris cones provide evidence for mass wasting processes. Nevertheless, former ice-marginal positions can, with care, be reconstructed from aerial photograph interpretation. Some of the steeply sloping moraines on Seatoller Fell are continued, in their upper reaches, by faint drift limits which extend beyond the confines of the valley and onto the spur of high ground that separates Little Gatesgarthdale from Gillercomb. Examples are marked DL1-DL2 and DL3-DL4 on Figure 5.12. These are continued by the low ridges shown in Figure 5.5. Such ice-marginal configurations imply that the Honister Pass palaeoglacier was contiguous in its upper reaches with the ice mass inferred to have occupied Gillercomb at that time (Section 5.2.5), which in turn can only be interpreted as evidence for a contemporaneous plateau icefield on Grey Knotts/Brandreth.

On the northern slopes, most ice-marginal moraines trend obliquely down the valley sides at gentler gradients than those on Seatoller Fell, although gradients increase in the vicinity of the gullied, slightly concave section of hillside identified as 'G' on Figure 5.12. This section of the hillside contrasts with slopes to the east on account of its gullied drift cover, on which low mounds and ridges occur in some places, which extend beyond the confines of the valley and onto the gentler slopes of Dale Head above. The gully network, particularly the relict features on the upper slopes, possess an intriguing regularity and symmetry about the main gully. This gullied drift cover is interpreted as evidence for a plateau icefield outlet glacier which descended into Little Gatesgarthdale, with the relict gullies defining former ice-marginal positions during deglaciation. Such an interpretation is supported by a drift limit (DL5-DL6) on the northern slopes of Honister Pass (Figure 5.12), which is consistent in terms of orientation with an outlet glacier entering Little Gatesgarthdale at 'G.' Initially steep, the gradient of this drift limit lessens over a short horizontal distance to become consistent with the system of ice-marginal moraines and meltwater channels in the middle and lower reaches of the valley.

Immediately east of the watershed, on the northern slopes above the Honister Pass Youth Hostel, is a sequence of moraines on the lower valley sides with a distinct upper drift limit (Figure 5.13). These are interpreted as successive ice-marginal positions

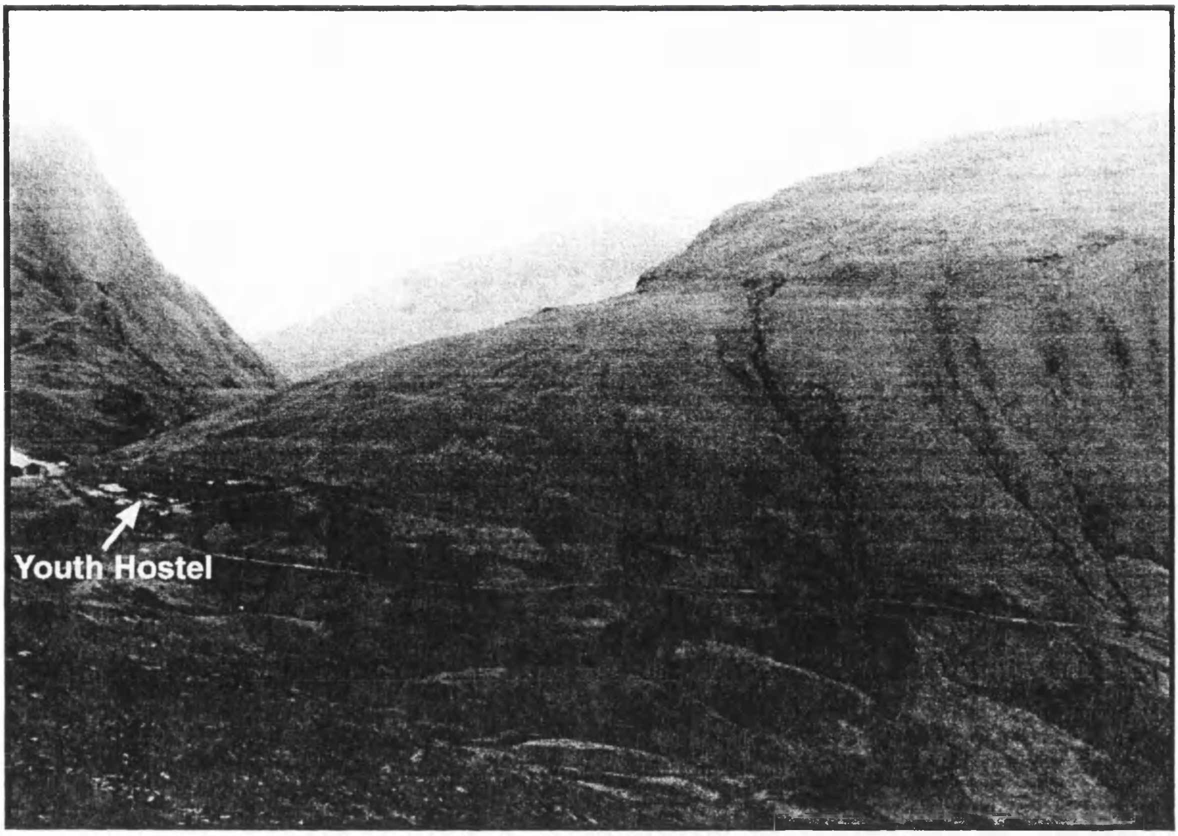


Figure 5.13 Ice-marginal moraines north of Youth Hostel
Former ice-marginal positions indicated by subtle drift benches.

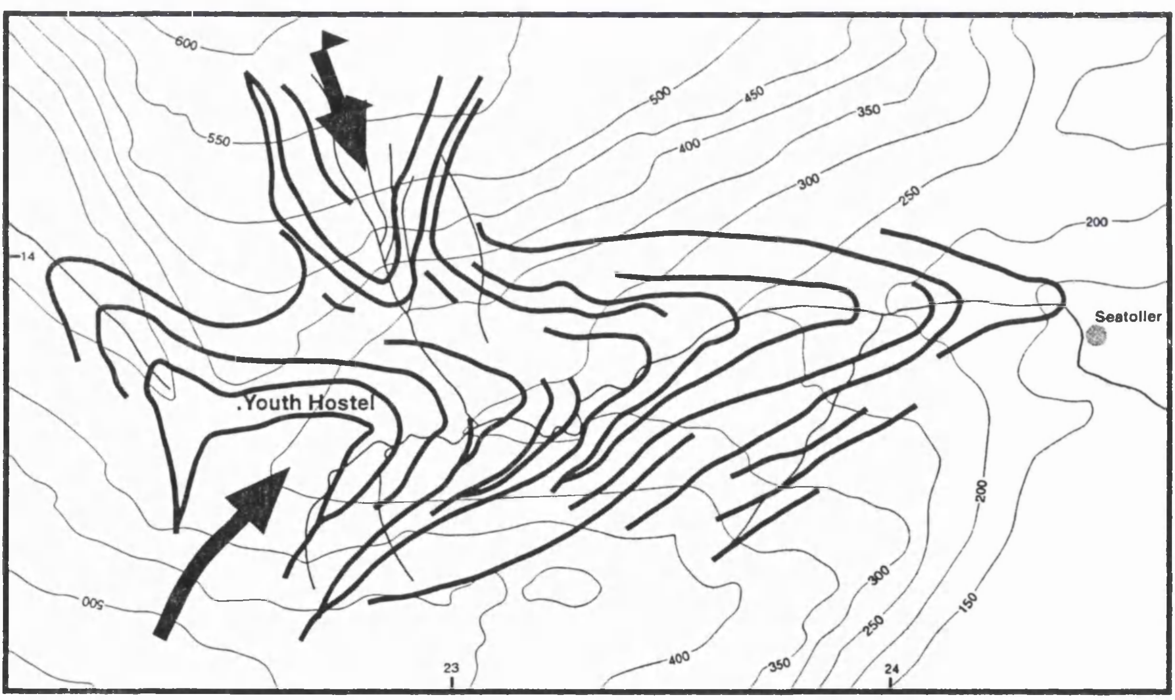


Figure 5.14 Reconstructed ice-marginal positions in the Honister Pass area.
These indicate decay centres on Dale Head and Grey Knotts/Brandreth

associated with ice descending from Grey Knotts on the opposite side of the valley. Their altitudes clearly demonstrate that ice flowed west of the watershed into Gatesgarthdale, although direct evidence for this is slight, being restricted to lateral moraines south of the road in the vicinity of the watershed itself. These would have been produced in the late stages of deglaciation, with evidence for more extensive positions (implied by the altitudes of the upper moraines on the slopes north of the Youth Hostel) either never produced or subsequently having been destroyed.

From the geomorphological evidence discussed above, a map showing approximate palaeoglacier configurations during deglaciation has been constructed (Figure 5.14). This clearly shows that the Honister Pass palaeoglacier was nourished by plateau icefield outlet glaciers descending from Grey Knotts/Brandreth and Dale Head, the former being a more significant source than the latter.

5.2.3 Newlands Valley

The Newlands valley lies north of Dale Head and is separated from it by precipitous cliffs (Figure 5.1). Although the geomorphological record in the Honister Pass area indicates that a plateau icefield developed on Dale Head during the Loch Lomond Stadial, clear and unequivocal evidence for contemporaneous glaciation is lacking from the head of Newlands valley. Nevertheless, subdued depositional ridges occur at the base of Dale Head crags, immediately north of the summit (Figure 5.15). Although there is no dating control, these may represent evidence for a small glacier nourished by snow and ice avalanching from the plateau icefield above. In addition, faint lateral moraines occur east of Dale Head Tarn, on west-facing slopes above Newlands Beck. These are interpreted as defining former ice margins of the Dale Head plateau icefield (Figure 5.16).

The absence of clear evidence for Loch Lomond Stadial glaciation in Newlands valley may reflect topographic controls on the development of the Dale Head plateau icefield. Although the northern and western margins of the summit are defined by steep crags which would have limited plateau icefield development, gentler, more extensive slopes to the south and southeast would have promoted icefield development in these directions. Thus, outlet glaciers descending into Newlands would probably have been very restricted

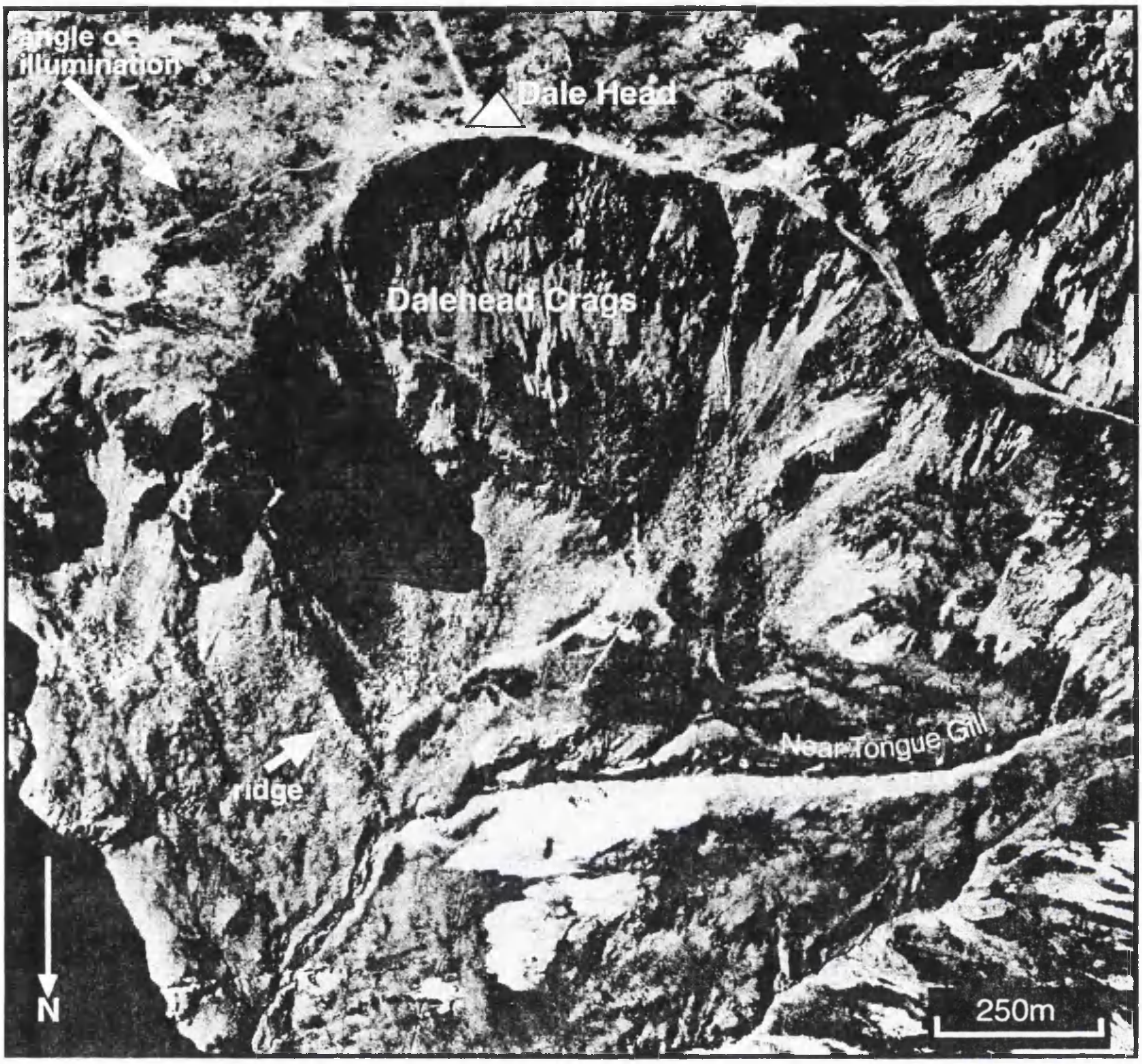


Figure 5.15 Subdued ridge interpreted as ice-marginal moraine at the base of Dalehead Crags. It is possible that this small palaeoglacier was nourished by snow and ice avalanching from the icefield which developed on Dale Head during the Loch Lomond Stadial.

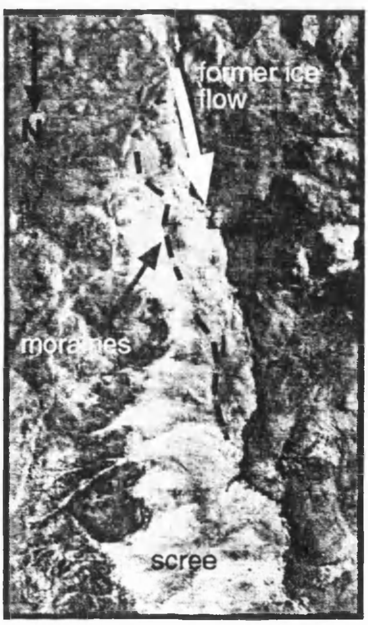


Figure 5.16 Extract from aerial photograph showing subdued mounds and ridges at the head of Newlands valley. These may be ice-marginal moraines associated with a Loch Lomond Stadial plateau icefield outlet glacier.

in extent. This, in combination with an absence of supraglacial debris carried from the source area, means that the potential for moraine production would have been very limited, and this may explain the somewhat equivocal nature of the geomorphological record in this area.

5.2.4 Ennerdale

The Ennerdale valley, which extends for approximately 12 km before opening out into the west Cumbrian coastal lowlands, was clearly a significant routeway for ice draining from the Lake District massif during times of regional glaciation (Figure 5.17). The impact of regional glaciation is particularly evident on the northern slopes of Pillar, where precipitous, arcuate cliffs appear to be backwall remnants of former corries which have subsequently been truncated by ice flowing west towards the Irish Sea.

Prominent glaciogenic landform assemblages at the head of Ennerdale represent evidence for the Loch Lomond Stadial glaciation (Figure 5.18). The lithostratigraphies of inter-moraine peat cores are consistent with this interpretation, the classic Lateglacial tripartite sequence being absent at all sites investigated (cf. Gray and Coxon, 1991). There is, however, a marked contrast in moraine development either side of the River Liza, with the most impressive glaciogenic landforms occurring to the north (Figure 5.19)

North of the river, ridges and hummocks are arranged in closely-spaced chains which trend obliquely down the valley sides at consistently steep gradients. On the southern slopes of Tongue (the spur of higher ground separating Tongue Beck and the River Liza), these predominantly take the form of very clearly defined ridges, but a transition to aligned hummocks and hummocky ridges occurs west of Loft Beck, a short distance downvalley (Figure 5.20, Figure 5.21). Clear evidence for paraglacial activity on the steeper slopes exists in the form of debris cones, particularly on the slopes downvalley from the Black Sail Youth Hostel (Figure 5.22). These debris cones consist of material which has been reworked from Loch Lomond Stadial lateral moraines at higher levels, only faint fragments of which survive. Nevertheless, these lateral moraine fragments, together with the more complete moraines on the lower slopes, have gradients which are entirely consistent with those of moraines upvalley and thus it is argued that these are part of the same system.

South of the River Liza, prominent glaciogenic landform assemblages cover a much more restricted area, extending not much more than 0.5 km upvalley from the forestry limit. In the vicinity of the river, these moraines comprise discontinuous, non-aligned

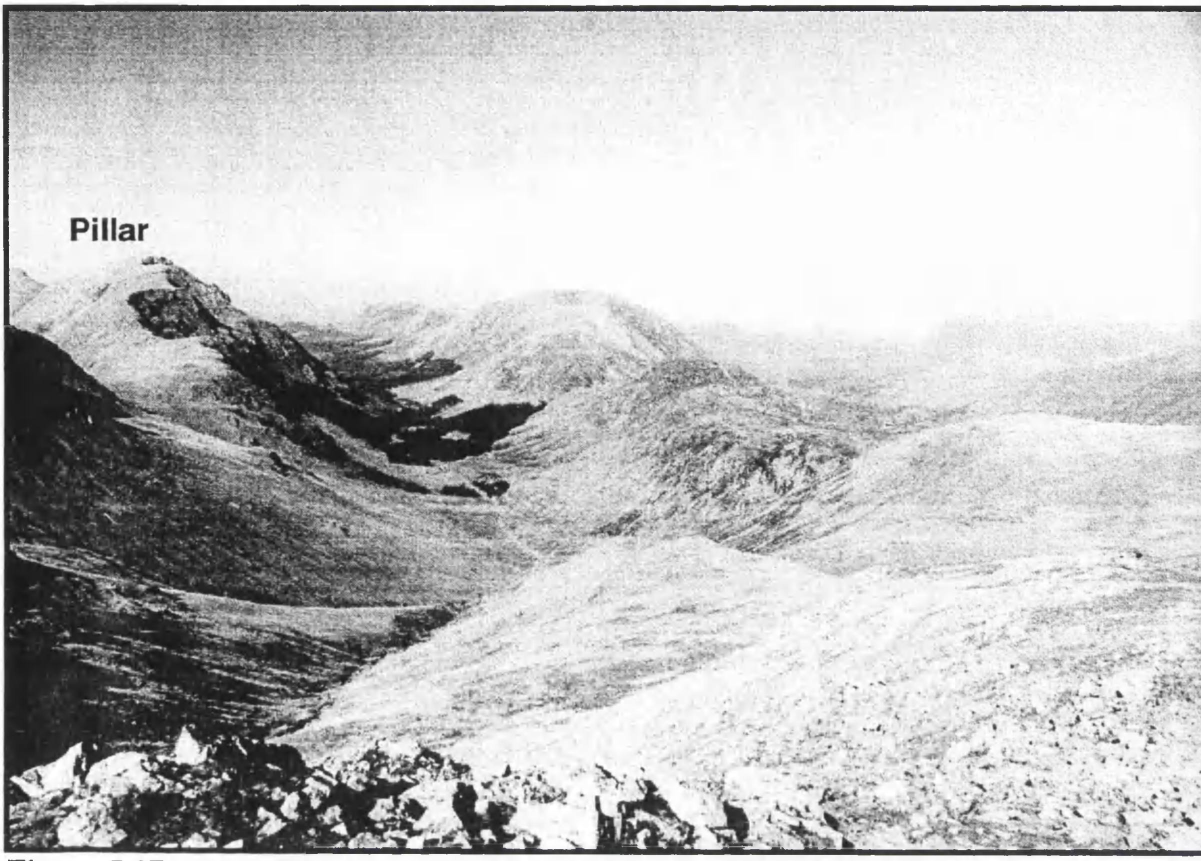


Figure 5.17 The Ennerdale valley. View looking west from Green Gable

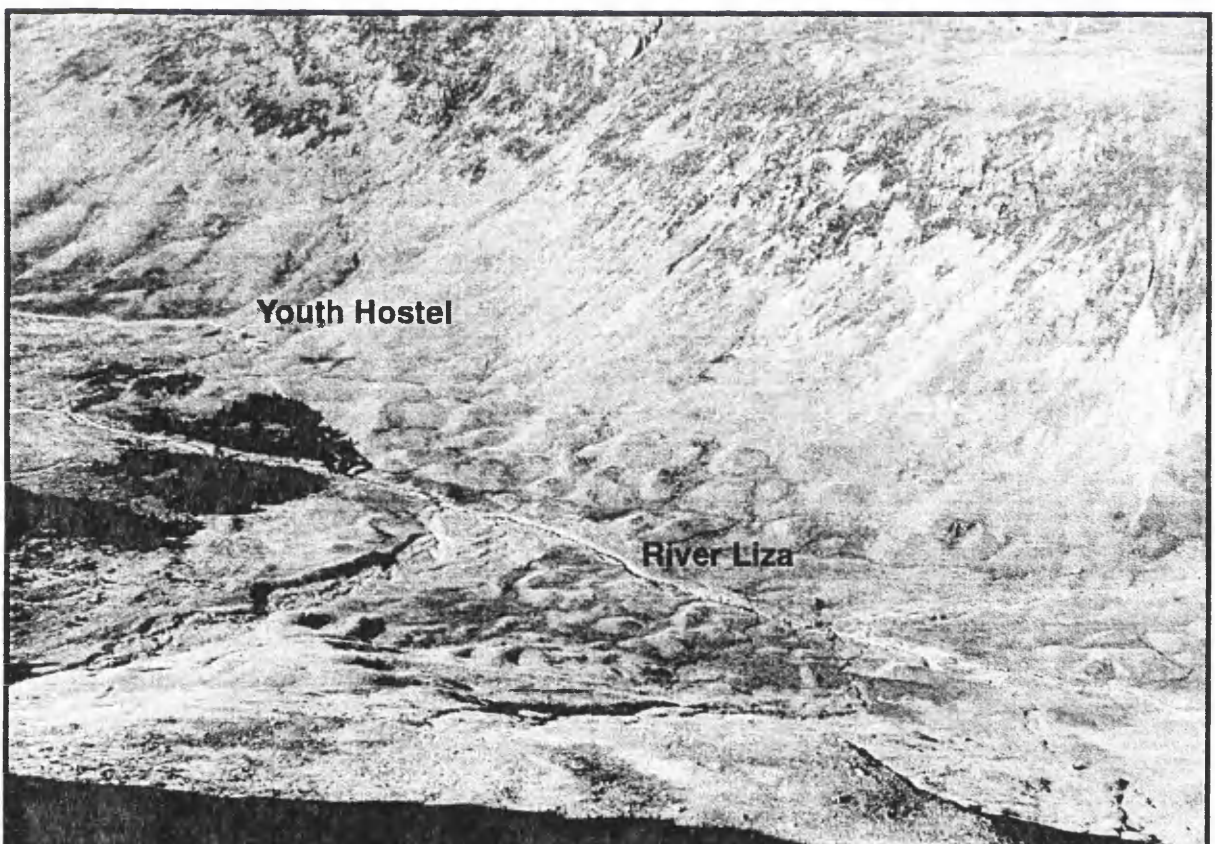


Figure 5.18 Loch Lomond Stadial glaciogenic landforms in upper Ennerdale. View looking northwest from Kirk Fell.

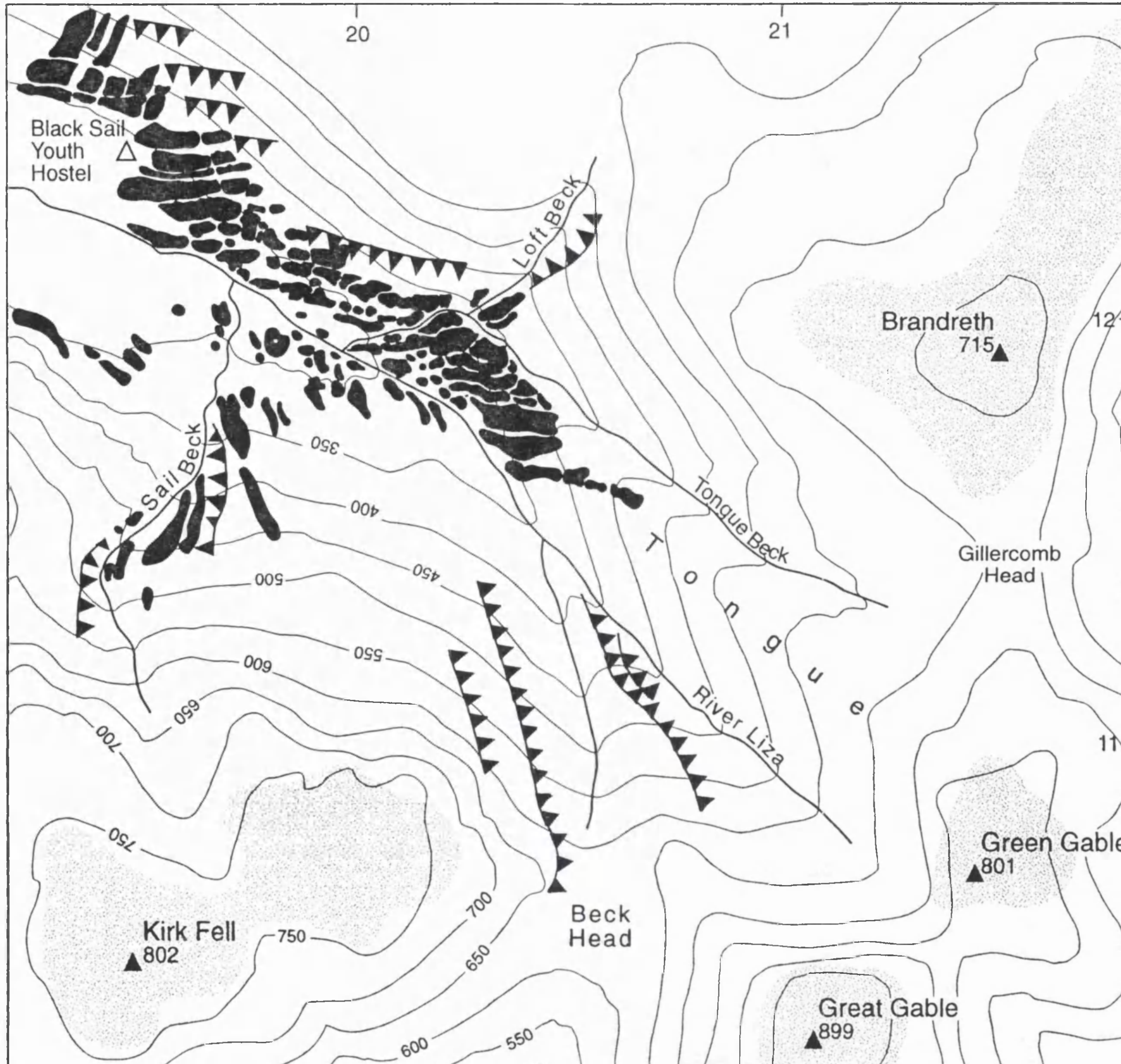


Figure 5.19 Loch Lomond Stadial moraines in Ennerdale. Scale and orientation given by British National Grid co-ordinates at 1km intervals. Key as for Figure 5.12 (except for blockfield - shown by grey shading).

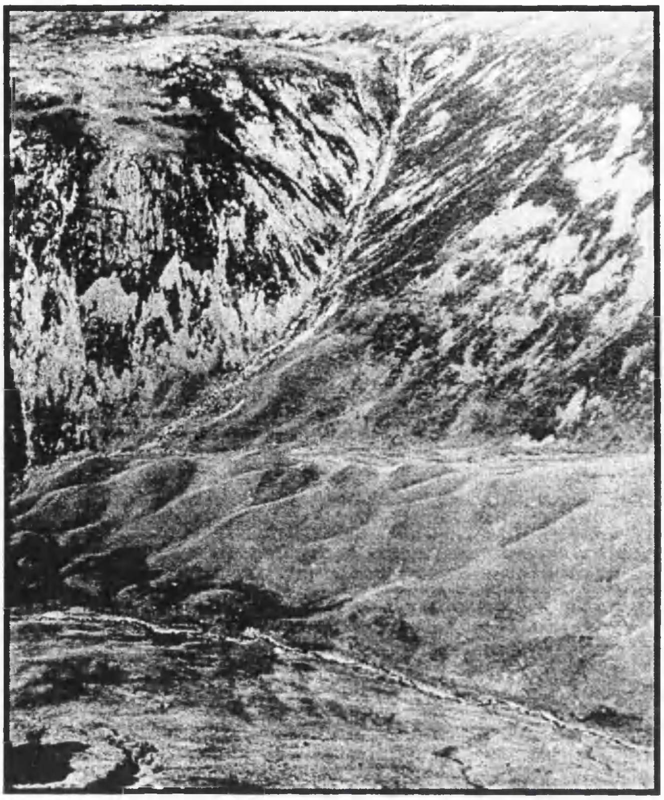


Figure 5.20
Lateral moraines on Tongue, Ennerdale

Figure 5.21
Hummocky moraines west of Loft Beck, Ennerdale

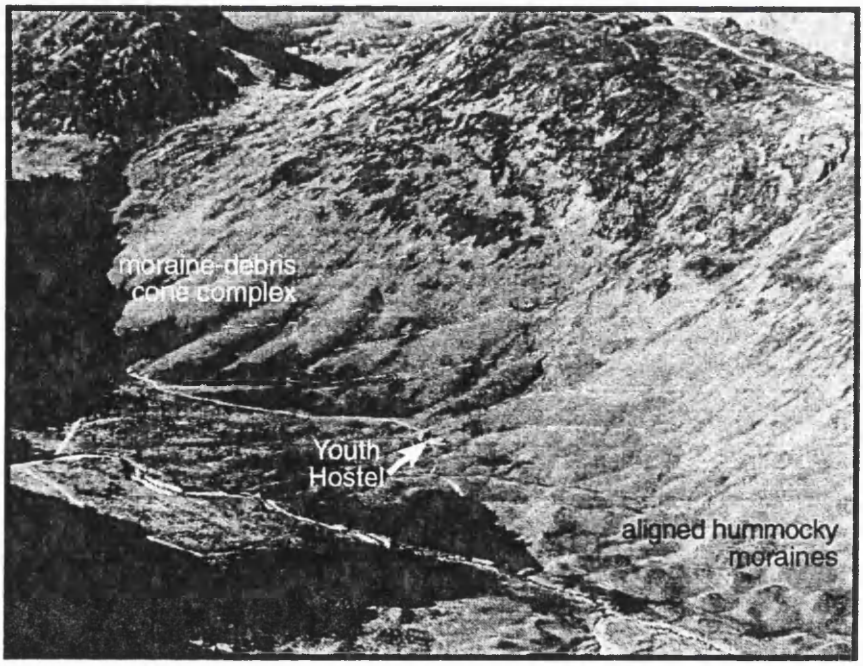


Figure 5.22
Moraine-debris cone complex west of Black Sail Youth Hostel

ridges and single or composite hummocks, which collectively convey a chaotic appearance. Nevertheless, there is a transition upslope to more ordered terrain, consisting of relatively subdued ridges which trend obliquely down the valley sides. Outwith this area, evidence for the Ennerdale palaeoglacier is limited to drift limits on the upper slopes.

On the basis of the geomorphological mapping shown in Figure 5.19, the glaciogenic landform assemblages in Ennerdale are interpreted as having been produced in association with an actively backwasting ice margin. As with the other sites investigated, the lack of sections means that this interpretation does not benefit from detailed sedimentological observations, which are clearly desirable in any discussion of moraine genesis (e.g. Benn, 1990). Nevertheless, in planform these moraines display a striking, if discontinuous, lobate arrangement which can only be explained as having been produced at successive margins of an actively backwasting glacier. The order is particularly clear north of the river where aligned hummocks and ridges trend obliquely down the valley sides at similar gradients, although few have corresponding ice-marginal moraines south of the river. The chaotic moraines which occur south of the river on the lower slopes are associated with lateral moraines on both the northern and southern slopes, which suggests that stagnation was limited to the ice margin on the valley floor. Eyles (1983) has argued that, if sufficient supraglacial debris was available, localised stagnation of the ice-margin would occur, a process he termed 'incremental ice-marginal stagnation.' Similar associations of valley floor hummocky moraines and valley side lateral moraines have been noted in Scotland (e.g. Benn, 1990; Bennett, 1991).

From the geomorphological evidence discussed above, a map showing approximate palaeoglacier configurations during deglaciation has been constructed (Figure 5.23). Commercial forestry plantations obscure the ground covered by the lower reaches of the glacier, and thus it has been necessary to extrapolate these limits downvalley from the lateral moraines and debris cones west of the Youth Hostel. Given the argument developed earlier that these debris cones were derived from Loch Lomond Stadial moraines, then it follows that their upper limits provide a minimum altitude for the ice surface.

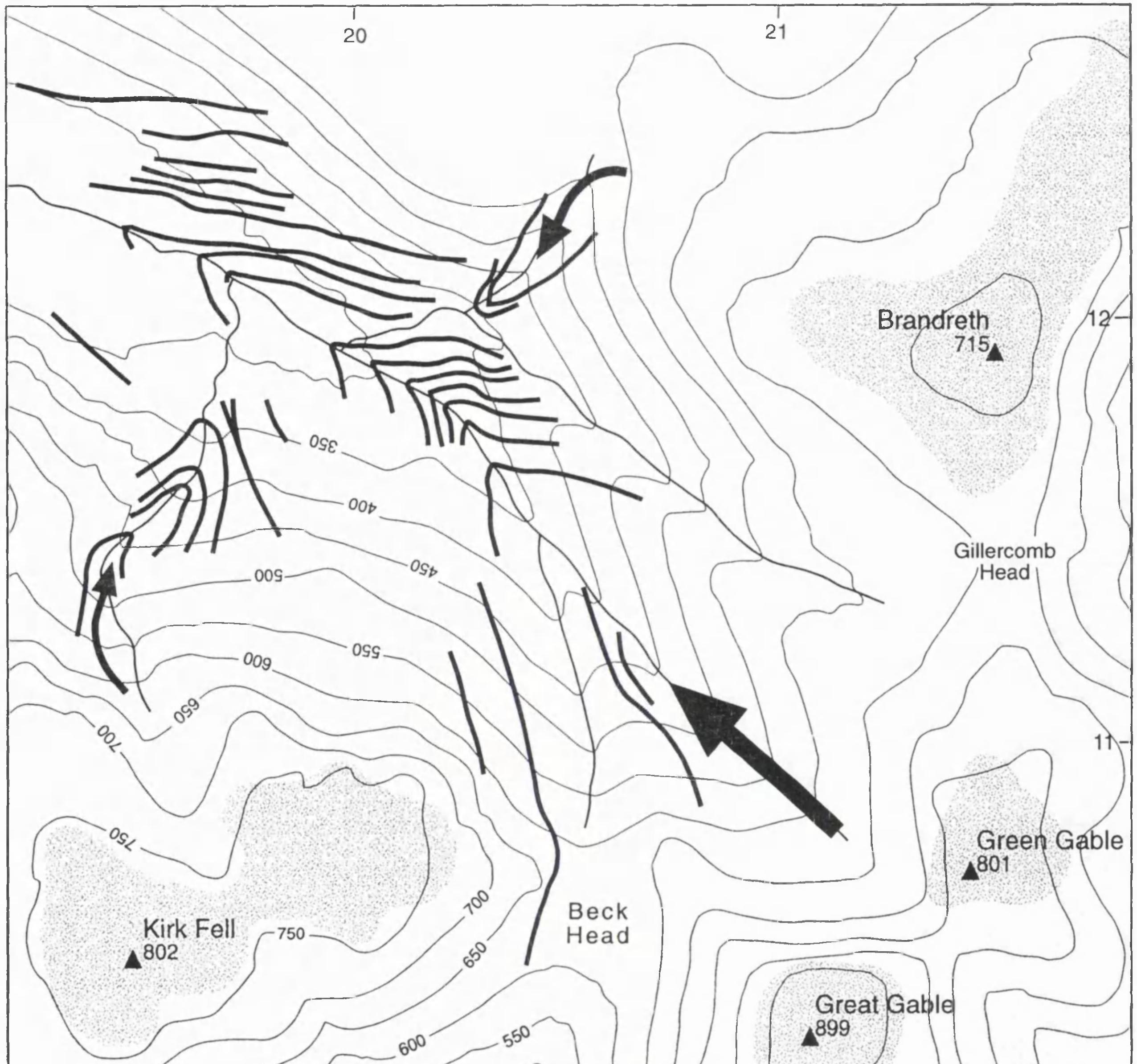


Figure 5.23 Loch Lomond Readvance decay pattern in upper Ennerdale.
Scale and orientation given by British National Grid co-ordinates at 1km intervals.

The existence of a contemporaneous plateau icefield on Grey Knotts/Brandreth is suggested by the uniformly steep gradients of the ice-marginal positions on the northern slopes. This arrangement can most readily be explained by ice from the plateau icefield descending relatively steeply into Ennerdale. More direct evidence for ice on the high ground west of Brandreth is provided by moraine fragments at the mouth of the Loft Beck gully, which define the former margins of a small tongue of ice descending from the higher ground above. Although a detailed sequence of deglaciation cannot be reconstructed on the basis of the available geomorphological evidence, faint drift limits near the valley head demonstrate that the contribution from Brandreth declined as deglaciation proceeded, with final deglaciation of the Ennerdale palaeoglacier centred on the sheltered ground overlooked by Great Gable and Green Gable.

The altitudes of the drift limits on the southern slopes of Ennerdale are consistent with the volumes of ice being proposed for the Ennerdale palaeoglacier. For example, the drift limits clearly show that at maximal extents, ice in upper Ennerdale overflowed through Beck Head (between Kirk Fell and Great Gable) to descend southwards for a short distance. There is also evidence for the development of a plateau icefield on Kirk Fell at this time. This takes the form of fragmentary ice-marginal moraines and drift limits either side of Sail Beck which appear to document ice-marginal positions of an outlet glacier during deglaciation.

5.2.5 Gillercomb

In contrast to Honister Pass and parts of upper Ennerdale, the glaciogenic landform assemblages in Gillercomb are relatively subdued (Figure 5.24). Single and composite hummocks, together with non-aligned ridges, convey a chaotic appearance in the lower reaches, although a transition to longitudinally-lineated forms occurs upvalley (Figure 5.25). These latter features are interpreted as flutings. Clasts obtained from shallow surface pits are typically sub-rounded and facetted, characteristics indicative of modification during active subglacial transport (cf. Boulton, 1978; Ballantyne, 1982). Scree control samples are substantially more angular.

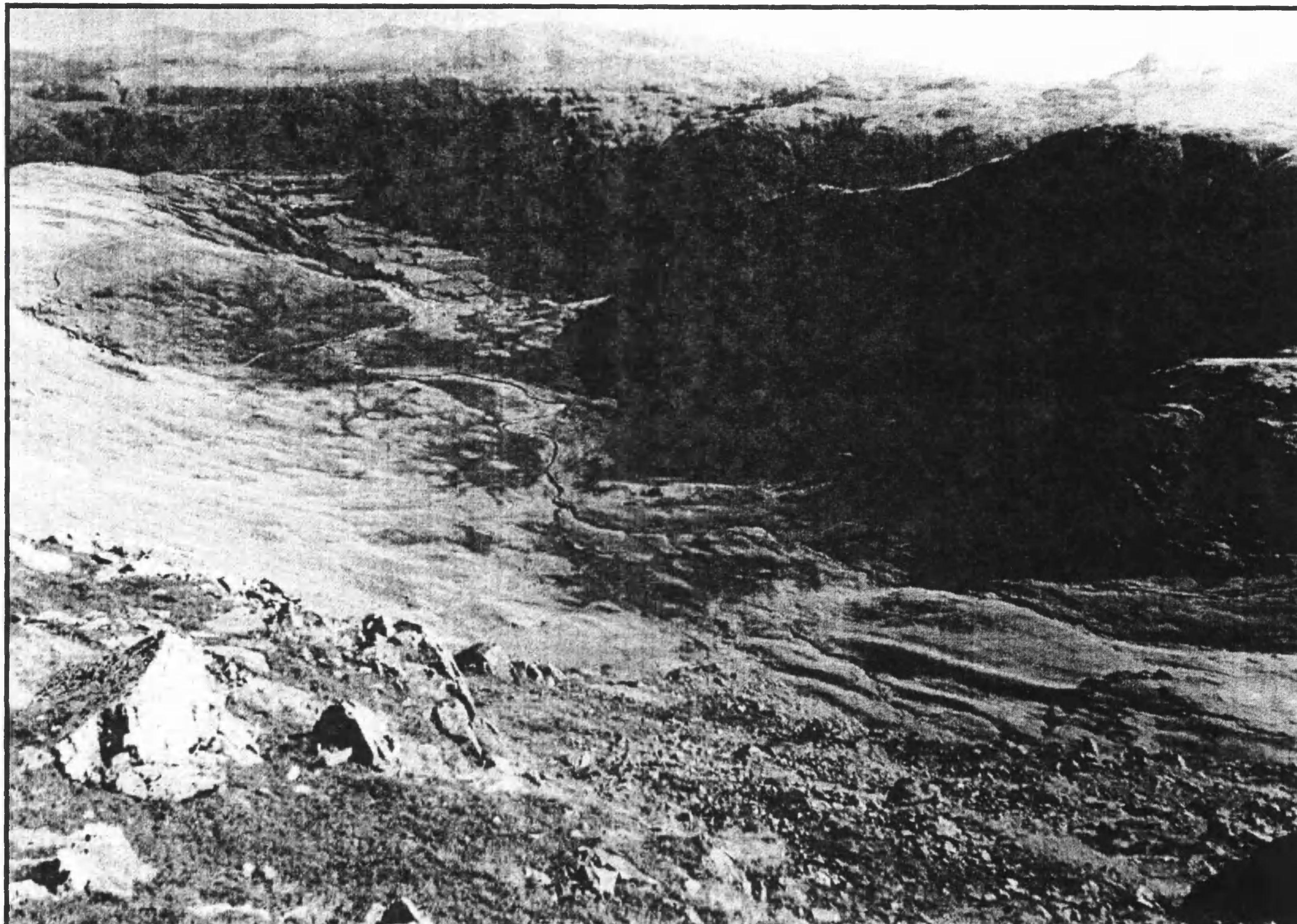


Figure 5.24 Loch Lomond Stadial moraines in Gillercomb

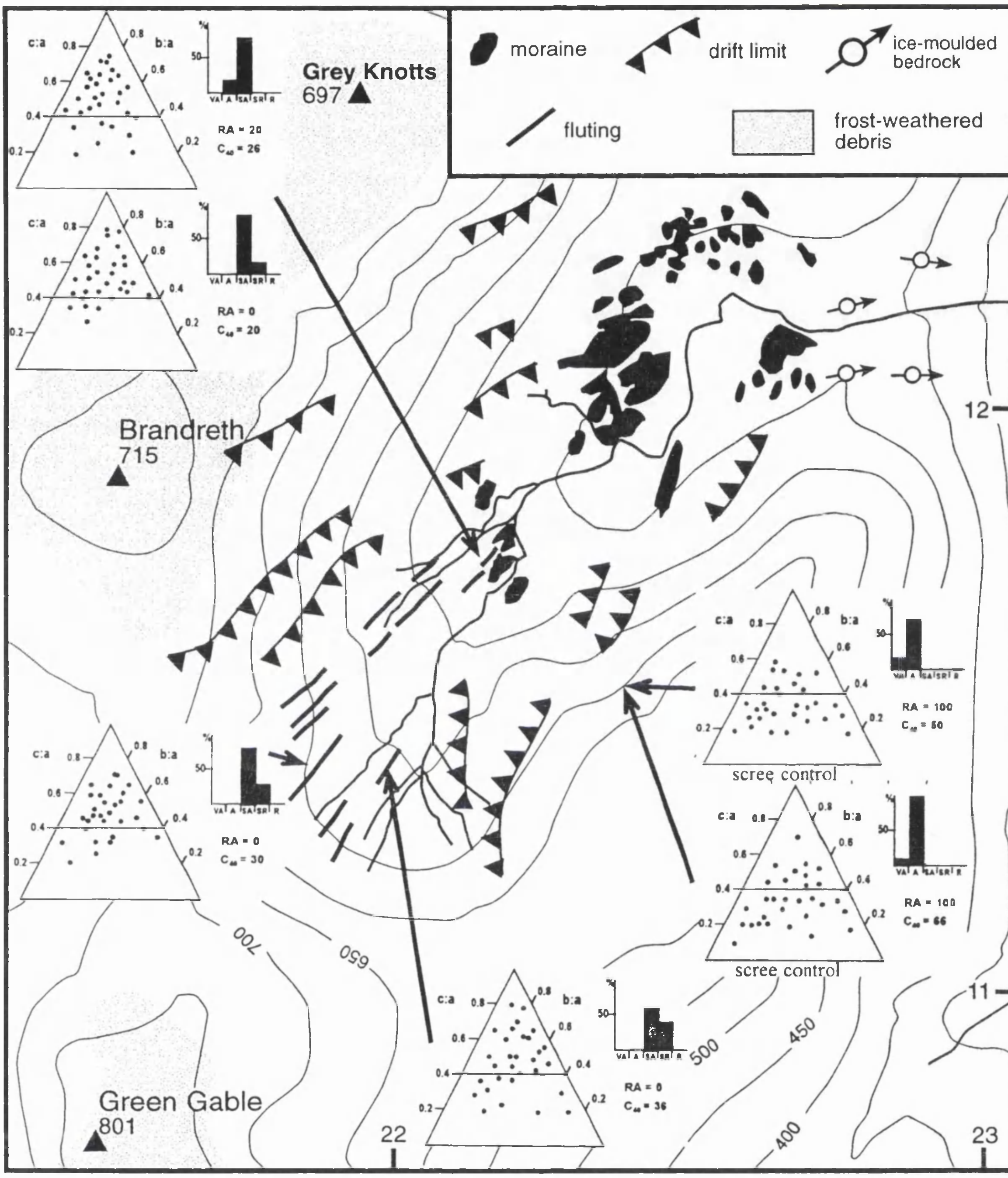


Figure 5.25 Map of Loch Lomond Stadial moraines in Gillercomb.
Scale and orientation given by British National Grid co-ordinates at 1km intervals.

The available geomorphological evidence in Gillercomb permits neither the delimitation of this palaeoglacier at maximal extent nor a detailed reconstruction of palaeo ice-marginal configurations during deglaciation. To a considerable extent, it is likely that the subdued nature of the geomorphological record reflects limited debris supply to the ice margin. The proximity and orientation of the Gillercomb valley with respect to the Grey Knotts/Brandreth plateau icefield would have ensured that, at maximal extent, ice entered the valley from both the valley head and via the western slopes. Given that no supraglacial debris would have been carried from the source area, the potential for supraglacial debris entrainment would have been very limited, dependent upon the contribution from slopes overlooking the eastern margin. Fragmentary drift limits clearly imply that as deglaciation proceeded the western slopes became exposed, with decay centred in the vicinity of Gillercomb Head rather than on the Grey Knotts-Brandreth summit.

It is likely that, at maximal extent, the Gillercomb palaeoglacier was confluent with the Seathwaite palaeoglacier. Although direct evidence for this is lacking, a combination of ice-moulded bedrock and locally thicker drift on the steep descent from Gillercomb to the Seathwaite valley may be associated with this event. Although the Seathwaite palaeoglacier was not considered as part of this research, fragments of lateral moraines at approximately 350 m OD on the eastern slopes of Seathwaite opposite Gillercomb suggest that, at maximal extent, it may have reached Seatoller or even Rosthwaite, which is substantially more extensive than suggested by the downvalley extent of prominent valley floor moraines. This is one of a number of examples considered in this investigation whereby the altitude and gradients of valley-side lateral moraines is at odds with limits proposed by other workers on the basis of the downvalley extents of prominent valley-floor moraines.

5.3 PLATEAU ICEFIELD RECONSTRUCTION

A tentative and, in places, somewhat speculative reconstruction of the plateau icefields which developed on Grey Knotts/Brandreth, Dale Head and Kirk Fell is presented in Figure 5.26. The general configurations of these coalescing systems are supported by the palaeo-ice-fronts considered in Section 5.2. For example, there is reasonably clear evidence to suggest that the Little Gatesgarthdale glacier was nourished by ice descending both the northern and southern slopes at the valley head, with a lobe of ice spilling westwards into Gatesgarthdale. On the other hand, these palaeo-ice-fronts were produced when deglaciation was well established and thus provide little information on ice-marginal positions at maximal extents, both in the upper and lower reaches.

In Ennerdale, forestry obscures the ground surface west of the Black Sail Youth Hostel. The downvalley limit of the Ennerdale palaeoglacier was therefore extrapolated from the uppermost recognisable drift limits on the northern slopes. There is no clear evidence for the maximal downvalley position attained by the small glacier which descended southwards between Kirk Fell and Great Gable. It is unlikely that moraines would have formed on these steep slopes. Limits have been extrapolated downvalley from fragmentary drift limits in the upper reaches. Lateral moraine fragments at NY217102 are the basis on which the narrow glacier between Great Gable and Green Gable has been inferred. Northeast of Green Gable, the contiguity of the ice surfaces in upper Gillercomb and Seathwaite (east of Green Gable) appears probable given the volumes of ice in the area, but direct evidence for this is lacking.

To the west of Grey Knotts/Brandreth, clear evidence for a former ice-margin is lacking. Of the two small tongues of ice which are inferred to have descended into Warnscale, the southern one is evidenced by a small lateral moraine. Whilst there is no geomorphological evidence that this was nourished by ice descending from higher ground, a valley glacier reconstruction (as proposed by Sissons - Section 5.4) results in a rather low firn line of 408 m (Sissons, 1980a). Evidence from High Raise suggests a figure of 500 m is more appropriate for the central fells.

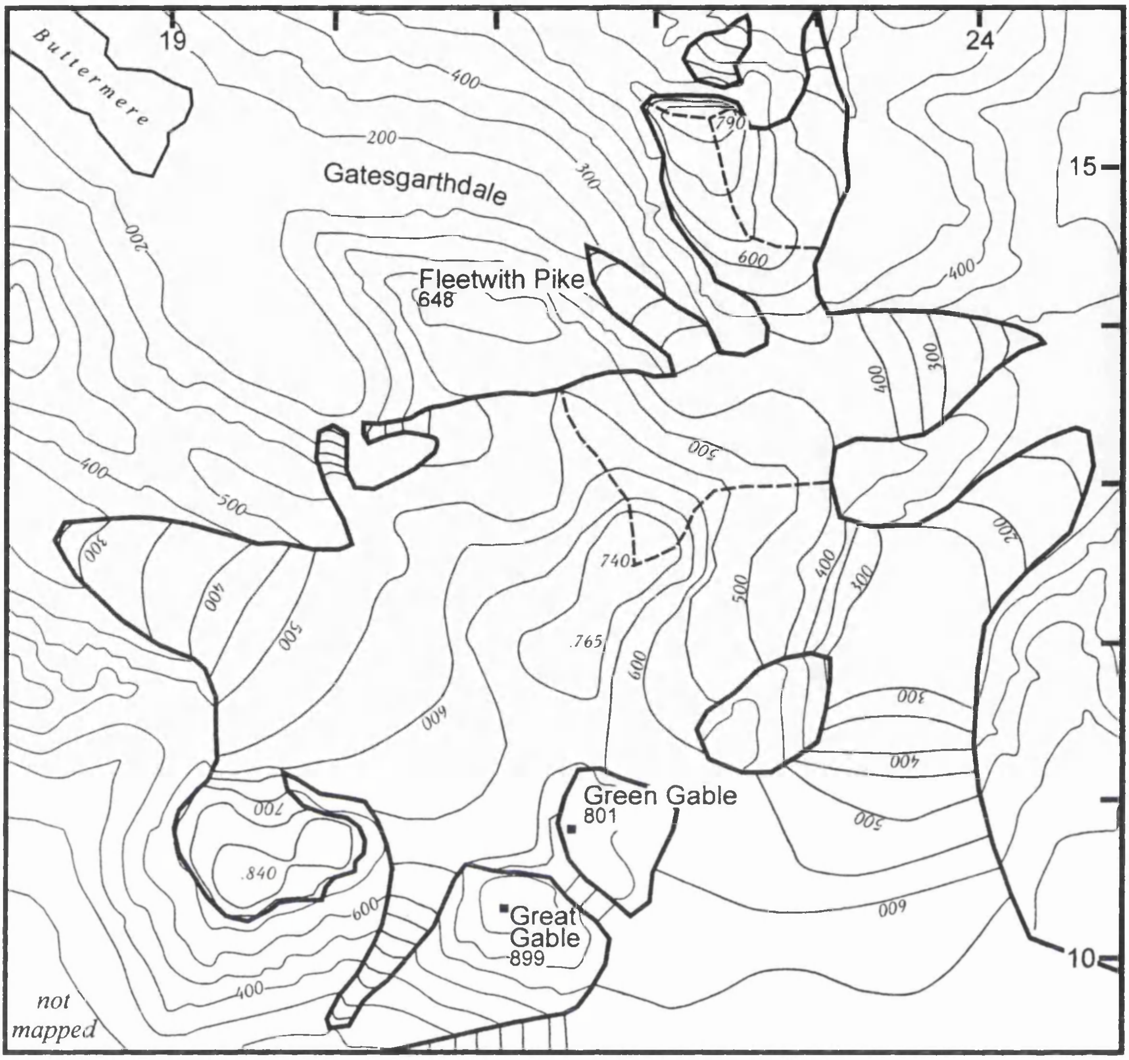


Figure 5.26 Reconstruction of the Loch Lomond Stadial plateau icefields which developed on Grey Knotts/Brandreth and Dale Head.
Scale and orientation given by British National Grid References at one kilometre intervals

As is the case with the High Raise plateau icefield system, uncertainty is introduced by the absence of ice-marginal control points in the upper reaches of these systems. Ice thicknesses on summits are estimates based on the theoretical relationship between plateau icefield depth and summit breadth shown in Figure 2.2.

Firn lines have been calculated for the glaciers in the Honister Pass area (elsewhere the glaciers merge with ice originating from outside the systems mapped) using Sissons's (1974) AWMA method (Section 3.4). This produced a firn line of 475 m, a value which is broadly consistent with those calculated for the High Raise plateau icefield system in Section 4.3.2.

5.4 COMPARISON WITH PREVIOUS WORK

Evidence has been presented in this chapter for the development of plateau icefields on Grey Knotts/Brandreth, Dale Head and possibly also on Kirk Fell during the Loch Lomond Stadial. Prominent glaciogenic landform assemblages in some of the surrounding valleys document successive ice-marginal positions of outlet glaciers during deglaciation. This interpretation differs from previous reconstructions in which an alpine style of glaciation has been assumed, with the hummocky moraines interpreted as evidence for the *in situ* stagnation of valley glaciers at or near maximal extents (e.g. Sissons, 1980a).

In mapping glaciogenic landforms in the valleys surrounding Grey Knotts, Brandreth and Dale Head, Sissons (1980a) was principally concerned with delineating the outlines of these former glaciers at maximal extents. The downvalley limits were defined on the basis of the areal extents of contiguous 'fresh' glaciogenic landform assemblages (Figure 5.27). Although the present worker did not employ moraine morphology as a relative dating technique, lithostratigraphic investigations in Little Gatesgarthdale and upper Ennerdale confirm a Loch Lomond Stadial age for these landforms. However, they also indicate a Loch Lomond Stadial age for the more rounded landforms in the lower reaches of Little Gatesgarthdale, which Sissons (1980a) considered to lie beyond the limits of the palaeoglacier. Geomorphological mapping reveals that these landforms, although relatively subdued, document ice-marginal positions which can be traced upvalley into the more prominent ('fresher') moraines upvalley. Thus, both lithostratigraphy and morphostratigraphy demonstrate that, for Little Gatesgarthdale at least, moraine morphology is an unreliable age discriminant at this locality.

There are a number of inter-related variables which will have influenced moraine morphology, including debris supply, ice margin behaviour during deglaciation, and paraglacial resedimentation following deglaciation. It is important to consider the likely effects of these when interpreting the glacial geomorphological record. On the northern slopes of Ennerdale, for example, ice-marginal moraines merge into debris cones derived from paraglacial reworking of moraines upslope. Whereas Sissons (1980a) considered that these 'smooth mounds of drift' pre-date the Loch Lomond Stadial, they are

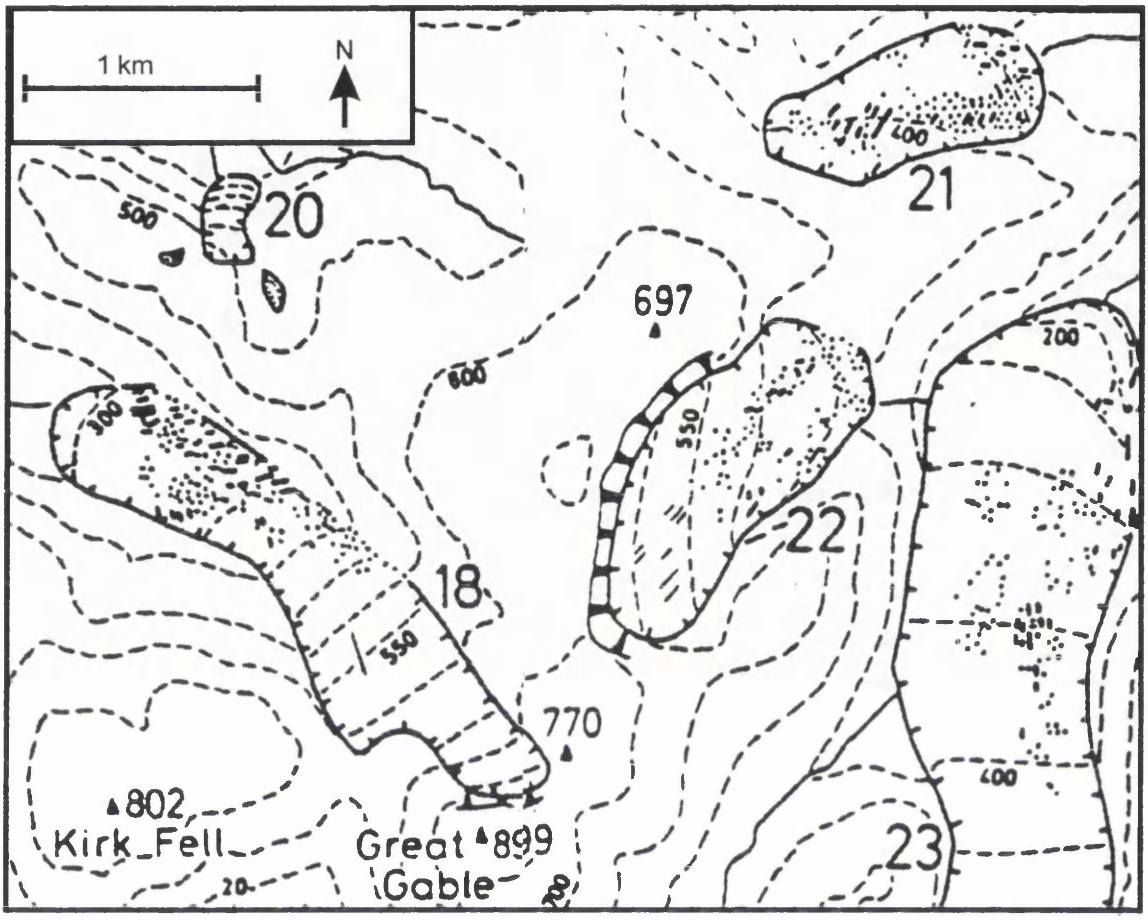


Figure 5.27 The Loch Lomond Readvance between Honister Pass and Ennerdale according to Sissons (1980a)
Key as for Figure 4.28

considered here to be contemporaneous on the basis of morphostratigraphy. Paraglacial activity has not entirely removed the evidence for ice-marginal positions, and relatively distinct lateral moraines immediately upvalley are continued by gentle ridges which trend obliquely across the surface of the debris cones. Indeed, the apexes of several debris cones are aligned and probably represent former ice surfaces. These ice-marginal positions trend obliquely to the near-horizontal glacier surface reconstructed by Sissons (1980a) for the lower reaches of the Ennerdale palaeoglacier. He considered that this low gradient, in combination with the well-developed hummocky moraine, could represent evidence for surging behaviour. It is argued here, however, that Sissons' glacier surface is actually based on an upslope transition from relatively prominent moraines on the lower slopes to thinner, drift-covered slopes in the steeper, upper slopes. The close proximity of successive ice-marginal positions serve to create the impression of an almost continuous drift 'limit.'

Differences in the interpretation of the geomorphological record also occur in the upper reaches of these former glaciers. Periglacial trimlines are not present in any of the valleys investigated and Sissons (1980a) extrapolated ice-marginal positions into these former accumulation zones, assuming an alpine style of glaciation. This contrasts with the present investigation in which detailed mapping has enabled successive ice-marginal positions to be reconstructed and decay patterns identified. On this basis, Loch Lomond Stadial plateau icefields are considered to have developed on Dale Head, Grey Knotts/Brandreth, and Kirk Fell. Although much of the evidence is faint and fragmentary (which is to be expected given the absence of periglacial trimlines), Sissons (1980a) presented no evidence whatsoever from these upper reaches. The extent of frost-weathered debris suggests that the geomorphological impact of these icefields was minimal on the summits.

5.5 SUMMARY

Evidence has been presented in this chapter for the development of small Loch Lomond Stadial plateau icefields on Grey Knotts/Brandreth, Dale Head and probably also on Kirk Fell. These covered areas of 7 km^2, 3 km^2 and 1 km^2 respectively. The geomorphological impact of these plateau icefield systems appear to have been minimal on the summits where blockfield and other frost-weathered debris implies that glacial erosion was minimal on summit areas. This is interpreted as evidence for at least patchy cold-based ice. The occurrence of ice-moulded bedrock at a number of localities on the western slopes of Grey Knotts/Brandreth suggests a transition to wet-based, erosive ice on steeper slopes, where strain heating would have been greater.

Reconstructed palaeo ice-marginal positions indicate that the glaciogenic landform assemblages east of the Honister Pass watershed were produced by plateau icefield outlet glaciers which descended from these summits. Similarly, reconstructed ice-marginal positions on the northern slopes of Ennerdale provide additional evidence for a contemporaneous Grey Knotts/Brandreth plateau icefield. The Kirk Fell plateau icefield is inferred from ice-marginal moraines on the southern slopes of Ennerdale which appear to document successive positions of an ice-tongue as it backwasted in the direction of the summit.

6

Discussion

6.1 INTRODUCTION

This study demonstrates that the Loch Lomond Stadial glaciation in the Lake District was characterised by plateau icefields, which at least partially nourished the corrie and valley glaciers already reported in the literature (e.g. Manley, 1959; Sissons, 1980a). Geomorphological evidence has been presented for the development of a plateau icefield system centred on High Raise in the central fells (Chapter 4). Including outlet glaciers, this system covered an area of approximately 55 km^2. To the west, smaller plateau icefields developed on Grey Knotts/Brandreth, Dale Head and probably also on Kirk Fell (Chapter 5), covering areas of 7 km^2, 3 km^2 and 1 km^2 respectively (Figure 6.1).

The reconstructions of these plateau icefield systems are somewhat speculative, inevitable given the virtual absence of ice-marginal control points in the upper reaches of these systems (Sections 4.3.2 and 5.3). Ice thicknesses on the summits are estimates based on the theoretical relationship between plateau icefield depth and summit breadth (Section 2.2.2), with general configurations extrapolated from palaeo-ice-marginal positions produced during deglaciation. This situation contrasts markedly with the more extensive icefield which developed on Skye at this time, where former ice margins in the accumulation zone are documented by relatively clear periglacial trimlines (Ballantyne, 1989). Similar uncertainties surround outlet glacier extents. Although the downvalley extents of some outlet glaciers appear to be defined by relatively clear end moraines (for example, at Wythburn and Rosthwaite), the evidence is generally subtle and fragmentary.

The failure of Sissons (1980a) to recognise the existence of these former plateau icefield systems most probably reflects a number of factors. Firstly, and arguably most importantly, he assumed an alpine style of glaciation when reconstructing Loch Lomond Stadial ice masses in the Lake District (and other parts of upland Britain supposedly marginal for glaciation at this time – Wales, the Southern Uplands and the Cairngorms) (Section 2.4.2.3). In doing so, no explicit consideration was given to the nature of the terrain, despite the fact that glaciation styles are determined by interactions between topography and climate (Section 2.3). Secondly, there was little or no appreciation at that time of the difficulties surrounding the recognition of plateau icefields in the geomorphological record. These have only become apparent following investigation of

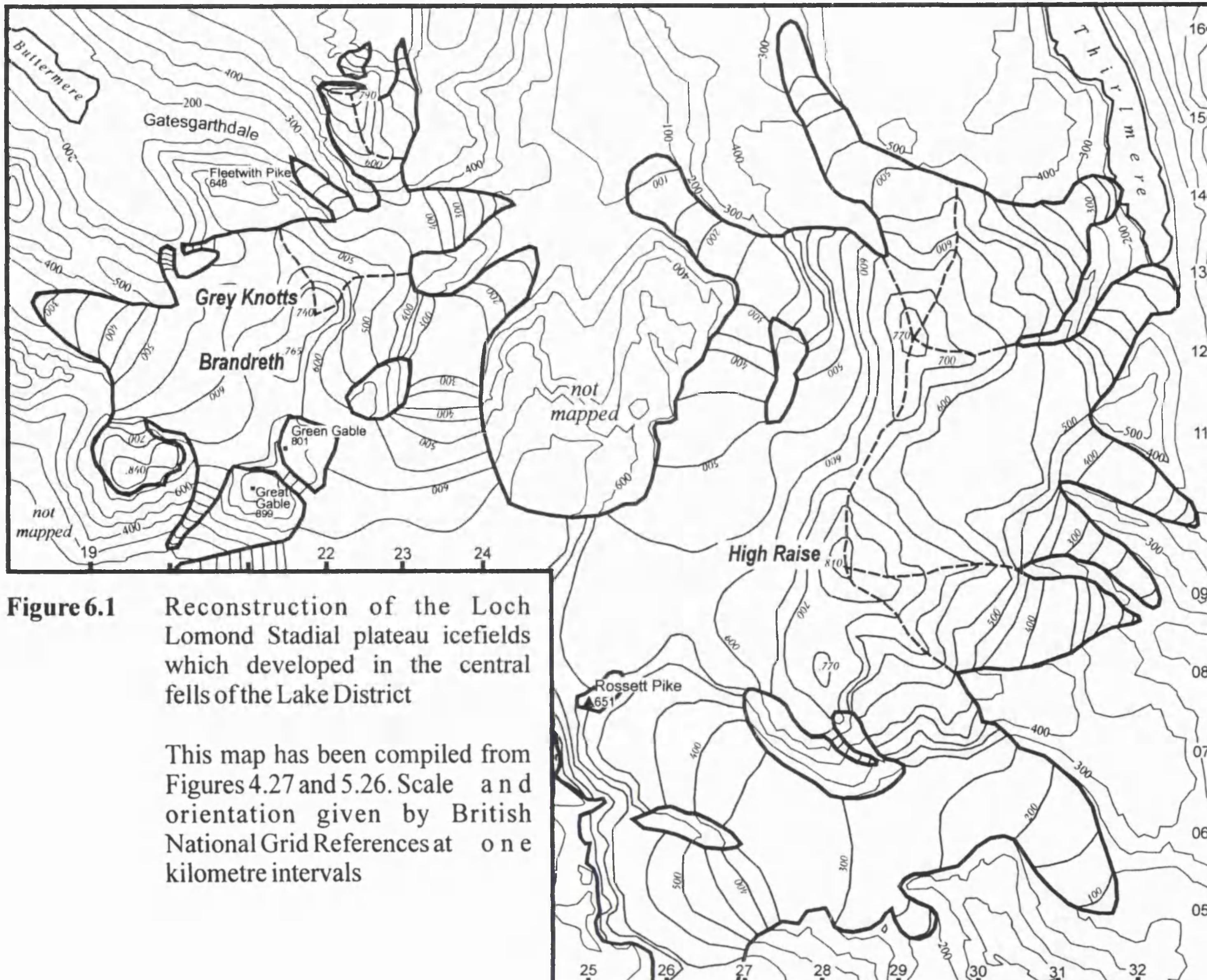


Figure 6.1 Reconstruction of the Loch Lomond Stadial plateau icefields which developed in the central fells of the Lake District

This map has been compiled from Figures 4.27 and 5.26. Scale and orientation given by British National Grid References at one kilometre intervals contemporary plateau icefields in north Norway (e.g. Gordon *et al.*, 1987; Gellatly *et al.*, 1988), and the present study is the first to relate these observations to the geomorphological record in upland Britain, a completely deglaciated area. Finally, the identification of these former plateau icefield systems in the central Lake District follows detailed geomorphological mapping of successive ice-marginal positions associated with deglaciation. Until relatively recently, it was believed that Loch Lomond Stadial glaciers stagnated *in situ* at or near maximal extents following rapid thermal amelioration (Section 2.4.2.1). In this study, the ability to identify decay centres is critical to the reconstruction of these plateau icefield systems.

The question thus arises as to whether former plateau icefield systems elsewhere have been incorrectly reconstructed as cirque/valley glaciers adjacent to ice-free summits. This applies not only to the British Isles but also to any other part of the world where assumptions about glaciation style have been made without due consideration to the nature of the terrain. The question is important because the failure to account for former plateau icefields will result in an overestimation of ELA lowering. The palaeoclimatic significance of this is considered in Section 6.2. This is followed by a discussion of the geomorphological evidence which may be used to reconstruct former plateau icefields elsewhere in the Lake District and beyond. Where the geomorphological evidence is equivocal, however, the identification of former plateau icefields will require an assessment of topoclimatic factors (Section 6.4). Approaches to reconstructing the surface profiles of former plateau icefields are discussed in Section 6.5, which is followed by a consideration of dating issues.

6.2 PLATEAU ICEFIELDS AND PALAEOCLIMATIC RECONSTRUCTIONS

Former glaciers have been used in many parts of the world as a basis for deriving palaeoclimatic inferences. In the British Isles, for example, glacier distributions, dimensions and altitudes suggest that the principal snow bearing winds during the Loch Lomond Stadial were south-easterly or south-westerly, with mean July temperatures of 6–8°C in northern Britain. The significance of the present research is that it shows how traditional approaches to palaeoglacier reconstructions can fail to take into account the former existence of relatively small plateau icefields. In turn, this will produce erroneous palaeoclimatic inferences.

The failure to account for former plateau icefields is significant where these were vital for the maintenance of glaciers in the surrounding corries and valleys, and will result in an overestimation of ELA lowering. Of the examples considered in this investigation, the largest ELA adjustment affects the Honister Pass palaeoglacier. Whereas Sissons' reconstruction of this former ice mass produced an ELA of 385 m, the addition of a plateau icefield source increases this value to 475 m (Section 5.3). In many of the cases considered, however, the impact of accumulation at higher levels on the ELA is partly offset by more extensive downvalley positions.

As the size of the plateau icefield increases in relation to the valley glacier component, so the magnitude of the error will increase (Rea *et al.*, in prep.). Thus, it should be expected that the potential for ELA discrepancies will be highest in areas characterised by extensive, high level plateaux (e.g. the Cairngorms). Relative relief is also important. It follows that as the altitudinal difference between the plateau top and the valley bottom increases, so the potential magnitude of the error will increase. Therefore, errors will be most pronounced in high relief mountain environments with extensive plateaux, such as parts of Norway. In the Lyngen Peninsula, precipitous cliffs more than 1000 m high separate icefields from the valley glaciers below (Gordon *et al.*, 1987, 1988). The valley glaciers appear to exist entirely within the ablation zone and can only survive because they are nourished by ice avalanching from the icefield above.

The failure to account for former plateau icefields in low relief mountain environments which lack extensive plateaux, such as the Lake District, can still be of significance in palaeoclimatic reconstructions. The plateau icefield reconstructions proposed in this investigation, for example, remove the need to invoke, as Sissons (1980a) did, marked snowfall variations over geographically limited areas or unusually high inputs of wind-blown snow in former accumulation zones (Section 2.4.4.5). Instead, variations in the extent of glacierization can be attributed to topoclimatic factors, specifically whether or not accumulation took place on the summits. Thus, relatively extensive outlet glaciers which debouched from plateau icefields are not inconsistent with much smaller ice masses which developed below narrow summits. It is likely that local variations in palaeoglacier development elsewhere in Britain and beyond may be explained more simply by summit accumulation rather than by invoking elaborate variations in snowfall intensity.

Revised estimates of palaeoprecipitation levels suggests that conditions in the central Lake District during the Loch Lomond Stadial were slightly more arid than Sissons envisaged. Palaeoclimatic inferences can be derived from reconstructed glaciers with the aid of graphs, such as the one shown in Figure 6.2, which relate accumulation at the ELA with mean summer temperatures for representative contemporary glaciers. Analyses of Coleopteran assemblages obtained from two sites in northern England suggest mean July temperatures at sea level were approximately 9°C during the Loch Lomond Stadial (Coope, 1994; Lowe *et al.*, 1995a). Using Figure 6.2, a mean July temperature of 9°C (equivalent to a mean summer temperature of 7.3°C) translates to a mean summer temperature at the firn line of 4.4°C (assuming a lapse rate of 0.58°C/100 m). In turn, this implies a mean annual precipitation of 2,000–2,500 mm. This is somewhat lower than the estimate of 2,700–4,000 mm provided by Sissons (1980a) (Section 2.4.4.5).

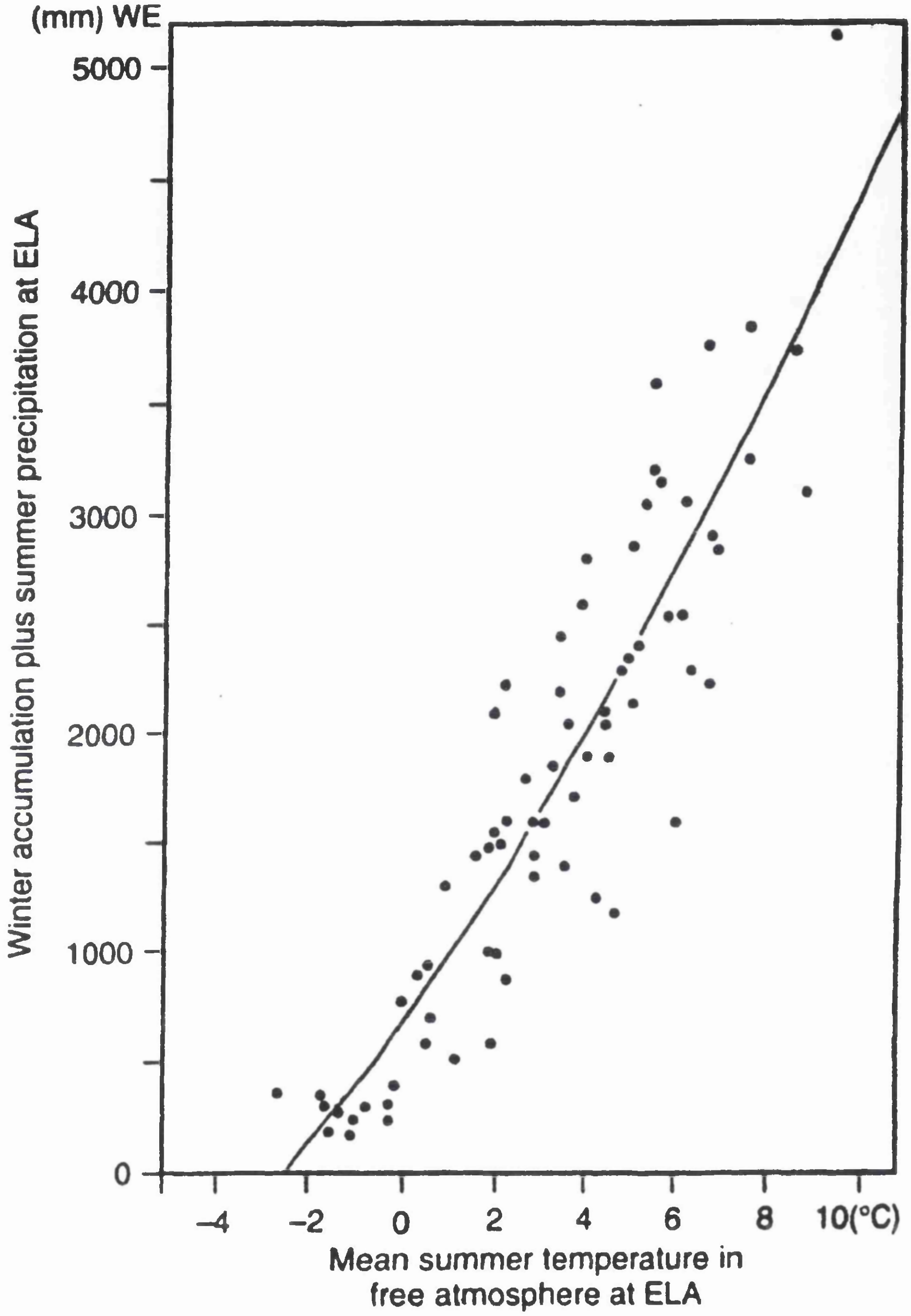


Figure 6.2 Annual precipitation and free-atmospheric temperature observed at the equilibrium line altitudes for seventy glaciers

Adapted from Benn and Evans (1998, Figure 2.23, redrawn from Ohmura et al., 1992)

6.3 PLATEAU ICEFIELDS AND THE GEOMORPHOLOGICAL RECORD

6.3.1 Introduction

Although the Loch Lomond Stadial plateau icefields which developed in the central fells of the Lake District appear to have had a minimal geomorphological impact on the summits, prominent moraine systems were produced by outlet glaciers which descended into surrounding valleys, where their margins became sediment traps for supraglacial debris and inwash. At a number of sites, it has been possible to map successive palaeo ice-marginal positions of outlet glaciers in the final stages of deglaciation and demonstrate that decay was centred on summits rather than valley heads.

Periglacial trimlines have been employed in reconstructions of Loch Lomond Stadial ice masses, such as the icefields which occupied the Western Grampians (Thorp, 1981, 1986) and the Cuillins on Skye (Ballantyne, 1989). However, valley-side trimlines do not occur everywhere. Thorp (1981) noted that Loch Lomond Stadial periglacial trimlines may be poorly developed or absent on frost-resistant lithologies such as lavas, coarse-grained granites and massive schists. This may explain the virtual absence of periglacial trimlines in the areas investigated in this study: the central fells of the Lake District are underlain by the Borrowdale Volcanic Group.

The virtual absence of periglacial trimlines in the central fells has meant that decay patterns recorded by ice-marginal moraines in the valley heads have been critical to the identification of former plateau icefields in the geomorphological record. This section briefly reviews the range of evidence presented. In so doing, it is hoped that the identification of former plateau icefields elsewhere in the Lake District and beyond will be made easier.

6.3.2 Summit Geomorphology

Various researchers have employed contrasts in weathering and periglacial landform development to assist in the delimitation of the higher reaches of Loch Lomond Readvance glaciers. This followed observations which suggested a mutually-exclusive relationship between the distribution of relict major periglacial landforms (such as blockfields and large solifluction features) and that of former glaciers (e.g. Sissons, 1967, 1972, 1974, 1976, 1979; Sissons and Grant, 1972). Thus, the presence of such features has been used as an indicator of areas not covered by Loch Lomond Readvance glaciers. Nevertheless, such an interpretation takes no account of the possibility that these periglacial landforms may pre-date the Loch Lomond Stadial and have survived beneath thin, cold-based ice on summits. This study has shown this to be the case for the blockfield and other frost-weathered debris which mantles High Raise, Thunacar Knott, Ullscarf, Grey Knotts/Brandreth, Dale Head and possibly also Kirk Fell (Chapters 4 and 5). This interpretation rests heavily on the assumption that these frost-weathered deposits could not have formed under the mild conditions of the present interglacial (e.g. Ballantyne and Harris, 1994).

The preservation of periglacial phenomena beneath plateau icefields requires cold-based, non-erosive ice. Such conditions are promoted by the dynamics of relatively thin, slow-moving ice, plus the effects of low mean annual air temperatures. Rea (1994b) calculated the relationship between plateau icefield thickness and summit breadth by employing a parabolic equation used to estimate ice sheet profiles (Section 2.2.2; Figure 2.2). For the Lake District plateau icefields, this suggests ice thicknesses of between 40 m and 50 m, although Rea (1994b) has stated that actual figures are likely to be even lower than those calculated due to the draw-down effect of outlet glaciers.

It is most unlikely that such conditions were unique to the Lake District summits during the Loch Lomond Stadial, which means the presence of well-developed periglacial phenomena on summits and plateaux elsewhere may not necessarily indicate an absence of glacier ice at this time. The survival of periglacial phenomena beneath former cold-based plateau icefields has been described by Evans (1988, 1990) in northwest Ellesmere Island where deglaciated summits lack obvious glacial erosional features and are often characterised by well-developed patterned ground and tors. In some places, however, ice-marginal meltwater channels define the successive margins of plateau icefields during deglaciation. Nevertheless, ice-marginal channels appear to be largely absent from the Lake District summits investigated, the exception being to the west and north of Ullscarf (Figure 4.24). It is possible that meltwater channels are masked by the extensive blanket peat which has developed on some of the broader summits, notably High Raise. On the other hand, they may not have formed at all if rates of meltwater production were low. This explanation has been proposed to account for the absence of channel formation at the retreating margins of the cold-based Balgesvarri plateau icefield in northern Norway (Gellatly *et al.*, 1988).

Similarly, moraines appear to be absent from the summits investigated (although it is possible that subdued ridges may be masked by blanket peat deposits). The absence of ice-marginal moraines on the summits investigated may reflect a number of factors. Firstly, and perhaps most importantly, it seems almost certain that these summits were entirely covered by ice during the Loch Lomond Stadial. Thus, it seems probable that moraines could only have formed at a relatively advanced stage in deglaciation, and then only if certain conditions were met. Deglaciation would have had to have been characterised by stillstands and/or readvances in the final stages of decay, and a source of debris for moraine formation would have been required. As far as the latter is concerned, an absence of a supraglacial debris source on the summits would have required debris to be derived subglacially and/or proglacially. Observations around Öksfjordjökelen in northern Norway led Gellatly *et al.* (1988) to conclude that subglacial erosion by thin, warm-based plateau icefields has been effectively limited to the evacuation of pre-existing blockfield and the resulting formation of boulder moraines.

The only clear high-level geomorphological evidence for former plateau icefields occurs at the margins of several summits where ice-moulded bedrock documents a transition to wet based, erosive conditions. Wet-based conditions are most likely to occur at plateau margins and above valley heads where increasing gradients result in increased strain heating. Perhaps the most impressive example occurs north of High Raise summit at Long Crag, where the wet-based, erosive conditions implied by the ice-moulded bedrock contrasts with the blockfield and other frost-weathered debris (indicative of cold-based ice) a short distance upslope (Section 4.2.1; Figure 4.6).

This investigation therefore provides clear support for the concerns expressed by Gordon *et al.* (1987, 1988) and Gellatly *et al.* (1988) regarding the identification of former plateau icefields in deglaciated areas (Section 2.2). Working at the margins of some contemporary plateau icefields in north Norway, they observed that recently deglaciated terrain exhibited little or no evidence for subglacial erosion, a situation which they attributed to low basal shear stresses and, in some places, cold-based ice. They argued that such conditions may have characterised areas long since deglaciated and, if so, little or no evidence for such plateau icefields would remain. Nevertheless, the extent to which plateau icefield systems in north Norway represent suitable analogues for Loch Lomond Stadial plateau icefields is a moot point. In the Lyngen Peninsula, for example, plateau icefields are separated from valley glaciers below by precipitous cliffs more than 1000 m high. The Loch Lomond Stadial plateau icefields which developed in the gentler relief of the Lake District were, by contrast, drained by outlet glaciers which descended into the surrounding valleys where they produced prominent moraine systems. This difference is significant because, at a number of sites, it has been possible to map successive palaeo-ice-marginal positions of outlet glaciers in the final stages of deglaciation and show that decay was centred on summits rather than valley heads.

6.3.3 Ice-marginal moraines

Ice-marginal moraines produced by actively-backwasting outlet glaciers in valleys surrounding the summits have proved to be the single most important line of evidence in the search for former plateau icefields in the Lake District. As explained in Chapter 3, only those moraines which appear to reflect the geometry of a former ice margin are referred to as 'ice-marginal' in this study, thus excluding moraines which are truly chaotic. Due to the paucity of sections within the moraines investigated, it was not possible to distinguish different types of ice-marginal moraine (e.g. push, dump etc.). The absence of this information does not compromise the study. Indeed, moraines are usually mapped independently of their internal compositions, which may vary significantly over short distances (e.g. Frye and Willman, 1962; Rose and Menzies, 1996). The morphological emphasis of this study is similar to the approach adopted by Benn (1990) and Bennett (1991) in their reconstructions of the decay patterns of Loch Lomond Stadial glaciers in the Northwest Highlands and islands of Scotland. They argue that by mapping the cross-valley concentric structures within Loch Lomond Stadial glacigenic landform assemblages, it is possible to obtain a picture of the pattern of deglaciation for these former glaciers.

The former existence of a plateau icefield may be inferred where the direction of deglaciation recorded by ice-marginal moraines differs from that which would be predicted for a valley glacier. This is the case, for example, with the suite of hummocky recessional moraines in Stake Pass and Langdale Combe, southwest of High Raise in the central Lake District (see Chapter 4). These moraines, which have been palynologically dated to the Loch Lomond Stadial by Walker (1965), define successive margins of an outlet glacier which actively-backwasted towards its source on High Raise, perpendicular to the valley axis. If these moraines had instead been produced by a valley glacier, as both Manley (1959) and Sissons (1980a) believed, then the direction of decay would have been at right angles to this, towards the valley head (Section 4.2.2.1 and 4.4)

A former plateau icefield may also be inferred where ice-marginal moraines can be shown to extend beyond the topographic confines of the valley and onto the higher ground above. This is the most common basis by which former plateau icefields have been inferred in the Lake District. In Honister Pass, for example, relatively clear moraines on the valley floor and lower slopes on the south side of the valley can be traced on aerial photographs more or less continuously upslope and beyond the confines of the valley onto gentler ground above. Although faint, these ice-marginal moraines and drift limits demonstrate the existence of ice at higher levels (Section 5.2). Similar evidence is provided by faint ice-marginal moraines which trend up the western slopes of Greenup and onto the higher ground above (Section 4.2.2.3). In both cases, however, the drift lineations become indistinct on the higher ground.

Perhaps the most significant obstacle to inferring former plateau icefields on the basis of high-level moraines and drift limits is their invariably subtle and fragmentary nature. To some extent, this can be attributed to paraglacial reworking following withdrawal of the ice margin. The efficacy of mass wasting processes has been emphasised by Ballantyne and Benn (1994a) who, working in Norway, have described the effects of paraglacial activity on recently deglaciated slopes on the northern side of Fåbergstølsdalen. The sediments derived from intense gullying in the upper sections of the steep drift-covered slopes have been redeposited on the lower slopes as coalescing debris cones. In some sections, the lateral moraine has been completely removed, leaving only debris cones below. Their investigations revealed that the principal agent of reworking was debris flow activity, triggered by both snowmelt at gully heads and autumnal rainstorms.

The close association of relict debris cones with ice-marginal moraines on valley sides in the Lake District is considered by the author to reflect paraglacial activity following withdrawal of the ice margin. Although a detailed investigation of paraglacial landforms and sediments has yet to be undertaken in the Lake District, observations within those areas studied reveal that the steepest slopes are associated with the best developed debris cones and the poorest preserved lateral moraines. For example, debris cones mantle the steep slopes of the narrow valley which links Greenup with Stonethwaite in the central Lake District, and are also widespread in the lower reaches of adjacent Langstrath. Nevertheless, evidence for former ice margins has not been completely removed by paraglacial activity at these locations and, with care, some ice margins can be reconstructed (Section 4.2.2.3). This is particularly well illustrated on the slopes north of Black Sail Youth Hostel in upper Ennerdale, where a moraine-debris cone complex has developed.

For palaeo-ice-fronts to be useful as a means of identifying former plateau icefields, two conditions must be met. Firstly, decay must be centred on the summit (at least until a very late stage in deglaciation). Investigations into the deglaciation of the Loch Lomond Stadial icefield in Northwest Scotland have revealed that corries were not important as local decay centres and decay was centred on the highest ground (Bennett, 1991; Bennett and Boulton, 1993a, 1993b). This also appears to apply for the Loch Lomond Stadial plateau icefields in the central fells of the Lake District. Secondly, active deglaciation must have occurred throughout (or at least in the final stages). Whilst this appears to have been the case for both the Northwest Highlands (where patterns of unordered stagnation terrain are relatively limited) and the valleys in the central Lake District, this apparently did not occur on Skye, where a climatically-driven two-stage deglaciation of the icefield is recorded in the geomorphological record (Benn *et al.*, 1992). Initial active retreat was followed by uninterrupted retreat and localised *in situ* stagnation. If this had occurred in the Lake District, then it would not be possible to trace lateral moraines onto the higher ground in the final stages of deglaciation and thus demonstrate a plateau icefield source.

Bennett and Boulton (1993b) have suggested that the difference in the styles of deglaciation between Skye and the Northwest Highlands may be due to the mainland icefield being sufficiently large to dampen the impact of climate forcing which drove the two-stage deglaciation in Skye. The Lake District evidence, although restricted in coverage, provides no evidence for the two-stage deglaciation recorded in Skye. Moreover, the ice masses were relatively small and thus would have had a minimal influence on the local climate. Nevertheless, it is probably premature at this stage to speculate on the significance of such differences given that most of the Lake District remains to be re-mapped.

6.4 TOPOCLIMATES AND PLATEAU ICEFIELDS

Where the geomorphological record proves to be equivocal, plateau icefield reconstruction will require an assessment of topographic and palaeoclimatic factors (Section 2.3). In his investigation into the topoclimatic controls on European icefields, Manley (1955) demonstrated that there is a close, non-linear relationship between summit breadth and altitude above the regional firn line (Figure 2.6). Essentially, as summit breadth decreases, the altitude above the firn line that a summit must attain if it is to support an icefield increases. Manley's curve is reproduced again here, this time showing data for the Loch Lomond Stadial plateau icefields which developed in the Lake District, with firn lines averaged at 500 m [Sections 4.3 and 5.3; (Sissons, 1980a)] (Figure 6.3).

High Raise, which undoubtedly has the clearest geomorphological evidence for a former plateau icefield (Chapter 4), lies on Manley's curve. The remaining summits which supported plateau icefields at this time all lie below the curve. For example, Thunacar Knott and Ullscarf are both approximately 70 m below the elevation considered by Manley (1955) to be necessary for plateau icefield development. The largest discrepancy is associated with Brandreth, which lies 150 m below the curve, although in this case the contiguous Grey Knotts summit creates a more extensive stretch of upland which was clearly favourable for plateau icefield development.

It is not clear to what extent Manley's curve (in conjunction with the data from this investigation) can be used to predict where plateau icefields developed elsewhere in the Lake District. Data from a number of areas in north Norway (Bergsfjord, Lyngen, Narvik, Skjomen) provide general support for Manley's (1955) suggestion that the existence of a small plateau icefield is controlled by the breadth of the summit in the direction of the prevailing wind during the accumulation season, and the height of the plateau above the local firn line (Rea *et al.*, in prep.). Nevertheless, Rea *et al.* argue that the simple relationship envisaged by Manley does not hold, with summits hosting plateau icefields at lower elevations than those suggested by the curve.

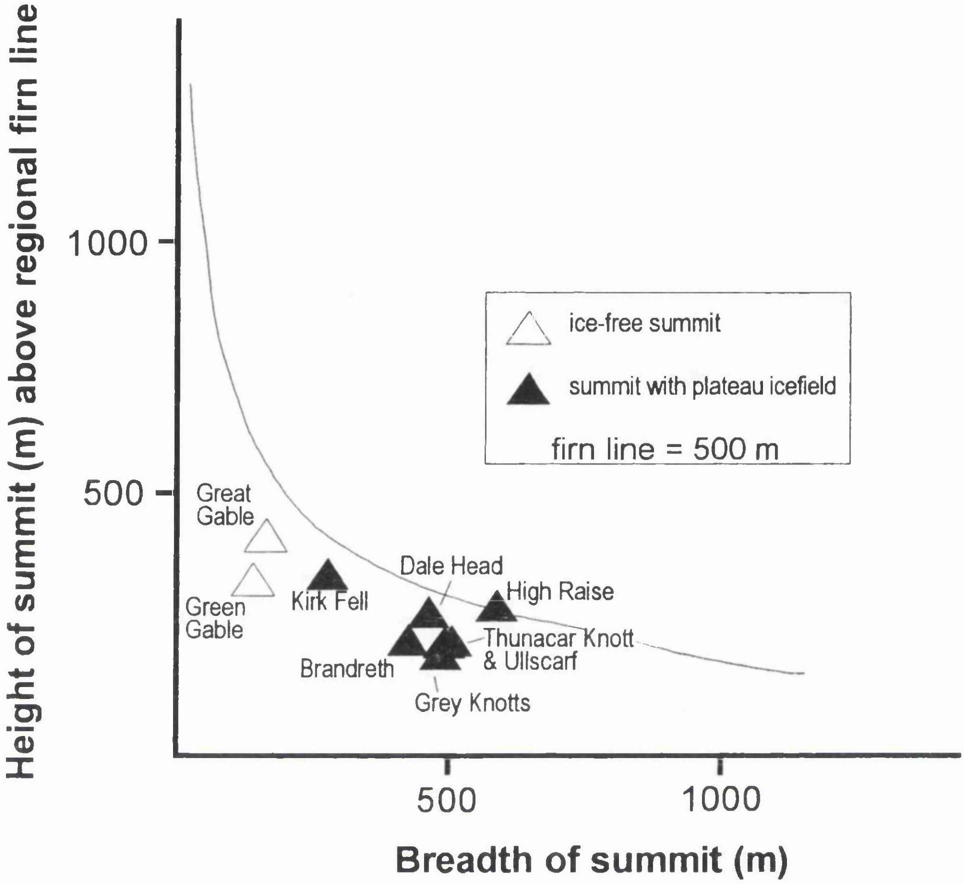


Figure 6.3 Manley's (1955) curve with additional Lake District data

The curve documents the relationship between summit breadth and altitude above the firn line for plateau icefield development (Section 2.3). The Lake District data are based on a Loch Lomond Stadial firn line of 500 m (Sections 4.3 and 5.3).

6.5 RECONSTRUCTING PLATEAU ICEFIELDS

This research shows how conventional approaches to palaeoglacier reconstructions can fail to take into account the former existence of relatively small plateau icefields, a situation which largely reflects the nature of the geomorphological record. Even where evidence for a former plateau icefield exists, there may be little or no geomorphological evidence to constrain its dimensions in the accumulation zone.

In this investigation, summit ice thicknesses are estimates based on the theoretical relationship between plateau icefield depth and summit breadth (Section 2.2.2), with general configurations extrapolated from reconstructed ice-marginal positions produced during deglaciation. The principal advantage of this approach is that it is simple to employ and has a glaciological basis, although the validity of the overall reconstruction is dependent upon the researcher's skill and judgement in contouring the former ice surface. This limitation could be addressed in future investigations by deriving surface contours from high-resolution, three-dimensional models.

Traditionally, three-dimensional models have not featured in Loch Lomond Stadial palaeoglacier reconstructions, although recent modelling by Hubbard (1997) of the ice cap which developed in the western Scottish Highlands at this time demonstrates their potential. Hubbard's model was constrained by the field data and used primarily to investigate hypotheses concerning temperature and precipitation changes during the Loch Lomond Stadial. However, it was also used to simulate the limits and geometry of the ice cap into areas where field evidence is currently lacking.

It is difficult to envisage that the discrepancies between these different approaches (i.e. three-dimensional modelling and the method employed in this investigation) would be sufficiently marked to have a significant bearing on ELAs. Nevertheless, it is clear from Hubbard's (1997) investigation that three-dimensional models could be usefully employed to explore the dynamics of individual plateau icefield systems. Similarly, glaciological modelling could also be used to check the validity of existing, geomorphologically-based reconstructions in order to check their validity [for example, those of Sissons (1980a)]. To this end, it may well be that outlet glaciers which have been incorrectly reconstructed as valley glaciers are glaciologically problematical.

Finally, glaciological modelling could assist in constraining the downvalley extents of plateau icefield outlet glaciers. Although the downvalley extents of some outlet glaciers in this investigation appear to be defined by relatively clear end moraines (for example, at Wythburn and Rosthwaite), the evidence elsewhere is generally subtle and fragmentary. Furthermore, all sites lack any form of dating control. Thus, distinguishing between Loch Lomond Stadial and Dimlington Stadial glaciogenic landform assemblages is problematical. Dating (or the lack of it) is arguably one of the most substantial obstacles to reconstructing former glaciers in upland Britain.

6.6 DATING PROBLEMS

A problem faced by many researchers reconstructing former glaciers in mountain environments is that of dating, particularly where it is necessary to distinguish between more than one event. Organics rarely survive reworking, so absolute dating control tends to be the exception rather than the rule. In upland Britain, the glaciogenic landform assemblages of the relatively restricted Loch Lomond Readvance need to be distinguished from those of the much more extensive Dimlington Stadial ice sheet. It has been argued that, unlike glaciogenic landforms associated with the decay of the Dimlington Stadial ice sheet, Loch Lomond Readvance landforms remain relatively unweathered and of a 'fresh' appearance (e.g. Lowe and Walker, 1984, p.27). Thus, most workers have employed moraine 'freshness' as a relative dating technique, although its use implies acceptance that Loch Lomond Readvance landforms should everywhere be morphologically similar and distinct from those produced during the decay of the last ice sheet.

Ignoring the problems of defining freshness and consistently applying it in the field, perhaps the attraction of freshness as a relative dating technique for many workers is that it is a 'quality' which can be visually assessed quite rapidly, and can be used on all glaciogenic landform assemblages. Although bio- and lithostratigraphic contrasts have been used at some localities to assist in constraining ice-marginal positions, these relatively time-consuming investigations have proved to be the exception rather than the rule (see Gray and Coxon, 1991, p.97). Furthermore, the spatial resolution of such investigations are usually quite poor because they are dependent upon the availability of suitable 'inside' and 'outside' sites.

Nevertheless, freshness is intuitively an unreliable relative dating technique. The notion that the downvalley extents of Loch Lomond Stadial palaeoglaciers should be everywhere defined by morphologically similar (fresh) moraines is difficult to sustain. It takes no account of the range of (interrelated) variables which determine moraine morphology. These variables include glacier morphology, ice margin behaviour during deglaciation and the effects of paraglacial reworking on both valley sides and floors (Section 3.3.4). The influence of these variables is obvious in contemporary glacial environments (Figures 6.4 and 6.5) (e.g. D.J.A. Evans, 1988; 1990; Ballantyne and Benn, 1994a; Maizels, 1995). For 'freshness' to be a valid age discriminant, all of these factors must have remained constant throughout upland Britain during the Loch Lomond Stadial and this was clearly not the case.

The potential pitfalls associated with defining former glacier margins on the basis of the areal distribution of 'fresh' glaciogenic landform assemblages are well illustrated in Stake Pass and Langdale Combe (Section 4.2), where both Manley (1959) and Sissons (1980a) incorrectly interpret the 'limit' of fresh moraines (in reality a transition to increasingly faint moraines) in an ice-marginal context (Section 4.4). Other examples are discussed in Chapters 4 and 5. Whilst it is undeniably the case that the most prominent glaciogenic landform assemblages in the valley heads of upland Britain relate to the Loch Lomond Readvance, it does not follow that all Loch Lomond Readvance landforms are similarly prominent.

Moraine 'freshness' was not employed in this investigation. Instead, landforms of different ages were differentiated on the basis of style of deglaciation. It is generally accepted that the last ice sheet stagnated within the Lake District (e.g. Boardman, 1981). If the margins of Loch Lomond Stadial glaciers actively backwasted, at least for a short distance, then this provides a means of differentiating between Dimlington Stadial and Loch Lomond Stadial glacigenic landforms which is independent of moraine morphology. Thus, the arcuate end moraines at Rosthwaite (Borrowdale) and those at Wythburn (Clark and Wilson, 1994) are not believed to be associated with the wastage of the last ice sheet. There are a number of limitations with this approach, the principal one being that the style of deglaciation of the last ice sheet has not been rigorously investigated. Furthermore, it also assumes that there was no advance of glaciers between the wastage of the last ice sheet and the renewed glaciation of the Loch Lomond Stadial. Finally, this approach will not work where localised stagnation of Loch Lomond Stadial glaciers occurred.

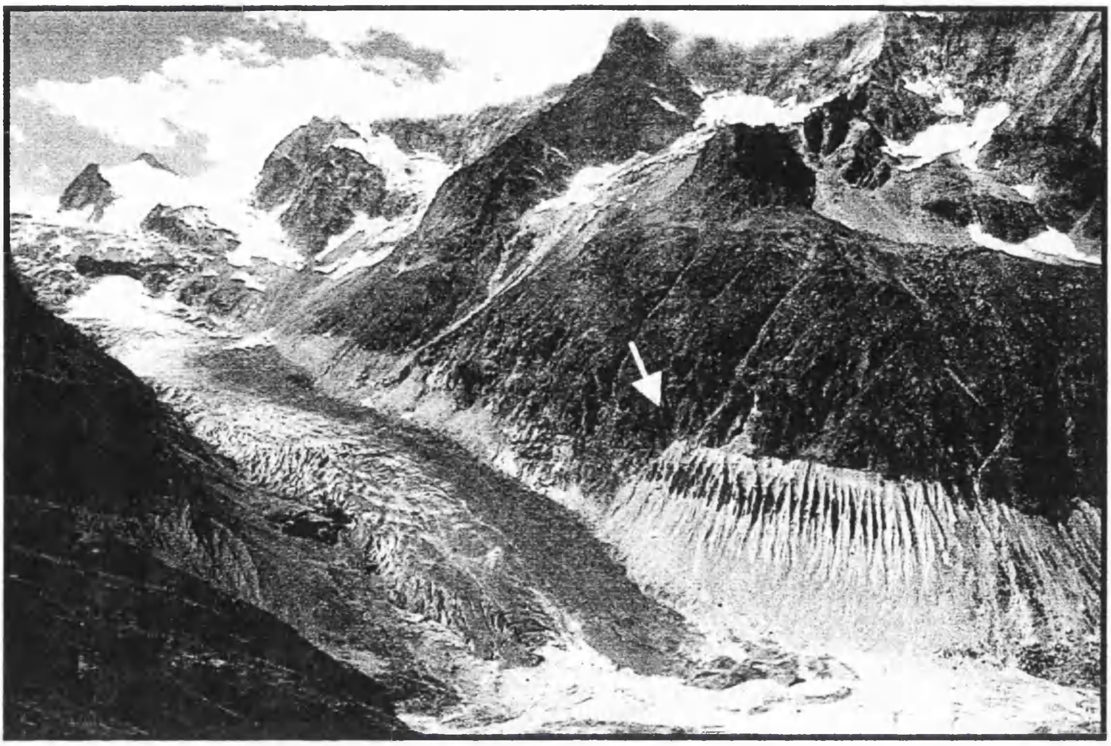


Figure 6.4 Paraglacial reworking on a Little Ice Age moraine in Val d'Hérens, Switzerland.

Mass wasting processes are evidenced by the gullies on the upper slopes of the moraine, with the material being redeposited as a series of coalescing debris cones on the lower slopes. A short section of the moraine has been removed altogether to reveal the underlying bedrock.

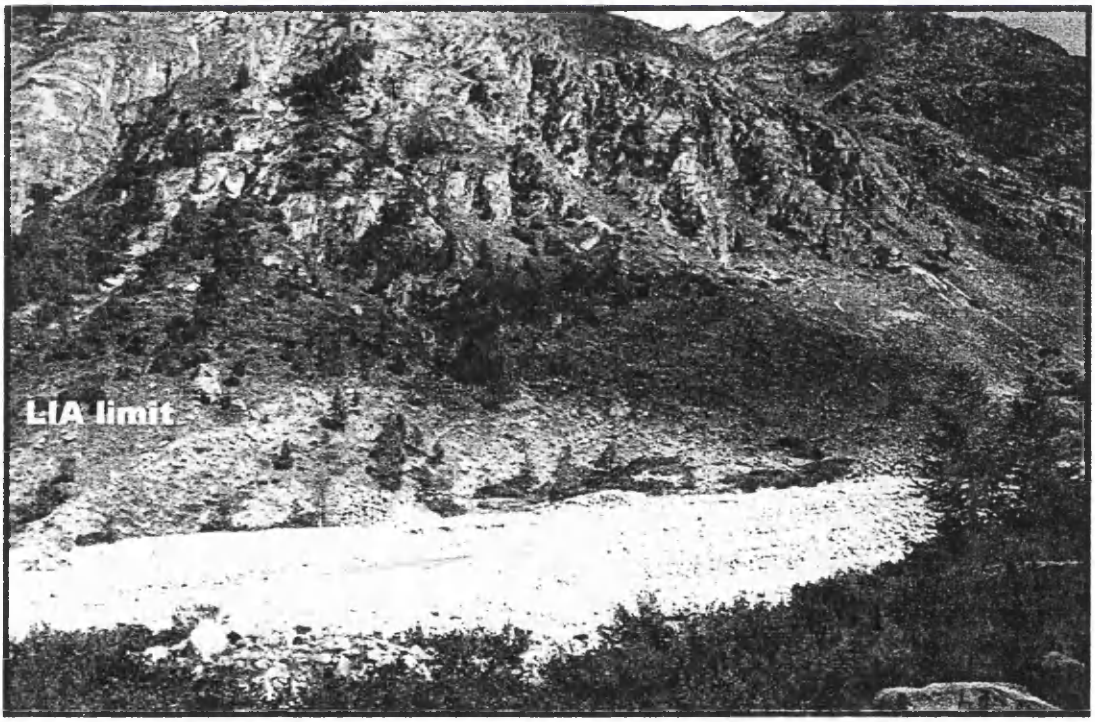


Figure 6.5 The Little Ice Age limit of the Arolla glacier, Val d'Hérens, Switzerland.

Although a clear drift limit on the valley side documents the Little Ice Age position of this glacier, paraglacial reworking on the valley floor has completely removed evidence for a former ice margin.

7

Conclusion

7.1 CONCLUSIONS

The main conclusions reached by this study are:

1. Detailed geomorphological mapping has revealed evidence for the development of plateau icefields in the central fells of the English Lake District during the Loch Lomond (Younger Dryas) Stadial.

The largest plateau icefield system, which covered an area of approximately 55 km^2 (including outlet glaciers), was centred on High Raise (NY281095). To the west, smaller plateau icefields developed on Grey Knotts/Brandreth, Dale Head and probably also on Kirk Fell, covering areas of 7 km^2, 3 km^2 and 1 km^2 respectively. This interpretation differs from those of previous workers, who assumed an alpine style of glaciation, with reconstructed glaciers emanating from corries and valley heads (Manley, 1959; Sissons, 1980a).

The geomorphological impact of these plateau icefields appears to have been minimal on the summits, where the survival of blockfields and other frost-weathered debris (which is mostly peat-covered) implies the existence of protective, cold-based ice. This interpretation assumes that the Loch Lomond Stadial was the last major episode of periglacial activity to have affected upland Britain (e.g. Ballantyne and Harris, 1994). As such, this represents the first reported occurrence of a Loch Lomond Stadial ice mass which was not wet-based throughout. Cold-based conditions would have been promoted by a combination of thin, slow-moving ice plus the influence of low mean annual air temperatures on the summits. Ice-moulded bedrock at some plateau edges, however, document a transition to wet-based, erosive conditions. At these locations, steeper slopes would have resulted in increased strain heating within the ice.

In many cases, prominent moraine systems were produced by outlet glaciers which descended into the surrounding valleys where their margins became sediment traps for supraglacial debris and inwash. In some valleys, ice-marginal moraines record successive positions of outlet glaciers which actively backwasted towards their plateau source. Given the virtual absence of periglacial trimlines in the study area, reconstructed palaeo-ice-margins constitute the single most important line of evidence in the identification of plateau icefields in the geomorphological record. Arguably the most impressive sequence of ice-marginal moraines occurs southwest of High Raise in Stake Pass and Langdale Combe. These hummocky recessional moraines, which have been palynologically dated to the Loch Lomond Stadial by Walker (1965), document in considerable detail the active retreat of a lobate ice margin as it backwasted towards the High Raise–Thunacar Knott summit area.

Given the virtual absence of ice-marginal control points in the upper reaches of these plateau icefield systems, their reconstructions are somewhat speculative. Ice thicknesses on the summits (40–50 m) are estimates based on the theoretical relationship between plateau icefield depth and summit breadth (Section 2.2.2). General ice mass configurations have been extrapolated from palaeo-ice-marginal positions produced during deglaciation. Support for these reconstructions is provided by the consistency in firn line values, all of which lie around 500 m OD. A Loch Lomond Stadial firn line of 500 m OD corresponds to a mean annual precipitation of 2,000–2,500 mm, assuming a mean July temperature of 9°C (e.g. Coope, 1994; Lowe *et al.*, 1995a) and a lapse rate of 0.58°C/100m (Section 6.2).

2. In the absence of periglacial trimlines, conventional approaches to reconstructing former glaciers do not take into account the subtle geomorphological impact of plateau icefields.

Delineating ice margins in former accumulation zones can be problematical because geomorphological evidence is invariably poorly developed or absent altogether. Many workers have attempted to circumvent this difficulty by extrapolating ice margins into accumulation zones. In the Lake District and other parts of upland Britain considered to be 'marginal' for glaciation during the Loch Lomond Stadial (Wales, the Southern Uplands and the Cairngorms), extrapolated ice margins have been based on the assumption of an alpine style of glaciation, with glaciers emanating from cirques and valley heads (e.g. Sissons, 1980a; Cornish, 1981) (Section 2.4.2.3). In doing so, no explicit consideration was given to the nature of the terrain, despite the fact that glaciation styles are determined by interactions between topography and climate (Section 2.3). This assumption of an alpine style of glaciation is certainly not appropriate for the study area, where plateau icefields developed on some of the more rounded summits.

The potential difficulties surrounding the identification of former plateau icefields in deglaciated areas were first discussed by Gordon *et al.* (1987, 1988) and Gellatly *et al.* (1988). Working in north Norway, they observed that the geomorphological impacts of some contemporary plateau icefields appeared to be negligible, with recession revealing areas which have experienced little or no subglacial erosion, a situation which has been attributed to low basal shear stresses and, in some places, cold-based ice (Gordon *et al.*, 1987, 1988; Gellatly *et al.*, 1988; Rea *et al.*, in prep.). They argued that if similar conditions prevailed in areas which are now deglaciated, such plateau icefields would be difficult or impossible to recognise in the geomorphological record. The significance of the present investigation is that it provides, for the first time, evidence from the British Isles which supports the Norway research.

In Britain, it is generally believed that there is a mutually exclusive relationship between the distribution of relict major periglacial landforms, such as blockfields, and the ground covered by Loch Lomond Stadial glaciers (e.g. Ballantyne, 1984; Ballantyne and Harris, 1994). As such, many researchers have employed contrasts in weathering and periglacial landform development to assist in delineating the higher reaches of Loch Lomond Readvance glaciers. However, this approach assumes wet-based conditions throughout, which does not appear to have been the case for the summits investigated. Rather, blockfields and other frost-weathered debris on these summits imply the former existence of protective, cold-based ice.

However, whilst providing broad support for the concerns of Gordon *et al.* (1987, 1988) and Gellatly *et al.* (1988) regarding the identification of former plateau icefields, this investigation has shown that in the gentler relief of the Lake District, the decay patterns produced by actively-backwasting outlet glaciers can demonstrate the former existence of a plateau icefield, even where the summit exhibits little or no evidence for subglacial erosion. Decay patterns have been reconstructed using the approach of Benn (1990) and Bennett (1991). However, such evidence is invariably subtle and fragmentary, particularly on the upper slopes. By itself, the absence of direct geomorphological evidence for a plateau icefield is inconclusive. Reconstruction of plateau icefields in such circumstances will require an assessment of topographic and palaeoclimatic factors.

The implications of this research are clear; former plateau icefield systems elsewhere may have been incorrectly reconstructed as cirque/valley glaciers adjacent to ice-free summits. This applies to any part of the world where assumptions about glaciation style have been made without due consideration to the nature of the terrain.

3. The failure to take account of former plateau icefields will result in erroneous palaeoclimatic inferences.

The failure to account for former plateau icefields is significant where they were vital for the maintenance of glaciers in the surrounding corries and valleys, and will result in an overestimation of equilibrium line altitude (ELA) lowering. In this investigation, the largest ELA adjustment affects the Honister Pass palaeoglacier. Whereas Sissons' reconstruction of this former ice mass produced an ELA of 385 m, the addition of a plateau icefield source increases this value to 475 m (Section 5.3). In many of the cases considered, however, the impact of accumulation at higher levels on the ELA is partially offset by more extensive downvalley positions. In general, ELA discrepancies will be most pronounced in high relief mountain environments with extensive plateaux, such as parts of Norway. For a given temperature, an increase in the altitude of the equilibrium line corresponds with a decrease in precipitation.

The plateau icefield reconstructions proposed in this investigation remove the need to invoke, as Sissons (1980a) did, marked snowfall variations over geographically limited areas or unusually high inputs of wind-blown snow in former accumulation zones (Section 2.4.4.5). Instead, variations in the extent of glacierization can be attributed to topoclimatic factors, specifically whether or not accumulation took place on the summits. Thus, relatively extensive outlet glaciers which drained plateau icefields are not inconsistent with much smaller ice masses which developed below narrow summits. It is likely that local variations in palaeoglacier development elsewhere in Britain and beyond may be explained more simply by summit accumulation rather than by invoking elaborate variations in snowfall intensity.

4. Detailed geomorphological mapping in valley-heads may reveal evidence for former plateau icefields.

At a number of sites in the central Lake District, it has been possible to reconstruct successive ice-marginal positions of outlet glaciers in the final stages of deglaciation and demonstrate that decay was centred on summits rather than valley heads. Ice-marginal moraines produced by actively-backwasting outlet glaciers have proved to be the single most important line of evidence in the search for former plateau icefields in the Lake District. A former plateau icefield may be inferred where the direction of deglaciation recorded by ice-marginal moraines differs from that which would be predicted for a valley glacier. For example, the suite of hummocky recessional moraines in Stake Pass and Langdale Combe define successive margins of an outlet glacier which actively-backwasted towards its source on High Raise, perpendicular to the valley axis (Chapter 4). A plateau icefield may also be inferred where ice-marginal moraines can be shown to extend beyond the topographic confines of the valley and onto the higher ground above. This is the most common basis by which former plateau icefields have been inferred in the Lake District. For palaeo-ice-fronts to be useful as a means of identifying former plateau icefields, decay must have been centred on the summit and active deglaciation must have occurred throughout.

5. Dating uncertainties remain a barrier to accurate palaeoglacier reconstructions in upland Britain.

It is often argued that, unlike glaciogenic landform assemblages associated with the decay of the last ice sheet, Loch Lomond Readvance landforms remain relatively unweathered and of a 'fresh' appearance (e.g. Lowe and Walker, 1984, p. 27). Thus, many workers have employed contrasts in moraine freshness to define the downvalley extents of Loch Lomond Stadial glaciers (e.g. Sissons, 1980a). Nevertheless, freshness is intuitively an unreliable relative dating technique. The notion that the downvalley extents of Loch Lomond Readvance glaciers should be everywhere defined by morphologically similar (fresh) moraines is difficult to sustain. It takes no account of the range of (interrelated) variables which determine moraine morphology. These variables include glacier morphology, ice margin behaviour during deglaciation and the effects of paraglacial reworking on both valley sides and floors (Section 3.3.4). For 'freshness' to be a valid age discriminant, all of these factors must have remained constant throughout upland Britain during the Loch Lomond Stadial and this was clearly not the case. Whilst it is true that the most prominent (or 'freshest') glaciogenic landform assemblages in the valley heads of upland Britain relate to the Loch Lomond Readvance, it does not follow that all Loch Lomond Readvance landforms are morphologically similar.

7.2 RECOMMENDATIONS FOR FURTHER RESEARCH

This study has only covered a relatively small part of the English Lake District and thus there are clear opportunities to extend this approach to the remainder of the region and beyond. Sites worthy of initial consideration within the Lake District would be highlighted by employing Manley's curve (plus the Lake District data) and assuming a firn line of 500 m, a figure which appears to be valid for the central fells at least. Given the limited size of the Lake District, it seems unlikely that the regional firn line would have varied markedly. Suitable summits may also be flagged through their association with any of Sissons' (1980a) reconstructed glaciers which appear anomalous in some way, perhaps due to low ELAs. Other parts of upland Britain may have proved even more conducive to their formation and thus it may be more productive to concentrate on those areas. Suitable candidates include the rounded summits of the Southern Uplands, the Cairngorms and the Monadliath Mountains. Clay weathering analysis of summit material may prove useful in identifying sediments which have survived numerous glaciations, although its role in distinguishing between Loch Lomond Stadial and Dimlington Stadial sediments is less certain due to the relatively small age differences involved.

In any such investigation, dating uncertainties and an incomplete geomorphological record may prove substantial barriers to plateau icefield reconstructions. It is possible that both issues could be at least partially addressed by employing high-resolution, three-dimensional models. These would be of particular value in reconstructing surface profiles in accumulation zones, where ice-marginal evidence may be poorly developed or (as in the case of the present study) absent altogether. Glaciological modelling could also assist in constraining the downvalley extents of plateau icefield outlet glaciers, where evidence may be fragmentary and/or dating control is unavailable.

Traditionally, glaciological modelling has not featured in Loch Lomond Stadial palaeoglacier reconstructions, although recent modelling by Hubbard (1997) of the ice cap which developed in the western Scottish Highlands at this time shows their potential. Apart from being used to simulate ice surfaces in areas where the geomorphological record is lacking, modelling could also be used to explore the dynamics of individual plateau icefield systems. Similarly, glaciological modelling could also be used to check the validity of existing, geomorphologically-based reconstructions in order to check their validity [for example, those of Sissons (1980a)].

Finally, it is important that investigations of contemporary plateau icefield systems continue. In particular, a detailed study of their dynamics will undoubtedly assist in the development of high resolution, three-dimensional models of former plateau icefield systems.

Appendix

Aerial photographs used in this investigation

Chapter 4: High Raise, Thunacar Knott and Ullscarf	
Ordnance Survey 1:7,000	(B&W)
Nos. 72/008/311, 312;	Mickleden; lateral moraines clearly visible
73/140/115	below Greenup; lateral moraines and paraglacial cones
M.A.F.F. 1:14,000	(B&W)
Nos. 158/33–158/41	contrasty, but minimal shadow; shows summit blockfields and related landforms clearly
159/251–159/259	as above
M.A.F.F. 1:18,000	(B&W)
Nos. 379/62–379/68	flutings by Stickle Tarn reasonably clear; moraines north of Ullscarf visible
379/101–379/107	Greenup moraines clear; pattern within Stake Pass moraines not visible due to adverse angle of illumination
379/187–379/189	eastern valleys
381/259–381/261	moraine/blockfield boundary at Rossett Pike clear
M.A.F.F. 1:20,000	(colour)
88/29/16–88/29/18	Rossett Pike and Head of Langstrath; moraine/blockfield boundary clear; ice marginal moraines in upper Langstrath very clear
88/32/08–88/32/12	Greenup moraines very clear; pattern within Stake Pass moraines not visible
88/33/34–88/33/40	flutings by Stickle Tarn clear
Ordnance Survey 1:23,000	(B&W)
72/262/384–72/262/387	eastern valleys, moraines generally clear (but those south of Easedale not); Wythburn moraine clear
72/262/415–72/262/422	moraines at Watendlath very clear; Greenup moraine structure not visible due to adverse angle of illumination; Stake Pass moraine structure very clear; flutings at Stickle Tarn visible
72/262/448–72/262/449	moraines in upper Langstrath difficult to discern, but clear drift limits on western slopes
Chapter 5: Grey Knotts, Brandreth and Dale Head	
M.A.F.F. 1:18,000	(B&W)
381/256–381/258	Honister Pass and Gillercomb
M.A.F.F. 1:20,000	(colour)
88/29/13–88/29/14	Honister Pass and Gillercomb; moraines reasonably clear
88/29/57	Moraines in upper Gillercomb very clear; Ennerdale moraines reasonably clear
Ordnance Survey 1:23,000	(B&W)
72/262/450–72/262/452	Honister Pass moraines clear; flutings in upper Gillercomb clear
72/342/275–72/342/277	Ennerdale moraines and debris cones clear

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