YOUNGER DRYAS MORAINES IN THE NW HIGHLANDS OF SCOTLAND : GENESIS, SIGNIFICANCE AND POTENTIAL MODERN ANALOGUES

Sven Lukas

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

2005

Younger Dryas moraines in the NW Highlands of Scotland: genesis, significance and potential modern analogues

Sven Lukas

Thesis submitted to the University of St Andrews
for the degree Doctor of Philosophy

School of Geography and Geosciences
University of St Andrews
September 2005

I, Sven Lukas, hereby certify that this thesis, which is approximately 58,000 words in length, has been written by me, that it is the record of work carried out by me and that it has not been submitted in any previous application for a higher degree.

Date 09.09.2005 Signature of candidate [redacted]

I was admitted as a research student in September 2002 and as a candidate for the degree of PhD in September 2003; the higher study for which this is a record was carried out in the University of St Andrews between 2002 and 2005.

Date 09.09.2005 Signature of candidate [redacted]

I hereby certify that the candidate has fulfilled the conditions of the Resolution and Regulations appropriate for the degree of PhD in the University of St Andrews and that the candidate is qualified to submit this thesis in application for that degree.

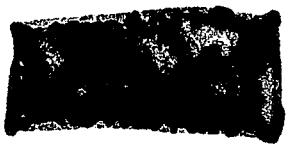
Date 09.09.2005 Signature of supervisor

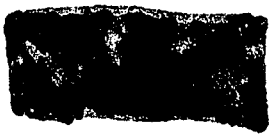
[redacted]

Unrestricted

In submitting this thesis to the University of St Andrews I understand that I am giving permission for it to be made available for use in accordance with the regulations of the University Library for the time being in force, subject to any copyright vested in the work not being affected thereby. I also understand that the title and abstract will be published, and that a copy of the work may be made and supplied to any bona fide library or research worker.

Date 09.09.2005 Signature of candidate [redacted]





Abstract

The Younger Dryas was the last period during which glaciers shaped large parts of the Scottish landscape. Reconstructing the palaeoclimate and glacial processes that operated during this time is crucial for the understanding of past atmosphere-cryosphere interactions and predicting future climate change.

This thesis presents results from geomorphological and geological mapping in the NW Highlands of Scotland that have resulted in the reconstruction of a Younger Dryas ice cap. Reconstruction of equilibrium-line altitudes and palaeo-precipitation values suggest that the Scottish west coast was wetter than at present.

Detailed sedimentological analyses of "hummocky moraines" allow the modes of moraine formation to be reconstructed in great detail and existing models to be tested. "Hummocky moraines" largely represent terrestrial ice-contact fans consisting of supraglacial debris flows and intercalated glaciofluvial units indicating an ice-marginal mode of formation. Different stages of deformation in these fans indicate highly dynamic glaciers that oscillated during retreat, partly or completely overriding previously formed landforms during readvances. Clast shape analyses reveal that debris was mostly subglacially derived and transported. The evidence is incompatible with a morphological model according to which the moraines could be formed by englacial thrusting.

Comparison with modern glacial landsystems indicates the following similarities with Scottish Younger Dryas glaciers. Low winter temperatures are similar to those on Svalbard, the marginal response of Younger Dryas glaciers to temperate environments and the modes of deposition to less responsive debris-covered glaciers. High precipitation along the Scottish west coast probably suppressed continuous permafrost development and caused high mass turnover and very dynamic, dominantly temperate Younger Dryas glaciers. Only a narrow zone around the margins appears to have been frozen to the ground, aiding elevation of basal debris and rapid deposition near the snout. The specific climatic and glaciological conditions during the Younger Dryas appear not to have a single modern analogue.

Acknowledgements

This thesis would never have taken its present shape without the input and support from the following individuals.

First and foremost I am grateful to my principal supervisor Dr. Doug Benn for the numerous enthusiastic and motivating discussions, his continuous encouragement and assistance in the field. Without his openness and willingness to explore different paths and to take risks, this thesis would have had a different focus. Furthermore, I would like to thank Prof. Colin Ballantyne for his input in the early stages of my PhD work. Dr. Tom Bradwell is thanked for providing invaluable input to the project at different stages, but also useful guidance and organisational assistance at the British Geological Survey (BGS) front.

Guidance on OSL-related matters was kindly given by Dr. Joel Spencer, Dr. Ruth Robinson (University of St Andrews), Dr. Frank Preusser (University of Bern, Switzerland) and Prof. Andreas Lang (University of Liverpool), and even though this thread of work did not fully come into effect, it was a useful and much appreciated learning process. Likewise, I thank Prof. Graeme Whittington (University of St Andrews) for his help in identifying plant macrofossils and his advice on sampling for ^{14}C-dating.

I have greatly benefited from discussions with the following individuals who have, in one way or another, had an impact on this thesis and my thinking about glacial and periglacial processes and environments or the NW Highlands: Dr. Jostein Bakke (University of Bergen, Norway), Dr. Hanne Christiansen (UNIS, Svalbard), Dr. Jez Everest, Nick Golledge, Jon Merritt (BGS), Prof. Ole Humlum (University of Oslo, Norway), David Jarman (Stirling), Prof. Reinhard Lampe (University of Greifswald, Germany) and the participants of the Quaternary Geology fieldtrip to the NW Highlands in May 2005 and Prof. Christian Schlüchter (University of Bern, Switzerland).

I am particularly grateful to Benny Reinardy (coring, surveying, digging), Tim d'Arville, Stefan Morrocco, Lindsey Nicholson (coring, surveying) and Fionna Ross (Norway, Svalbard) for the assistance with strenuous tasks and their company during fieldwork in the harshest weather conditions. I am indebted to Graeme Sandeman (University of St Andrews) for sharing his cartographic expertise and his comments on, and the printing of, Appendix 2.

My (partly ex-) postgraduate colleagues and friends at the School of Geography and Geosciences, most notably Dave Cheer, Laura Foster, Dr. Eleanor Haresign, Ruth Mottram, Dr. Lindsey Nicholson, Stefan Morrocco, Fionna Ross and Dr. Lindsay Wilson, made my stay even more worthwhile and kept me going during the more difficult phases of the project.

This work was carried out while I was in receipt of a UCAC studentship (2K01E22) which is hereby gratefully acknowledged. The fieldwork over the past three years was funded entirely by the Moine Thrust Project of the BGS, and I would like to thank Dr. Maarten Krabbendam in particular for his willingness to pay for two coring trips and ^{14}C dates. Further financial support for all the "extra"

work outwith the original scope of the PhD came from the School of Geography and Geosciences (Norway, Svalbard), the Bill Bishop Memorial Trust, the British Geomorphological Research Group and the Dudley Stamp Memorial Fund (Svalbard) as well as the Russell Trust, University of St Andrews (OSL-dating). Various smaller contributions from various funding bodies that supported trips to conferences and field meetings allowed me to present my findings early on.

Last but not least I am eternally grateful to my wife Tina for being there and for enduring all the ups and downs a PhD comes with. Her understanding and support, which includes fieldwork, entering clast shape data, digitising the bulk of moraines in Appendix 2 and checking the bibliography, contributed a great deal to this thesis.

TABLE OF CONTENTS

CHAPTER 1 INTRODUCTION

1.1 Background and state of research

Quaternary research in Scotland has a long tradition, and the clarity of some Scottish glacial landforms contributed significantly to the widespread acceptance of the glacial theory (e.g. Agassiz, 1840; Forbes, 1846; Bonney, 1871). Due to the repeated net erosion by successive ice sheets, evidence of earlier glaciations has largely been removed, however, and it is particularly in areas covered by glaciers during the Loch Lomond Stadial or Younger Dryas (ca. 12.7-11.5 cal ka BP), the last period of mountain glaciation in Scotland, that glacial depositional landforms are best-developed (cf. Sissons, 1967, 1976; Sutherland, 1984a; Ballantyne and Harris, 1994; Benn, 1997a, b). A landform assemblage found widely in areas covered by Younger Dryas glaciers is "hummocky moraine", so-called because of its apparent hummocky appearance when viewed from the ground (cf. Sissons, 1967). In this context it is worth pointing out that Scottish "hummocky moraine" is different from landform assemblages in other parts of the world that carry the same name (e.g. Benn and Evans, 1998; Eyles *et al.*, 1999; Dyke and Savelle, 2000; Dyke and Evans, 2003; Johnson and Clayton, 2003). Scottish "hummocky moraines" have rarely been subject to detailed investigations and despite advances made in the understanding of modern glacier environments and the availability of sedimentological methods in the past 20 years or so the only detailed sedimentological study remains that of Benn (1990, 1992a) on moraines on the Isle of Skye. More recent genetic interpretations, which attribute some "hummocky moraines" in Scotland to englacial and proglacial thrusting observed at polythermal glaciers in Svalbard (e.g. Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998; Graham and Midgley, 2000a), are almost entirely based on geomorphological evidence. Although this model has not been tested against sedimentological field evidence, it would, if true, have large implications for palaeo-environmental reconstructions. This thesis seeks to close this gap. It aims to (a) reconstruct the detailed modes of formation of Scottish "hummocky moraines", (b) discuss the implications of the findings in the context of glacier dynamics, (c) evaluate potential modern analogues and (d) reconstruct and assess the palaeoglaciological and palaeoclimatic significance of Younger Dryas glaciers in a large area in the far NW Highlands of Scotland that has so far received little attention from glacial scientists. In order to achieve these aims, geomorphological field mapping, detailed sedimentological analyses and comparison with two modern glacial analogues will be employed.

This chapter will review literature on "hummocky moraine" in Scotland to provide the context for the present study. It will focus on two major themes that have emerged in the literature, namely (1) the significance of Scottish "hummocky moraine" as an indicator of Younger Dryas glacier limits and (2) genetic models of Scottish "hummocky moraine" formation and their significance for glacier dynamics.

1.1.1 The role of "hummocky moraine" in Scottish Younger Dryas research

1.1.1.1 History and significance of Younger Dryas glaciation in Scotland

Early attempts to reconstruct the retreat patterns of the last ice sheet in Scotland (e.g. Charlesworth, 1926, 1955; Simpson, 1933) invoked numerous possible readvances that interrupted overall retreat. The advent of radiocarbon dating and detailed geomorphological mapping during the 1960s, however, led to the abandonment of most of the readvances proposed earlier, and by 1976 only the Loch Lomond Readvance, which corresponds to the global Younger Dryas, had been widely accepted (Sissons, 1976: 85; Ballantyne and Gray, 1984). This was later supplemented by evidence for a regional ice sheet readvance along the west coast of Scotland, the Wester Ross Readvance (Robinson and Ballantyne, 1979).

Research into the extent and duration of the Younger Dryas in Scotland has been enormous in the past 30 years and stems from the notion that this period was the last significant glaciation to affect Scotland on a large scale following ice sheet deglaciation (cf. Sissons, 1976, 1979a; Gray, 1997; Benn, 1997a, b). The current knowledge of ice extent during this period is shown in Fig. 1.1. Although persistence of glaciers into the early Holocene in high-level corries has been discussed on the basis of palynological work or geomorphological relationships (Kelletat, 1970; Walther, 1984; Rapson, 1985; Tipping *et al.*, 2003), no substantial evidence has so far been presented, and the widely accepted view is that glaciers had vanished shortly after termination of the Younger Dryas (Ballantyne, 1991, 2002a; Benn *et al.*, 1992; Benn, 1997a; Gray, 1997; Clapperton, 1997). Due to this special status of the Younger Dryas, which is sandwiched between ice sheet deglaciation and early Holocene warming, there is a need to understand palaeoclimatic conditions during this period, explaining the large amount of research conducted with respects to the Younger Dryas in Scotland (e.g. Sissons, 1974, 1979a, 1980; Sissons and Sutherland, 1976; Ballantyne, 1989, 2002a; Gray and Coxon, 1991; Bennett and Boulton, 1993a, b; Benn and Ballantyne, 2005).

In addition to a local or regional interest into the extent and dynamics of Younger Dryas glaciers in Scotland and Britain, this period is of global research interest as it represents the last short-lived return to cold conditions at the transition from full-glacial to full-interglacial conditions (e.g. Anderson, 1997; Rahmstorf, 2002; Tarasov and Peltier, 2005). The occurrence of this event emphasises the rapidity of climate change and is thought to have been triggered by instabilities in the thermohaline circulation of the North Atlantic, probably brought about by the sudden discharge of glacial meltwater from the Laurentide Ice Sheet (Tarasov and Peltier, 2005, and references therein). The moraines in Scotland formed during the Younger Dryas are potentially very valuable terrestrial archives, as they represent a case where glacier dynamics of relatively small palaeo-ice caps can be studied on land and their response linked to palaeoclimate proxies (e.g. Brooks and Birks, 2000; Ballantyne, 2002a; Benn and Ballantyne, 2005).

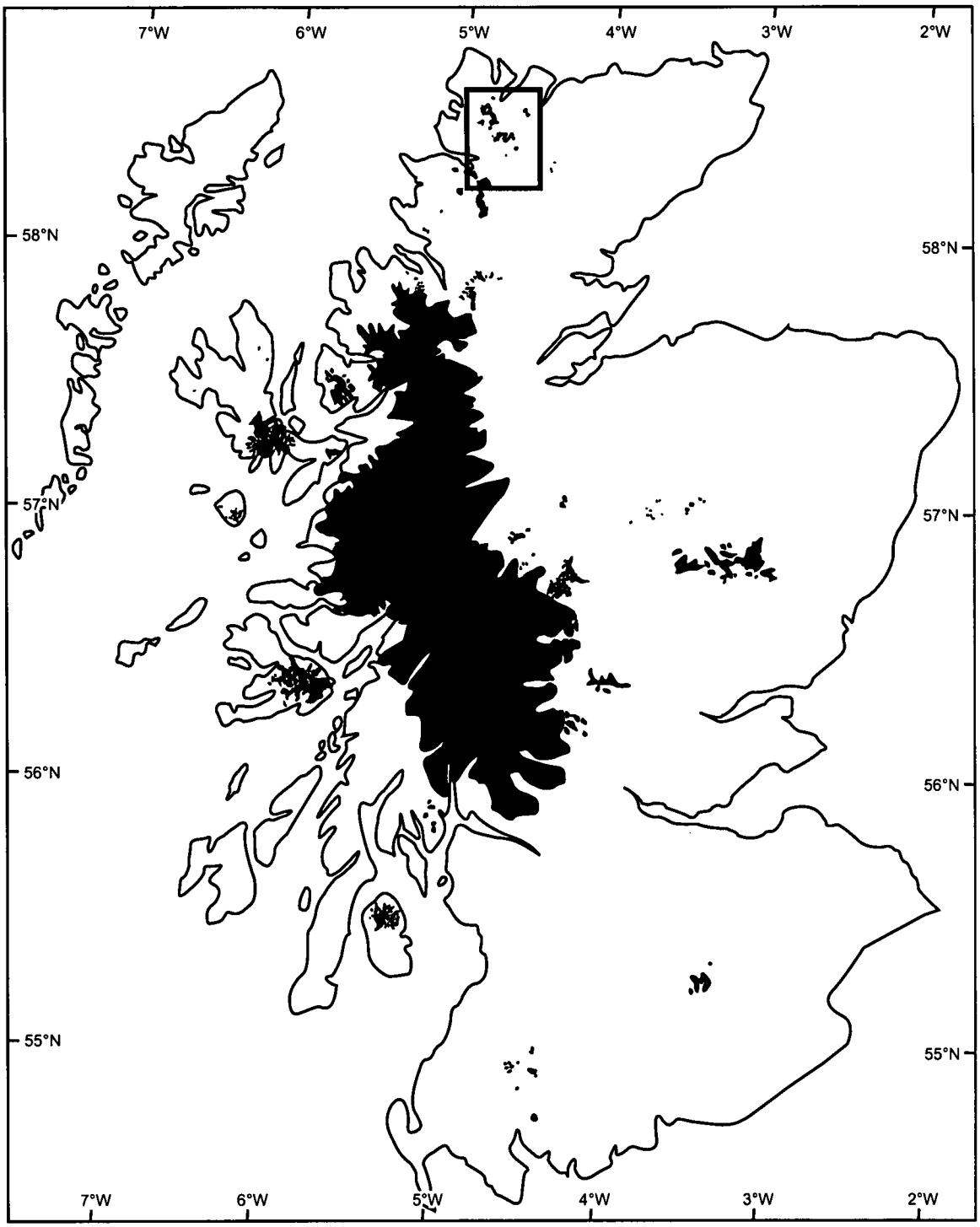


Fig. 1.1 The present knowledge of glacier extent during the Younger Dryas in Scotland (compiled from: Sugden, 1970; Sissons and Grant, 1972; Sissons, 1972, 1977, 1980; Gemmell, 1973; Ballantyne and Wain-Hobson, 1980; Lawson, 1986; Thorp, 1986; Gray and Coxon, 1991; Benn *et al.*, 1992; Bennett and Boulton, 1993b; Ballantyne and Benn, 1994; Ballantyne, 2002a; Benn and Ballantyne, 2005). The black frame marks the location of the present study area.

1.1.1.2 Dating the Younger Dryas in Scotland – the significance of "hummocky moraine"

Younger Dryas research in Scotland (and elsewhere in Europe) saw its beginnings through the use of palaeoecological, in particular palynological, studies in the 1950s through to the present day (cf.

Tipping, 2003). This work has shown a distinct pollen zone in which arboreal species were much reduced and arctic mosses, shrubs and sedges dominated the pollen spectrum between ca. 10.8 to 10.3 ^{14}C ka BP; this period was termed the Loch Lomond Stadial and linked to the European Younger Dryas (cf. Sissons and Walker, 1974; Walker, 1975a, b; Sissons, 1976; Vasari, 1977; Walker *et al.*, 1988; Gray and Coxon, 1991). Following Charlesworth's (1955) attempt to reconstruct lateglacial ice retreat which was only assigned speculative ages, J.B. Sissons and his co-workers were the first to recognise that glaciation during the Younger Dryas was an important and separate significant event that shaped the Scottish landscape through the renewed growth of glaciers following the Windermere Interstadial (e.g. Sissons, 1967, 1974, 1979a; Sissons and Sutherland, 1976). Through a combined program of radiocarbon dating, palynological analyses of kettle hole infills and geomorphological research, Sissons and his co-workers set out to reconstruct the extent of Younger Dryas glaciers in Scotland, a task in which "hummocky moraine" had an important place as delineator, although the understanding of what "hummocky moraine" represented and how it formed was very limited and not backed up by sedimentological work.

Sissons and his co-workers mapped *inter alia* the distribution of "hummocky moraine" throughout Scotland to determine the maximum extent and distribution of glaciated terrain during the Younger Dryas and calculate palaeoclimatic variables from these data (e.g. Sissons, 1967, 1974, 1976, 1979a; Sissons and Sutherland, 1976; Lawson, 1983, 1986). The age of this event was only constrained in a few places by radiocarbon dates, but it was argued that "hummocky moraine" was a characteristic landform assemblage formed during the Younger Dryas (cf. Sissons, 1976; Lowe and Walker, 1997). The inference of a Younger Dryas age based purely on the morphology and occurrence of "fresh hummocky moraine" and its use for glacier reconstruction was strongly opposed by Sugden (1970, 1974, 1980), Clapperton and Sugden (1977) and Clapperton *et al.* (1975) who stated that "hummocky moraine" can be produced in different topographic settings where, for example, parts of a retreating ice sheet were cut off by bedrock thresholds, stagnated and melted out *in situ*. Clapperton *et al.* (1975) were the first to show that "hummocky moraine" also exists in areas outside the Younger Dryas limits in the northern Cairngorms, confirming earlier morphological interpretations that "hummocky moraine" was also formed as the last ice sheet retreated (e.g. Sugden, 1970; Clapperton and Sugden, 1977). These authors argued that the presence of "hummocky moraine" does therefore not necessarily imply certain climatic conditions characteristic of only one chronological period, in this case the Younger Dryas, and that unless the origin and age of these moraines could be determined more accurately, such moraines should not *per se* be used in reconstructions of Younger Dryas glaciers.

The debate over whether "hummocky moraine" represents a time-marker continues, though to a more limited extent, to the present day. Wilson and Evans (2000) have shown that "hummocky moraines" that were produced during ice sheet retreat were overprinted by flutes during the Younger

Dryas to produce a palimpsest landscape (cf. Kleman, 1992). Furthermore, Everest (2003) and Everest and Golledge (2004), using cosmogenic radionuclide and optically-stimulated luminescence dating, have demonstrated that classical "hummocky moraines" in Glen Geusachan, Cairngorms that were previously attributed to the Younger Dryas (Sissons, 1979a; Bennett and Glasser, 1991; Bennett, 1996) pre-date the Younger Dryas by several millennia. Thus, using "hummocky moraine" on its own to infer a Younger Dryas age is problematic, and a new approach involving the use of more than one line of geomorphological evidence and numerical dating techniques is needed in future reconstructions of Younger Dryas glacier extent.

1.1.2 "Hummocky moraine" – changing genetic interpretations

At the beginning of glacial studies in Scotland, moraines were consistently interpreted as latero-frontal moraines and primarily used to delineate the extent of former ice masses, both in the earliest studies (e.g. Agassiz, 1840; Forbes, 1846; Bonney, 1871) and subsequent work carried out by officers of the British Geological Survey (e.g. Barrow *et al.*, 1913; Read *et al.*, 1926; Read, 1931). This early work was informed by modern analogues from the European Alps, and the moraines were interpreted as the product of "active" retreat of valley glaciers (cf. Benn, 1992a; Bennett, 1994). Some local assemblages of moraine, however, were interpreted to be the result of glacier stagnation (Harker, 1901). In later studies, Charlesworth (1955: 777) also interpreted the moraines, especially those "along the lower parts of the glens and straths and on the lower parts of the hillsides where the topography is mostly moundy" as "terminal and lateral features" indicative of active recession. In a pioneering effort to reconstruct the pattern of deglaciation across the Scottish Highlands he linked consecutive moraines to former ice-marginal positions which he attempted to correlate. Despite the problems he acknowledged as inherent in his approach, Charlesworth (1955) distinguished two phases of readvance that interrupted overall deglaciation, an earlier "Highland Glaciation" and a later "Moraine Glaciation" (Charlesworth, 1955: 904ff.). However, the largest problem associated with his work is that he never presented detailed geomorphological maps showing the distribution of moraines, so that his reconstructions could not be readily tested against field evidence, a point for which he earned a lot of criticism (cf. Sissons, 1967, 1974).

Contrasting with most previous work, J.B. Sissons and his co-workers described large tracts of moraines formerly seen as closely-spaced, distinct ridges and mounds as a "sea of chaotic mounds lacking any systematic arrangement" (Sissons, 1967: 97). These moraines were now commonly referred to as "hummocky moraine" (e.g. Sissons, 1967, 1974, 1976, 1979a). Based on their apparent chaotic appearance, their association with sites dated to the Younger Dryas (Loch Lomond Stadial, ca. 12.7-11.5 cal ka BP) and emerging palaeoclimatic data that suggested a very abrupt warming at the end of that period (Atkinson *et al.*, 1987), they were interpreted as a result of widespread *in situ* glacier stagnation (e.g. Sissons, 1967, 1974, 1979a; Sissons and Sutherland, 1976). Contrasting with

earlier work, this interpretation drew parallels with contemporaneous studies from areas covered by the Pleistocene Laurentide ice sheet that viewed "hummocky moraine" as the product of widespread *in situ* stagnation (e.g. Gravenor and Kupsch, 1959). Although the presence of linear elements within "hummocky moraine" was generally acknowledged, these were interpreted as basal crevasse-squeeze-ridges or flutes (e.g. Sissons, 1967, 1976).

As a result of these contrasting interpretations, case studies were subsequently conducted to gain a better understanding of the variety of processes that might lead to "hummocky moraine" formation. Hodgson (1982, 1986) studied shallow surface exposures and undertook clast fabric analysis in linear ridges with a downvalley orientation in Glen Torridon. He demonstrated that the material contained within these lineations was of subglacial origin, that the ridges were orientated parallel to ice flow direction and interpreted them as flutes that were formed underneath an active glacier. Eyles (1983a), using Icelandic glaciers for comparison, concluded that "hummocky moraine" might reflect the product of controlled and uncontrolled deposition of supraglacial debris. This model deviates from the rapid and widespread stagnation model adduced by Sissons and co-workers in that zones of stagnating ice can be cut off in front of actively, incrementally retreating glacier margins (cf. Kjær and Krüger, 2001; Evans, 2003a). Crucially, however, Eyles (1983a) did not provide sedimentological data from Scotland and thus, "hummocky moraine" was not largely accepted as a product of "active" incremental retreat.

Detailed geomorphological mapping and sedimentological analysis carried out by Benn (1990, 1992a) revealed that "hummocky moraine" consists of individual moraine ridges and mounds that are commonly aligned in individual chains. The moraines were composed of supraglacial debris flow units and intercalated fluvial units that contained signs of ice-marginal glaciotectonism. Together, this evidence demonstrated an ice-marginal formation of the transverse moraine fragments during oscillatory retreat. Bennett (1990), Bennett and Glasser (1991) and Bennett and Boulton (1993a) found similar geomorphological evidence when mapping moraines from aerial photographs; in particular they argued that the clear order recognised in areas of "hummocky moraine" and the association with ice-marginal landforms such as meltwater channels, outwash fans or terraces and kames could be used as evidence of an ice-marginal origin. These latter studies, however, were not backed up by detailed sedimentological evidence, although shallow surface exposures apparently also indicated ice-marginal sedimentation in a similar style to that reported by Benn (1992a). So far, the following sedimentary processes leading to the formation of Scottish ice-marginal ("hummocky") moraines have been recognised and generally accepted: (a) sedimentation along stationary ice margins leading to the formation of terrestrial ice-contact fans (Benn, 1990, 1992a; Mitchell and Lukas, 2004); (b) ice-marginal pushing of glacigenic material to form a push moraine (Benn, 1990, 1992a; Bennett and Boulton, 1993a); (c) dumping of supraglacial and englacial material following *in situ* ice stagnation forming unorganised, "chaotic" moraines (Eyles, 1983a; Benn, 1990, 1992a); (d) subglacial moulding,

i.e. formation of flutes, megaflutes and drumlins (Hodgson, 1982, 1986; Benn, 1992a; Bennett, 1995; Wilson and Evans, 2000). Actively formed recessional moraines (the majority of Scottish "hummocky moraine") have been used to reconstruct the retreat patterns of glaciers across Scotland in many studies since (Benn, 1990, 1992a; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; Lukas, 2003, 2004a, b).

More recently it has been suggested that moraines with clear proximal rectilinear slopes might have been formed at the margins of polythermal glaciers such as those that exist in Svalbard today. The earliest published suggestion that this model might be applicable to Scottish "hummocky moraine" stems from Bennett (1996) who used the morphological similarity of moraines in Glen Dee, Cairngorms with those observed in glacier forelands of high-arctic Svalbard glaciers (Hambrey and Huddart, 1995) to suggest that debris transfer and elevation of material along englacial thrust planes could be invoked for their formation. According to this model, englacial thrust planes form at the transition from warm- to cold-based ice near the snout as a result of flow compression. Likewise, flow compression is inferred to be induced by ice advancing up a reverse bedrock slope (e.g. Bennett, 1996; Graham and Midgley, 2000a; Midgley, 2001). Basal sediment is apparently elevated along these thrust planes into an englacial position where it forms englacial debris septa (Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Hambrey and Glasser, 2002; Glasser and Hambrey, 2003). During glacier retreat, this material is inferred to melt out without alteration to leave a sequence of stacked moraines with characteristic proximal rectilinear slopes (Bennett *et al.,* 1998). Stress propagation into the foreland induces stacking of proglacial material to produce a series of moraines. Sedimentological evidence in support of this model is restricted to shallow surface exposures, but it was stated that "each mound is composed of a single facies or facies association" (Hambrey *et al.,* 1997: 624; Bennett *et al.,* 1998: 19). This evidence was then used to infer the transfer of distinct facies in slices along thrusts, usually depicted by different ornaments in figures published by the thrusting group (cf. Bennett *et al.*, 1998).

Subsequently, the englacial thrusting model has been applied to two other sites in Britain, Coire a' Cheud-chnoic (Hambrey *et al.*, 1997; Bennett *et al.*, 1998) and Cwm Idwal (Graham and Midgley, 2000a). Graham (2002) attempted to test the applicability of the englacial thrusting model at six other sites across upland Britain, but his data remain equivocal due to the lack of detailed sedimentological work. The main line of evidence adduced by proponents of the thrusting model is the morphological similarity of moraines, notably the presence of proximal rectilinear slopes dipping upglacier at around 30°, and hints of a sedimentological similarity of the debris contained within such moraines. However, the sedimentological context in which the debris is found – e.g. facies associations, contacts between units, sedimentary and deformation structures and structural data – is consistently ignored.

This recent proposal, however, contradicts earlier interpretations and has opened up new problems. If it holds true, then several moraines could be formed in a single event, with far-reaching

implications for the interpretation of Scottish moraines that had thus far been interpreted as recessional moraines. As such the latter carry some information about the direction and nature of ice retreat (Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; Lukas, 2003). The proponents of the englacial thrusting model have shown that debris within the moraines observed in Svalbard (e.g. Hambrey and Glasser, 2003, Hubbard *et al.*, 2004), and also in Britain (Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002), is of subglacial origin. However, the link between a subglacial origin and the mode of deposition is far from clear, and a subglacial transport path of clasts does not necessarily imply deposition in a subglacial environment. The uncertainties associated with the thrusting model are mainly due to the lack of clearly documented sedimentological logs through stacked moraine sequences and present the weakest point in the argument for the englacial thrusting model.

In addition to these inherent uncertainties, recent re-evaluation of the structures interpreted as thrusts at Kongsvegen, Svalbard, has led to the conclusion that the role of thrusting may have been overestimated (Woodward *et al.,* 2002, 2003). As a consequence, Woodward *et al.* (2002: 207) conclude that the model of englacial thrusting as an explanation of British "hummocky moraine" formed during the Younger Dryas "also must be questioned". Similarly, the landforms in Glen Torridon interpreted as a result of thrusting by Hambrey *et al.* (1997) and Bennett *et al.* (1998) have been more convincingly re-interpreted as a palimpsest landscape in which older moraines have been overprinted by flutes during the Younger Dryas (Wilson and Evans, 2000).

1.1.3 Summary and nature of the problem

1.1.3.1 Genesis of "hummocky moraine"

The Scottish landform assemblage termed "hummocky moraine" has attracted a considerable amount of research in the past 50 years and resulted in conflicting interpretations with different implications for the palaeoclimate and palaeoglaciological conditions at the time of formation. The vast majority of studies dealing with Scottish "hummocky moraine", however, have to this day been merely guided by geomorphological interpretations and have thus not been able to address how this landform assemblage forms.

Despite a large amount of research into the nature of moraines at modern ice margins, few detailed studies exist that constrain palaeo-glacier dynamics effectively (e.g. Benn, 1990, 1992a; Benn and Evans, 1993, 1996) due to the lack of exposures or detailed sedimentological work (Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; Lukas, 2003). As a result, it is still not known by what processes moraines were formed during the Younger Dryas in most parts of Scotland and what the implications for palaeoglaciological and palaeoclimatic parameters are.

1.1.3.2 Models of "hummocky moraine" formation and modern analogues

The work carried out on "hummocky moraine" so far implies two different landsystems to explain the formation of moraines. In the first model, active incremental retreat is responsible for the formation of successive ice-marginal, recessional moraines which were formed at the glacier margin; moraines formed in such a way have been used to reconstruct patterns of palaeo-glacier retreat (e.g. Benn, 1990, 1992a; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003, 2004a, b).

Eyles (1983a) argued that Icelandic temperate glaciers might be regarded as a modern analogue. In Iceland, "hummocky moraine" forms as a result of meltout and transfer of supraglacial debris from stagnant ice bodies into topographic troughs resulting in topographic inversion and a "chaotic" appearance (Eyles, 1983a; Kjær and Krüger, 2001; Evans, 2003c). Sedimentological studies in Scotland have not been able to confirm the importance of this process, and apart from a few isolated moraines there is very little evidence for dead ice meltout and stagnation (Benn, 1990, 1992a). Apart from the brief note by Benn (1992a: 784) that the erosional and depositional evidence on Skye indicates that the ice was "wet-based throughout", little is known about the former thermal regime and dynamics of the glaciers elsewhere in Scotland.

The second model envisages englacial and proglacial thrusting, described from polythermal surging and non-surging glaciers on Svalbard, as the mechanism responsible for the formation of moraines with characteristic proximal rectilinear slopes (cf. Bennett, 1996; Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002). As discussed above, this work highlights the morphological similarity of ice-marginal moraines at ice margins in Svalbard, one site in the Scottish Highlands (Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998) and one site in Wales (Graham and Midgley, 2000a). As a logical consequence of morphological similarity and despite the absence of detailed sedimentological evidence (see above), however, it has been proposed that Svalbard may form a good analogue for Upland Britain during the Younger Dryas (Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998; Hambrey and Glasser, 2002; Glasser and Hambrey, 2003). Moraines formed under these circumstances cannot be used to reconstruct patterns of ice retreat and carry a different glaciodynamic signature to recessional moraines which imply contrasting palaeoglaciological conditions in upland Britain during the Younger Dryas (cf. Hambrey and Glasser, 2002).

It has yet not been possible to thoroughly test geomorphologically and sedimentologically if either of the two models applies to the Scottish palaeo-context. Although the model of incremental retreat is often preferred over the thrusting hypothesis on morphological grounds (e.g. the presence of crestline bifurcations of moraines), this has rarely been backed up by sufficient sedimentological evidence (e.g. Wilson and Evans, 2000; McDougall, 2001; Ballantyne, 2002a; Wilson, 2002; Lukas, 2003).

A thorough test of the applicability of the two models introduced above is thus necessary to advance our understanding of the nature of glaciation during the Younger Dryas in Scotland and to resolve the question as to whether Scottish "hummocky moraines" might best be represented by either a temperate, a polythermal or a different glacial landsystem. Finding such an analogue is important to understand the palaeoclimatic and palaeoglaciological conditions and their interplay during the Younger Dryas in order to link them to palaeoclimatic reconstructions elsewhere (Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005). Younger Dryas glaciers in Scotland have the potential to allow an insight into onshore glacier response to a global and sudden climatic transition in the North Atlantic region throughout the Younger Dryas while elsewhere, the record might be restricted to isolated ice-marginal positions demarcating the Younger Dryas maximum position (e.g. Fyfe, 1990; Lønne, 1993).

1.1.3.3 Extent of the Younger Dryas glaciation in the far NW Highlands, Scotland

In addition to the question of "hummocky moraine" formation and its implications reviewed above, the extent of glaciers during the Younger Dryas in the study area in the far NW Highlands of Scotland is uncertain. Investigations into the Quaternary history of this area are rare and date back to the earliest geological mapping projects carried out in Britain (Peach and Horne, 1892; Read *et al.*, 1926; Read, 1931). Later studies (Sissons, 1977; Lawson, 1986) reconstructed the extent of Younger Dryas glaciers in a larger area, but were largely restricted to mapping from aerial photographs. Glaciers reconstructed in these investigations are surprisingly small compared to the large ice caps along the western Scottish seaboard during the Younger Dryas (Fig. 1.1), so that a modern re-evaluation of this area is overdue. This is especially the case since work carried out in other parts of Scotland has shown many aspects of these earlier reconstructions to be erroneous with a tendency to underestimate the extent of glaciers in the west of Scotland (Ballantyne, 1989; Tipping, 1989; Benn *et al.*, 1992; Bennett and Boulton, 1993a). Due to aforementioned disparities, and as significant advances in our understanding of modern and ancient glacial environments have been made in the past 20 years, a reassessment of the evidence in the far NW Highlands, a much under-researched area, is timely. The far NW Highlands thus might be a large missing piece of the jigsaw in our understanding of the extent, palaeoclimatic conditions and significance of the Younger Dryas in Scotland. This aspect of the present thesis feeds into the Moine Thrust mapping project of the British Geological Survey.

1.2 Aims and objectives

This thesis seeks to resolve the problems outlined above, and the aims of the present study can be divided into two main threads, specific and general aims. The specific aims can be defined as follows:

(1) to map the glacial geomorphology of the NW Highlands to arrive at a modern interpretation of glacial events in this area with special reference to glacial activity during the Younger Dryas,
(2) to reconstruct the ways in which "hummocky moraines" formed and to gain an insight into the variability of moraine-forming processes during the Younger Dryas in the NW Highlands of Scotland,
(3) to reconstruct the dynamics of Younger Dryas glaciers in the NW Highlands of Scotland,
(4) to test if either of the two hitherto-proposed models of "hummocky moraine" formation is applicable to the Younger Dryas in Scotland,
(5) to shed light on the palaeoglaciological significance of Younger Dryas glaciers and their role in the overall Scottish Younger Dryas landsystem and
(6) to evaluate modern analogues for Younger Dryas glaciers in Scotland using examples from two contrasting modern environments.

Additional, and implicit, general aims are outlined below:

(1) to explore how moraines can contribute to the understanding of palaeo-glacier dynamics,
(2) to contribute to the understanding of processes that lead to moraine formation in two different modern environments where the climatic, glaciological and historical context of glacier fluctuations is well-established,
(3) to compare the dynamics of two present-day glaciated valley landsystems and their sedimentological signatures in different settings and
(4) to make informed statements about possible postdepositional alteration of moraines in these contrasting environments and to transfer this knowledge to Scotland.

The overall objective of this study is to increase the understanding of the significance of moraines as archives in palaeoglaciological research using data from modern glacial landsystems. The main emphasis is on establishing methods that allow realistic reconstructions of palaeo-glacier dynamics using examples from Younger Dryas "hummocky moraines" in the NW Highlands of Scotland. A secondary emphasis, however, is on contributing new data to the understanding of a temperate and a polythermal to cold-based present-day glacial landsystem.

CHAPTER 2 METHODS AND STUDY AREA

2.1 Methods

This chapter will discuss the methods used to solve the problems outlined above. First, sedimentological methods used to log and describe exposures in moraines will be examined, including approaches to quantify clast shape and roundness characteristics. Second, geomorphological mapping techniques will be discussed in the light of a morphostratigraphical framework used for glacier reconstruction. The chapter will conclude with an introduction to the study area in the far NW Highlands, its physiography, bedrock geology and structure and a review of Quaternary research carried out in the area.

2.1.1 Sedimentology

Sediments form important archives that contain information on the processes that led to their deposition. In order to obtain this information and make further inferences from it, careful and detailed recording (logging) of available exposures is required. Furthermore, a knowledge of modern sedimentary processes through the study of present-day analogues from the literature and/or in the field is required if a meaningful interpretation is to be achieved (cf. Benn and Evans, 2004). In modern glacial environments, large advances have been made in recent years in unravelling the processes leading to the formation of moraines, particularly when this data is coupled with historical, glaciological and climatic archives (e.g. Sharp, 1984; Humlum, 1985; Eybergen, 1987; Krüger, 1993, 1994, 1995, 1996, 1997; Winkler and Nesje, 1999; Krüger and Kjær, 2000; Kjær and Krüger, 2001; Bennett, 2001). Studies carried out in palaeo-environments can only benefit from these advances if the interpretation of sedimentary units is closely linked to observations made at modern ice-margins where moraines are currently being formed.

Aside from a few examples (Benn, 1990, 1992a; Benn and Evans, 1993, 1996; Phillips *et al.*, 2002), previous studies attempting to reconstruct the dynamics of former glaciers in Scotland have been hampered by the lack of available exposures. Consequently, point data had to be extrapolated (e.g. Bennett and Boulton, 1993a; Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Lukas, 2003; Mitchell and Lukas, 2004), making it difficult to test to what extent these data are representative of conditions over a larger area. In order to overcome this problem in this study, available sections were supplemented by numerous newly-created exposures through the whole width of moraines by enlarging smaller surface pits and cleaning larger areas of landslips triggered by fluvial undercutting. Fifty-two exposures in moraines of varying extent and quality were examined in total (Appendix 1).

Sedimentological logging of available sections was carried out using two methods: (a) a measured drawing on square millimetre paper was made in the field, using marker cairns along the

base of the section and a tape measure hung from the top of the exposure as a vertical scale. Prominent boulders and unit boundaries were drawn first, with increasing detail being added progressively. In order to increase planimetric accuracy, a photomosaic was taken and the field log later transferred. (b) in cases where sections could be revisited, logging was carried out on overlays of enlarged photomosaics of the cleaned sections in the field (Jones *et al.*, 1999; Evans and Benn, 2004). Section logs presented in this thesis employ a common style of depicting sedimentary structures (Fig. 2.1).

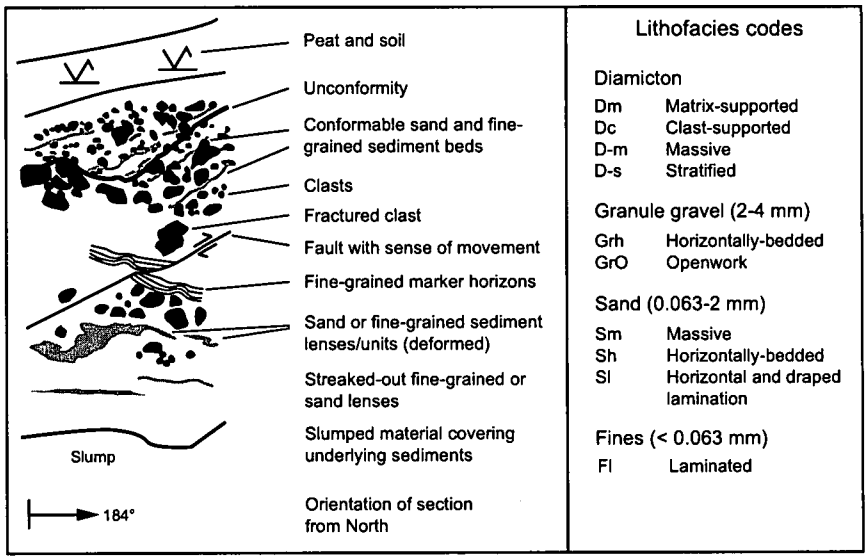


Fig. 2.1 Style of sedimentary logs and lithofacies codes used in this thesis.

Individual sedimentary units were identified and distinguished on the basis of their visual physical properties including grain size range, sorting, compaction, sedimentary structures (depositional, erosional and deformation structures) and clast shape and roundness (see below). The nature of contacts between individual units was also recorded. A slightly modified version of the lithofacies code introduced by Eyles *et al.* (1983) is employed for effective and rapid description in sedimentary logs (Evans and Benn, 2004). The dip and strike of moraine surface slopes, selected units, fault planes and fold axes was measured using a Recta compass-clinometer.

2.1.2 Clast shape and roundness

Clast morphology is a widely-used criterion to discriminate between different erosion and transport histories in differing environmental settings (Ballantyne, 1982; Benn, 1990, 1992a, 1994; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Salt and Ballantyne, 1997; Evans, 1999; Glasser *et al.*, 1999; Graham and Midgley, 2000a, b; Etienne *et al.*, 2003). Morphological characteristics can be divided into three properties: (a) *clast shape*, defined in terms of ratios between three orthogonal axes, the longest, intermediate and shortest axes, denoted as a, b and c axis, respectively; (b) *clast roundness* describes the degree of curvature around the clast edges. This is usually quantified using visual

comparison charts or descriptive criteria (Benn and Ballantyne, 1994; Table 2.1); (c) *surface texture* includes more delicate features such as striae and is more difficult to quantify (Benn, 2004).

Triangular diagrams following Sneed and Folk (1958) are particularly useful to display clast shape in an undistorted way that is free from bias (Benn and Ballantyne, 1993; Benn, 2004). On such ternary diagrams, the ratios of c:a and b:a axis are used to distinguish three endmembers of a clast continuum: equant or blocky shapes where $a \approx b \approx c$, prolate or elongate shapes where $a >> b \approx c$ and oblate or slabby shapes where $a \approx b > c$ (Benn, 2004). Ballantyne (1982), Benn (1992a) and Benn and Ballantyne (1993, 1994) have shown that the C_{40}-Index, defined as the percentage of clasts with a c:a ratio <0.4, is a powerful index to distinguish blocky from elongated clast shapes and works well in glaciated environments. Clast roundness characteristics (Table 2.1) can also be used to distinguish between, for example, edge-rounded, subglacially-transported and angular clasts that were affected by frost-weathering. For this purpose, the RA index, which denotes the percentage of very angular and angular clasts in a sample, will be used (cf. Benn, 1990, 1992a; Benn and Ballantyne, 1994). In order to discriminate between clasts with different histories, the C_{40}-Index is plotted against the RA-index in a co-variance plot (Benn and Ballantyne, 1994).

In this study, clast shape and roundness will be used to distinguish different clast erosional and transport histories as this combined approach has been shown to work very effectively in glacial environments (Benn, 1992a, 1994; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Evans, 1999; Lukas *et al.*, 2005a). The shape of samples of 50 clasts picked from a restricted area of max. 25*25 cm within individual facies units was measured with a tape measure and plotted using Tri-Plot (Graham and Midgley, 2000b). Clast roundness was estimated following the criteria outlined in Table 2.1 and the frequency of individual roundness classes plotted using Microsoft Excel. Control samples of known origin were obtained so that samples with an unknown transport history could be compared against a known datum to enable a meaningful reconstruction of the transport paths and source areas of clasts (cf. Benn and Ballantyne, 1994; Benn, 2004). In order not to introduce potential errors, clasts were always chosen from a restricted size range between 3 and 15 cm. It has been shown that different lithologies affect clast shape (Benn and Ballantyne, 1994; Krüger and Kjær, 1999; Benn, 2004), although some have argued that this might not affect results greatly (Bennett *et al.*, 1997). However, a close look at the data used in the latter study reveals that envelopes of individual lithologies overlap significantly and that the use of a mixture of lithologies can introduce a larger scatter of data, thereby making results less precise and reproducible. To exclude such potential complications, all samples were thus restricted to Moine psammite, the dominant lithology in the study area.

Table 2.1 Criteria employed in the identification of clast roundness classes (modified from Benn and Ballantyne, 1994).

Roundness class	Description
Very angular (VA)	Very acute edges and/or sharp protuberances
Angular (A)	Acute edges with no evidence of rounding
Sub-angular (SA)	Rounding confined to edges; faces intact
Sub-rounded (SR)	Rounding of edges and faces; often facetted
Rounded (R)	Marked rounding of both edges and faces; merging of edges and faces
Well-rounded (WR)	Distinction between faces and edges not possible

2.1.3 Geomorphological mapping, morphostratigraphy and glacier reconstruction

2.1.3.1 Geomorphological mapping

The distribution of moraines and other glacial landforms in the NW Highlands was mapped onto topographic maps at scales of 1:25,000. Aerial photographs at scales of ca. 1:25,000 were used in addition to determine the exact location, shape and planform of glacial and periglacial landforms (Appendix 4).

Due to the nature of projection, aerial photographs differ significantly from topographic maps, and these differences have to be borne in mind when transferring information from one to the other. Aerial photographs display geometric distortion, which has two major practical effects: (a) relief displacement and (b) variation of scale over the whole aerial photograph (Lillesand and Kiefer, 2000). Relief displacement can be explained in terms of the type of photographic (perspective) projection that is used. The distortion of features with different altitudes above a datum level is caused by the fact that their top is closer to the camera lens than their base so that objects appear to lean away from the centre of the photograph. Also, slopes are vertically exaggerated by about three to four times due to this effect (Albertz, 2001). Scale variation is a direct result of (a) as distances between two points are rarely at the same altitude; thus, measuring only *one* scale on an aerial photograph does not give a good result as scale variation increases towards the edge of aerial photographs. Thus, a few distances from different parts of the same aerial photograph should be measured to determine an average scale (Lillesand and Kiefer, 2000), and this has been done for the ones used in this study throughout.

In Scotland, where the majority of fieldwork was carried out, aerial photograph interpretation was used prior to and after fieldwork, thereby enhancing the accuracy of the maps and reducing possible errors that might arise from misinterpretation of features when only viewed once from aerial photographs or on the ground. For example, the recognition of different terrace altitudes from aerial photographs proved difficult so that individual terrace fragments were mapped entirely in the field. Conversely, mapping of moraines from aerial photographs yielded much better results with respect to their orientation and shape, particularly where larger areas of "hummocky moraine" exist. In order to record the location and planform of moraines accurately, mapping was done on an acetate overlying

one aerial photograph of a stereopair. Drawing on acetates was carried out while looking through a mirror stereoscope with threefold magnification.

The main problem associated with this method is the geometric distortion present in aerial photographs as described above so that the overlays could not simply be transferred to the base maps even though the scales were very similar. In order to get rid of this distortion, datum lines and reference points on the base maps were joined up with those transferred from aerial photographs (cf. Hubbard and Glasser, 2005). Drainage features such as rivers and lakes are the most obvious and least distorted features since they are located in lower parts of the relief (Kronberg, 1984). Moreover, rivers often show distinct and unique meander bends that enable easy and reliable matching of the information contained on the acetates and the topographic base maps. This is especially the case in the vicinity of river confluences as the number of points that can be used to "lock" the position of a feature increases. Due to the fact that the problem of distortion can be solved quite easily, this method presents a quick and accurate way of mapping geomorphological and geological features (Kronberg, 1984) and proved invaluable for the successful mapping of the study area. This process of joining and locking information to produce the final geomorphological map was facilitated electronically in Adobe Photoshop to un-distort the scanned acetates using the transform functions. The final geomorphological map (Appendix 2) was then produced by importing this information in Adobe Illustrator 10.0 where it was digitised together with features such as trimlines, ice-moulded bedrock and roches moutonnées that had been mapped on to the field maps.

In the field, geomorphological mapping was conducted with reference to landmarks, with the aid of a compass or, where clear reference points were lacking, a Garmin Summit 12-channel GPS. The orientation of ice flow-directional indicators such as striae, roches moutonnées and ice-moulded bedrock were measured using a compass and corrected from magnetic north to grid north before being placed on the map. In the case of striae, one arrow on the map represents the arithmetic mean of five measurements taken from outcrop areas of not larger than 25 m^2. Single measurements were taken perpendicular to the lee faces of roches moutonnées and parallel to larger-scale grooves in areas of ice-moulded bedrock.

2.1.3.2 Morphostratigraphy

Introduction and background

Morphostratigraphy refers to the application of stratigraphic principles to geomorphology, in other words utilising the spatial relationships between individual landforms to assign them to events or periods (e.g. Skinner and Porter, 1987; Schellmann and Radtke, 2004). This approach has been successfully applied in glacial geomorphology (e.g. Colhoun, 1988; Lemmen and England, 1992; Chadwick *et al.*, 1997; Lehmkuhl, 1998; Richards *et al.*, 2000; Dahms, 2002), and, although it underlies many glacier reconstructions in Scotland, clear morphostratigraphic criteria for identifying

specific periods of glaciations, e.g. the Younger Dryas, are still lacking. In glacial geology and geomorphology – as in any discipline of earth sciences – numerical dating is crucial to understand the rates and timing of change (cf. Skinner and Porter, 1987; Nesje and Dahl, 2000). Direct dating of glacier limits, i.e. moraines, is often hampered by the restricted availability of dateable material (cf. Nesje and Dahl, 2000), and usually proglacial sites have been used to date a given period of glaciation (e.g. Nesje and Dahl, 2000; Orvis and Horn, 2000; Geirsdóttir *et al.*, 2000; Matthews *et al.*, 2000; Dahl *et al.*, 2003; Bakke *et al.*, 2005). New techniques that allow the direct dating of moraine formation have been developed more recently, for example dating by cosmogenic radionuclides and/or optically-stimulated luminescence techniques applied to glacigenic and associated deposits (e.g. Ivy-Ochs *et al.*, 1996, 1999; Richards *et al.*, 2000; Everest and Golledge, 2004; Kelly *et al.*, 2004a; Kaplan *et al.*, 2004; Preusser, 2003, 2004).

In Scotland, dating of glacial limits has traditionally been carried out by comparing the peat stratigraphy contained within kettle holes inside and outside a glacial limit (e.g. Sissons and Walker, 1974; Gray and Coxon, 1991). This holds especially true in the context of establishing the extent and duration of glaciation during the Loch Lomond Stadial or Younger Dryas (cf. Sissons and Walker, 1974; Walker, 1975a, b; Sissons, 1976; Walker *et al.*, 1988; Gray and Coxon, 1991; Benn *et al.*, 1992). However, the scarcity of suitable sites has resulted in only few well-dated areas where the timing of the glaciation during the Younger Dryas has been established beyond doubt (e.g. Gray and Brooks, 1972; Walker *et al.*, 1988; Benn *et al.*, 1992; Ballantyne, 2002a). In addition, a few sites exist where the extent and timing of the Younger Dryas maximum could be established in settings different from the kettle hole approach (e.g. Rose *et al.*, 1988; Merritt *et al.*, 1990; Tipping *et al.*, 2003).

Although numerically constraining a glacial event, such as the Younger Dryas in Scotland, is of critical importance, comparison of characteristic sediment-landform assemblages inside and outside a proposed Younger Dryas limit are recognised as a useful means of glacier reconstruction (e.g. Benn and Ballantyne, 2004, 2005). For example, due to the limited availability of dateable sites or material and for economic reasons, it is rarely possible to date every outlet glacier lobe of a former ice cap and demonstrate synchroneity (e.g. Walker *et al.*, 1988; Ballantyne, 1989, 2002a; Benn *et al.*, 1992; Benn, 1997a; Benn and Ballantyne, 2005). Thus, the necessity of extrapolating characteristic geomorphological evidence from dated sites to those that remain undated arises (Lowe and Walker, 1997; Benn and Ballantyne, 2005).

As discussed above (Chapter 1.1.1.2), extrapolation of only one line of evidence, in this case the occurrence of "hummocky moraine", to infer the extent of glaciation during the Younger Dryas, can potentially lead to erroneous reconstructions. It is thus crucial to rely on assemblages of geomorphological features unequivocally associated with the Younger Dryas to extrapolate evidence across an area covered by an ice cap. Surprisingly, despite a large number of reconstructions that have been carried out across Scotland, a holistic and uniform approach incorporating available

geomorphological evidence and characteristic sediment-landform associations has never been formalised apart from early attempts by Sissons and his co-workers (e.g. Sissons, 1974, 1976, 1977). To overcome these shortcomings, a unifying approach utilising information from published sources is developed below (Table 2.2).

Characteristic sediment-landform associations

Moraines. Early glacier reconstructions in Scotland referred to large tracts of moraines with an apparently chaotic appearance when viewed from the ground as "hummocky moraine" and frequently used them to delineate the maximum extent of glaciers (e.g. Sissons, 1967, 1974, 1977, 1979a; Ballantyne and Wain-Hobson, 1980; Gray, 1982; Ballantyne, 1989, 2002a; Hughes, 2002). Although the use of "hummocky moraines" on its own can be problematic (Chapter 1.1.1.2) the coincidence of the termination of such fields of moraine ridges and mounds with the maximum extent of the Younger Dryas – where it has been dated – is conspicuous (cf. Benn and Ballantyne, 2004, 2005). On the Isles of Skye and Mull, and elsewhere in Scotland, more subdued moraines tend to occur outside the Younger Dryas maximum extent (e.g. Sissons, 1976; Robinson and Ballantyne, 1979; Ballantyne, 1988; Brown, 1993; Benn, 1997a), whereas inside the Younger Dryas limits, "hummocky" recessional moraines prevail (cf. Ballantyne, 1989, 2002a; Benn, 1992a; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b; Lowe and Walker, 1997; Benn and Ballantyne, 2005). The contrast between areas inside and outside the Younger Dryas glaciation was recognised early on and is fittingly described by Read (1931: 210): "During this epoch of the valley-glaciers there were deposited in the valleys the tumultuous morainic mounds and dumps which form so marked a contrast to the smooth surfaces of the ground-moraine of the earlier stages." In a number of cases, especially where the bedrock is very resistant to erosion and fine material is not readily produced by glacier grinding, or where glaciers have been inferred to be relatively inactive, end moraines may be represented by bouldery ridges devoid of fine material (e.g. Sissons, 1977; Everest, 2003).

River terraces. Evidence that has been used to reconstruct the extent and discuss the broad timing of glacial events also comes from fluvial terraces. Early models (e.g. Penck and Brückner, 1901/1909) correlated terraces to distinct end moraines and assigned them to different glacial stages. Although simplistic, recent research has demonstrated that this pioneering model still holds true in its basic assumptions and that terrace formation, which is a response of the river to changing boundary conditions such as discharge, sediment load and transport capacity (Knighton, 1998; Gurnell *et al.*, 1999), is essentially controlled by climate change (e.g. Lowe and Walker, 1997; Bridgland, 2000; Eden *et al.*, 2001) and the position of the glacier front in relation to the sandur fan or valley train (Gurnell *et al.*, 1999; Marren, 2002). It is now generally accepted that the formation of large river terraces is linked to the transition from a cold to a warm period as discharge and sediment load are

generally increased during this period thereby leading to degradation (cf. Bridgland, 2000; Eden *et al.*, 2001). This background can be transferred into a concept which allows the relative dating of a glacial event if this was distinct and dominant in the past glacial record. Research in numerous areas of Scotland has led to the general notion that extensive terrace sequences are only found outside the areas covered by glaciers during the Younger Dryas (cf. Young, 1974, 1978; Auton, 1990; Benn, 1991a), although this evidence has never been consciously employed in glacier reconstruction.

Although the above review suggests that river terraces might form a good indication of a former ice margin, a note of caution must be added. In some areas such as the periphery of the Cairngorms, which has not been glaciated since the Late Devensian ice sheet retreated from it, up to 5 terraces can be found which have been distinguished by the degree of soil formation and radiocarbon dating (Robinson-Rintoul, 1986). In coastal areas that are directly affected by sea level, and thus base level, change, this relationship is also rendered useless (Tipping *et al.*, 1994). In other cases, morphological terrace forms might not be the result of river aggradation and degradation, but might represent former lake floors that have been incised following lake drainage. Such cases are well-documented in a number of locations in Scotland that both date to, and pre-date, the Younger Dryas (e.g. Sissons, 1982; Peacock and Cornish, 1989; Benn, 1989, 1992b, 1996; Gray, 1992; Golledge, 2003), thus illustrating that the simple relationships between climate and river response breaks down under certain circumstances. These examples illustrate potential problems of misinterpreting terrace sequences when used on their own.

Glaciofluvial landforms and sediments. Landforms and sediments produced by glacial meltwater have been recognised in large areas that were last covered by the decaying Late Devensian ice sheet and usually take the form of thick outwash terraces or valley trains (e.g. Auton, 1990), kames (Young, 1974, 1978) and eskers (Young, 1974, 1978; Thomas and Montague, 1997). Such landforms indicative of large-scale stagnation have not been described from areas covered by glaciers during the Younger Dryas (cf. Sissons, 1974, 1977; Thorp, 1986, 1991; Ballantyne, 1989, 2002a; Benn, 1990, 1991a, 1992a; Bennett and Boulton, 1993a, b; Benn and Ballantyne, 2005). A gradual or sudden contrast in such markedly different landform inventories inside and outside the Younger Dryas limits could thus help determine the location of such a limit.

Shorelines. In areas where glaciers during the Younger Dryas extended below sea-level, such as in parts of the western Isles of Skye and Mull, raised shorelines can give important information as to the age of a glacial event (Gray and Brooks, 1972; Walker *et al.*, 1988; Ballantyne, 1989, 2002a; Benn, 1991b). In such areas, a Younger Dryas age of moraines and other landforms can be inferred by the presence of higher, older raised shorelines outside the glacial limits and their absence inside. Inside Younger Dryas limits, only raised shorelines of Holocene age can be found (cf. Walker *et al.*, 1988;

Benn, 1991b). This is possibly due to a combination of two factors. (a) the shorelines were being produced while adjacent areas were occupied by a glacier and no shoreline could be formed; only after the ice had retreated could younger shorelines form in formerly glaciated areas. (b) older shorelines were eroded as the glaciers advanced to their maximum extents (cf. Sissons and Dawson, 1981).

Sediment-covered slopes. Previous research has shown that the extent of thick debris mantles on slopes, which are frequently extensively gullied, often coincides with areas covered by glaciers during the Younger Dryas as this limit can be linked to other evidence such as periglacial trimlines (e.g. Ballantyne, 1989, 2002a; Hughes, 2002). Sediment accumulations are either markedly thinner or completely absent in upslope parts that remained above the Younger Dryas glacier limits (e.g. Thorp, 1986; Ballantyne, 1989, 2002a; Gray and Coxon, 1991; Hughes, 2002). Where such a clear upslope termination of the sediment cover (often referred to as "drift limit" in the literature) exists, it has been frequently utilised in the reconstruction of the Younger Dryas ice surface (e.g. Ballantyne and Wain-Hobson, 1980; Gray, 1982; Thorp, 1986; Ballantyne, 1989, 2002a; Gray and Coxon, 1991; Hughes, 2002; Benn and Ballantyne, 2005).

A variant to the use of "drift limits" is to consider the form and thickness of talus slopes. Although this has apparently never explicitly been used in previous studies, observations in the study area suggest that the thickness of talus cones and sheets varies inside and outside the areas covered by glaciers during the Younger Dryas. Previous work has demonstrated that the Younger Dryas was the last period of intense periglacial activity affecting Upland Scotland (e.g. Ballantyne, 1984, 1991, 1998; Ballantyne and Harris, 1994). Thus, extensive and "mature" talus slopes would have been produced at non-glaciated sites that were affected by enhanced frost weathering during this period (Ballantyne and Eckford, 1984; Ballantyne, 1991: 91) with comparatively little addition of material throughout the Holocene (Ballantyne, 1991; Hinchcliffe *et al.*, 1998; Curry, 2000). Likewise, protalus ramparts would only have developed outside the areas covered by glaciers during the Younger Dryas (Ballantyne, 1986, 1991; Ballantyne and Kirkbride, 1986, 1987). Conversely, the presence of laterally terminating thick talus sheets along slopes and the presence/absence of protalus ramparts could potentially be used to reconstruct a distinct glacial limit.

Glacially transported boulders/erratics. Similar to the upslope termination of a glacigenic sediment cover, the occurrence of numerous large erratics or glacially-transported boulders on slopes and in the valley bottoms has been observed to coincide with the former Younger Dryas glacial limit (e.g. Sissons, 1977; Ballantyne and Wain-Hobson, 1980; Gray, 1982; Thorp, 1986; Benn, 1990, 1992a; Gray and Coxon, 1991; Hughes, 2002). This might be attributed to the fact that boulders in modern glacial environments are released from the glacier surface by toppling and sliding forming either distinct ridges or confined boulder spreads that mark the maximum extent of a glacier (cf. Sharp,

1984; Winkler, 1996a; Benn and Evans, 1998). Furthermore, erratic dispersal trains, together with striae, roches moutonnées and ice-moulded bedrock, can in this respect be used to reconstruct the palaeo-ice flow direction and to constrain ice-surface contours (Ballantyne and Wain-Hobson, 1980; Sutherland, 1984a; Thorp, 1986; Ballantyne, 1989, 2002a; Benn, 1990, 1992a; Benn and Ballantyne, 2005).

Periglacial trimlines. Areas that remained above the ice masses during the Younger Dryas experienced intense frost weathering that led to the production of *in situ* blockfields, large-scale solifluction sheets or terraces and large-scale patterned ground phenomena (Sissons, 1974; Ballantyne and Wain-Hobson, 1980; Ballantyne, 1984, 1997; Ballantyne and Harris, 1994). Such features, summarised under the term "mountain-top detritus" by Ballantyne (1998) often have a clear boundary with ice-moulded bedrock, roches moutonnées and a cover of glacigenic sediments that record glacial activity, and hence protection from prolonged severe frost weathering, in the lower parts of valleys and mountainsides. This boundary, which is referred to as a periglacial trimline, can occur over 10-50 m vertically, depending on bedrock lithology (Ballantyne *et al.*, 1998a, b) and has been widely used in glacier reconstructions to delineate the former ice surface, in Scotland (Ballantyne and Wain-Hobson, 1980; Gray, 1982; Thorp, 1986; Ballantyne and Harris, 1994; Ballantyne, 1997, 1998; Ballantyne *et al.*, 1997, 1998a, b; Benn and Ballantyne, 2005) and elsewhere (e.g. Florineth and Schlüchter, 1998, 2000; Kelly *et al.*, 2004b).

Towards a reliable morphostratigraphic approach

The evidence reviewed above has extensively been used in glacier reconstructions in Scotland (and elsewhere) so far and has proven successful on a number of occasions. However, as indicated by the debate over the origin and timing of "hummocky moraine" formation (Chapter 1.1.1.2), caution is required not to fall into the trap of attaching too many assumptions to only one of these lines of evidence. The author suggests that, in order to eliminate this problem, it is crucial to remain critical of one's own mapping. Ideas developed during mapping should constantly be tested and supporting or contrasting evidence should actively be sought. For example, if a clear downvalley termination of "hummocky moraine" is encountered, this potential indication for a Younger Dryas maximum position (cf. Sissons, 1974, 1976, 1979a, b) should be tested with respects to its links to other landform assemblages reviewed above, e.g. whether a transition from one to two river terraces coincides with this limit and whether this in turn can be linked to the extent of sediment blankets and trimlines on slopes. In comparison with other earth-science disciplines, this translates into a multi-proxy approach (Mann, 2002). Thus, the more strands of evidence converge, the more reliable the geomorphological model will be. Ultimately, a sound morphostratigraphy can either be used as a guideline for site selection in dating programmes or to extrapolate existing dates with confidence.

Table 2.2 Summary of contrasting geomorphological evidence used in the reconstruction of the maximum extent of Younger Dryas glaciers in Upland Britain.

Landform/Landform assemblage	Inside Younger Dryas limits	Outside Younger Dryas limits	References
Moraines	"Hummocky moraine", clear end moraines, bouldery moraines	Subdued, isolated moraine ridges	Sissons, 1967, 1974, 1976, 1977, 1979b; Robinson and Ballantyne, 1979; Ballantyne and Wain-Hobson, 1980; Gray, 1982; Ballantyne, 1988, 1989, 2002a; Benn, 1990, 1992a; Hughes, 2002
Terraces	One river terrace above present floodplain	Flights of distinct river terraces	Previously not explicitly used in glacier reconstruction
Glaciofluvial landforms	Outwash fans and smaller terraces, occasional beaded eskers	Large kames, eskers, extensive kettled outwash plains	Young, 1974, 1978; Sissons, 1974, 1976; Thomas and Montague, 1997
Shorelines	Only Holocene shorelines present	Lateglacial shoreline present	Charlesworth, 1955; Gray and Brooks, 1972; Walker *et al.*, 1988; Ballantyne, 1989, 2002a; Benn, 1991b
Slope features	-Intensively gullied slopes, small debris cones, localised "immature" talus sheets or cones -"freshly" glacially polished bedrock with little debris accumulation at foot	-Large debris fans at foot of slope -Extensive "mature" talus sheets	Sissons, 1974; Ballantyne and Wain-Hobson, 1980; Ballantyne, 1984, 1989, 1991, 2002a; Ballantyne and Eckford, 1984; Ballantyne and Harris, 1994; Salt and Ballantyne, 1997; Hinchcliffe *et al.*, 1998; Curry and Ballantyne, 1999; Curry, 2000; Benn and Ballantyne, 2005
Vertical limits on bedrock (periglacial trimlines)	Ice-moulded and plucked bedrock	Heavily jointed bedrock, blockfields, thick *in situ* regolith, patterned ground ("mountain-top detritus")	Ballantyne, 1982, 1989, 1997, 1998; 2002a; Thorp, 1986; McCarroll *et al.*, 1995; Ballantyne *et al.*, 1997, 1998a, b; Hughes, 2002; Benn and Ballantyne, 2005
Vertical limits on sediment-covered slopes	Upslope and lateral termination of thick sediment cover ("drift limit"), limits of glacially-transported boulders, meltwater channels, lateral moraines	Large solifluction lobes, patterned ground, scree,	Gray and Brooks, 1972; Thorp, 1986; Ballantyne, 1989, 2002a; Hughes, 2002; Benn and Ballantyne, 2004, 2005

2.1.3.3 Glacier reconstruction

Introduction and background

The importance of reconstructions of past glacier extent in palaeoenvironmental research has been discussed above. This subchapter deals with information that can be derived once the glacier limits are correlated and combined into a reconstruction of a single ice mass. In addition to understanding the distribution of past glaciers, it has long been recognised that they act as terrestrial palaeoclimatic archives from which much can be learned about former cryosphere-atmosphere interactions. This includes the reconstruction of the former moisture-bearing winds based on the distribution of glaciers and their relationship to the topography (e.g. Sissons, 1974, 1977, 1979a; Sissons and Sutherland, 1976; Florineth and Schlüchter, 1998, 2000; Kelly *et al.*, 2004b), the equilibrium-line altitude (Benn *et al.*, 2005, references therein and below), and, based on the latter, the calculation of palaeo-precipitation values (e.g. Ballantyne, 1989, 2002a; Kerschner *et al.*, 2000; Benn and Ballantyne, 2005). It is crucial to determine these palaeoclimatic variables at the time of glacier existence for two reasons: (a) to understand the relationships between climate and glacierisation and assess the changes in temperature and precipitation compared to modern values (e.g. Ballantyne, 1989, 2002a; Kerschner *et al.*, 2000; Benn and Ballantyne, 2005). (b) to use known palaeoclimatic relationships to constrain boundary conditions of numerical models used to predict future climate change (e.g. Hubbard, 1999; Isarin and Renssen, 1999; McAvaney *et al.*, 2001; Golledge and Hubbard, 2005).

Equilibrium-line altitudes (ELAs)

The equilibrium-line altitude (ELA) separates the accumulation from the ablation area and thus connects the points on a glacier surface where the annual mass balance is zero, i.e. where accumulation equals ablation (Benn and Evans, 1998). It is therefore closely coupled with, and dependent on, local climatic conditions, with air temperature and precipitation being the most important variables (Ohmura *et al.*, 1992; Benn and Evans, 1998). In reconstructions of palaeo-ELAs it has to be assumed that the reconstructed ice mass was in equilibrium with climate, and therefore, a steady-state ELA is usually reconstructed (Benn and Evans, 1998; Benn and Lehmkuhl, 2000; Benn and Ballantyne, 2005; Osmaston, 2005).

In order to allow comparison with previous studies in the NW Highlands, but also elsewhere in Scotland, three different methods were employed to estimate the ELA. First, the Accumulation-Area Ratio (AAR) was employed. This method assumes that the accumulation area occupies some fixed proportion of the glacier area and has been used in a variety of settings (Porter, 1975; Meierding, 1982; Leonard, 1989; Nesje, 1992; Torsnes *et al.*, 1993; Benn and Evans, 1998). To allow comparison with previous studies (Ballantyne, 2002a; Benn and Ballantyne, 2005), ratios of 0.5 and 0.6 will be employed here. Second, the area-weighted mean altitudes of glacier surfaces, developed by Sissons (1974) will be used here as this will allow comparison with the original reconstruction of Younger

Dryas glaciers in the far NW Highlands by Sissons (1977). However, both these methods are based on assumptions that are rigid and need to be adjusted to specific glacier catchment characteristics. While the area-weighted mean altitude method (Sissons, 1974, 1977) assumes that ablation and accumulation gradients are identical, observations on modern glaciers have shown that the ablation gradient is typically much steeper than the accumulation gradient (e.g. Oerlemans, 2001; Kaser and Osmaston, 2002). On the other hand, the AAR method is based on the assumption that every part of the accumulation and ablation areas contributes equally to mass balance. However, both accumulation and ablation increase significantly with distance form the ELA implying that the glacier terminus and source area will contribute more to the overall mass balance than those near the ELA (Benn and Evans, 1998). Thus, both methods do not yield very realistic results, and comparison between sites is limited (Nesje, 1992; Torsnes *et al.* 1993; Benn and Evans, 1998; Ballantyne, 2002a; Benn and Ballantyne, 2005; Osmaston, 2005). Such shortcomings have been alleviated by the development of balance ratio methods which take into account glacier hypsometry and variable mass balance gradients (Furbish and Andrews, 1984; Benn and Gemmell, 1997; Osmaston, 2005), and thus an Area-Altitude Balance Ratio (AABR) will be employed in addition.

All three methods rely on accurately constrained and contoured ice masses. Contour lines were drawn at 50 m intervals following the methods established by Sissons (1974, 1977) and Ballantyne (1989, 2002a). This involved constructing the contours where the ice surface intersects contours on the topographic base map, and the ice surface contours were drawn at right angles to the orientation of ice-flow directional indicators. In order to further constrain the dimensions of individual glacier catchments, ice divides were drawn where flow was diffluent (cf. Sissons, 1974, 1980; Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005). Diffluences have been reconstructed where ice-flow directional indicators or the trend of the ice surface contours suggested divergence of flow, for example around nunataks.

Calculation of the different ELA estimates using the methods described above were carried out using Excel spreadsheets. The planimetric area of each individual altitudinal envelope (e.g. 50 to 100 m) was measured for each individual glacier basin and the value entered into the spreadsheet. In case of the AAR values, cumulative frequencies of the glacier area in each altitudinal envelope were calculated and plotted against altitude. Values of AAR = 0.5 and AAR = 0.6 were then read off the graph. In the case of the area-weighted mean altitude the area of each altitudinal envelope is multiplied by its mid-point value and its sum divided by the total glacier area. The AABR approach utilises a spreadsheet presented by Osmaston (2005) which had been developed from an earlier, more complex version of Benn and Gemmell (1997).

Since summer air temperature and annual precipitation control the position of the ELA (see above), it is possible to use the ELA values to calculate palaeoclimatic variables from it (e.g. Ohmura *et al.*, 1992; Benn and Evans, 1998; Benn and Ballantyne, 2005). However, the methods involved in

this will be discussed in Chapter 3.4.2, since they are only meaningful if placed in the context of actual data and to avoid repetition.

2.2 Study area

2.2.1 Physiography

The study area comprises ca. 1000 km^2 of the North-West Scottish Highlands in the county of Sutherland between NC 250 540 in the northwest, NC 250 140 in the southwest, NC 500 540 in the northeast and NC 500 140 in the southeast (58°5'-58°29'N; 4°58'-4°34'W; Fig. 2.2).

This area was chosen for investigation as its Quaternary geology and geomorphology have received very little research attention over the past century. While the Quaternary geomorphology of the area west of the main watershed (Fig. 2.2) is relatively well-known (see below), virtually no Quaternary research has been carried out to the east of the watershed except for palaeoecological studies based on Holocene bog and marine sediment successions. Knowledge about the geomorphological inventory of this area is thus scarce, and, as a result of this, our understanding of the glacial dynamics in this area is very incomplete.

Three physiographic regions can be distinguished. The first comprises large areas of low ground (< 350 m), especially in the west and northwest, underlain by Lewisian Gneiss. Such areas are dominated by bedrock at or near the surface, steep-sided rock knolls and numerous small lochans, a type of terrain that Linton (1963) termed "knock and lochan topography" (cf. Rea and Evans, 1996). Till is generally absent from this surface and only where the lithology of the underlying bedrock changes to Torridonian Sandstone (e.g. at the Point of Stoer), can a cover of glacigenic sediments be found.

The second region comprises a belt of mountains with an approximate width of ca. 20-30 km that extends from Ben More Assynt (998 m, NC 318 202) in the south via Beinn Leoid (792 m, NC 320 295) and Ben Hee (873 m, NC 426 341) to Foinaven (914 m, NC 317 508) and Ben Hope (927 m, NC 478 502) in the north. The main watershed is located within this zone and crosses the area in a roughly SSW-NNE direction (Johnstone and Mykura, 1989; Fig. 2.2).

The third physiographic region consists of the ground to the east which is dominated by a variably thick cover of blanket bog that has developed on generally acidic, base-poor Quaternary sediments that in turn overlie Moine schists (Lindsay *et al.*, 1988; Charman, 1994). This gives the land a gently undulating to near-horizontal surface expression below 250 m altitude. However, some areas of higher relief stand out of this peat-covered surface, e.g. Ben Klibreck (961 m, NC 585 299) and the mountains to the southeast where Creag Mhòr (NC 699 240) forms the highest peak at 713 m. Such peaks are the result of differential erosion as the rocks underlying these mountains are more resistant than the surrounding Moine rocks.

2.2.2 Pre-Quaternary geology

2.2.2.1 Stratigraphic succession

The pre-Quaternary geological strata can be divided into five units (Fig. 2.3). The oldest unit found in the NW Highlands is the Lewisian Complex with an age between 3.0-2.7 Ga (Corfu *et al.*, 1998; Park *et al.*, 2002). It consists largely of gneisses, which experienced multiple deformation during metamorphic events ranging from high-grade to low-grade metamorphism (Park *et al.*, 2002).

Following a prolonged period of burial deep in the Earth's crust, erosion of the Lewisian rocks commenced once they were exposed to weathering processes on the earth surface. Sedimentary rocks derived from erosion of the Lewisian complex consist of arkosic sandstones and conglomerates with subordinate shales that represent fluvial and lacustrine sediment facies. These sediments are generally attributed to the Torridonian succession ("Torridonian Sandstone") which are of late Precambrian age (ca. 1000 Ma; Strachan *et al.*, 2002) and crop out along the west coast.

Farther east, thick sedimentary successions, which might also have been derived from the Lewisian, were deposited in a transition from continental shelf to shallow sea environments. Unlike the Torridonian rocks, however, these were affected by metamorphic alteration and deformation during the Grampian Orogeny around ca. 470-460 Ma and are now represented by psammitic, semipelitic and pelitic gneisses and schists (Strachan *et al.*, 2002). These rocks belong to the Morar Group of the Moine Supergroup and crop out east of the Moine Thrust Zone (see below). Moinian metasedimentary rocks are, however, locally interrupted by inliers of Lewisian gneiss that have been interpreted as thrust slices and fold cores and are commonly referred to as "Lewisianoid" (e.g. Soper *et al.*, 1998; Strachan *et al.*, 2002). It has not yet been possible to establish correlations between the Moine and Torridonian rocks or with the lithologically similar deposits belonging to the Grampian Group south of the Great Glen Fault (Harris, 1995; Strachan *et al.*, 2002).

Following the deposition of the Torridonian Sandstone, crustal warping led to gentle folding of the Caledonian Foreland. During the following period of erosion, hundreds of metres of Torridonian strata were removed, locally leading to exposure of the underlying Lewisian.

In early Cambrian and Ordovician times, marine transgression led to the deposition of a series of clastic and carbonate sediments that only crop out in a narrow zone, the Moine Thrust Zone (see below). From the oldest these are the false-bedded quartzite, basal quartzite or pipe rock, the Fucoid Beds, and the Saltarella Grit formation (Johnstone and Mykura, 1989). The Durness Group, for which seven formations have been identified at the type site at Durness, consists of limestones, dolomitic limestones and dolomites (Walton and Oliver, 1991) and crops out in a relatively small area just west of the Moine Thrust (Fig. 2.3).

Igneous intrusions into aforementioned strata comprise ultrabasic and granitic bodies, of which the biggest occur outside the present study area. Within the study area, numerous smaller veins and intrusive bodies of ultrabasic composition are currently being investigated in order to constrain the

age of the development of the Moine Thrust Zone (cf. Strachan *et al.*, 2002; D. Cheer, pers. comm., 2005).

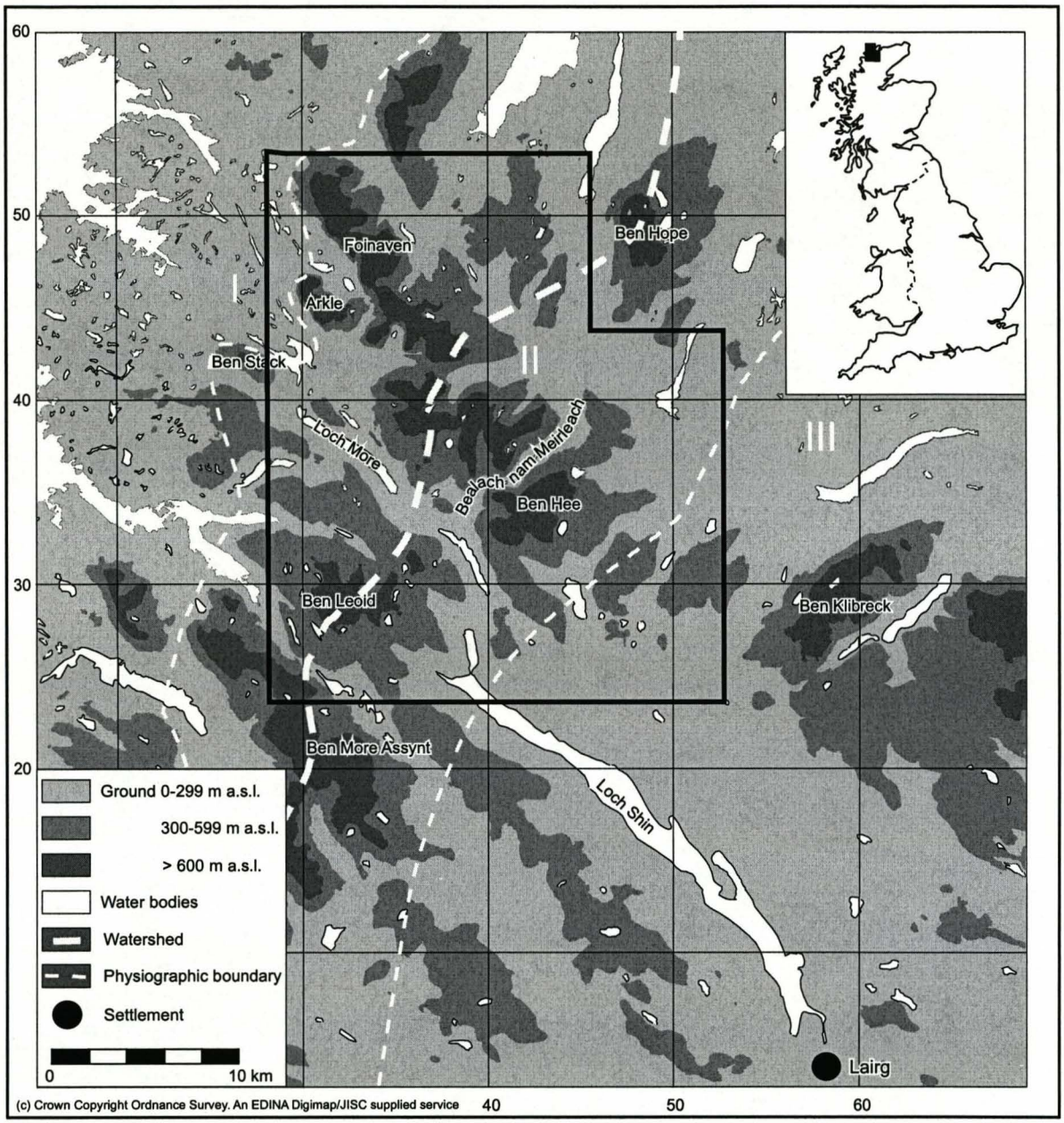


Fig. 2.2 Location and principal topography of the study area. The location of the watershed follows Johnstone and Mykura (1989), the approximate boundary of the study area is shown by black frame. For a full explanation of physiographic regions I-III see text.

2.2.2.2 Geological structure

Structurally, the western, north-western and northern surroundings of the study area are dominated by the Moine Thrust Zone (Fig. 2.3), which comprises a series of low-angle thrust planes that dip towards the east-south-east (Johnstone and Mykura, 1989). From east to west, these are the Moine, Ben More, Glencoul and Sole Thrust Planes. These thrust planes delineate several tectonic nappes that were

displaced in an ESE-WNW direction during the Scandian Orogeny. During compression, the individual nappes are thought to have been stacked on top of each other in a piggyback fashion, so that the Sole Thrust represents the youngest and the Moine Thrust the oldest in this sequence (Harris and Johnson, 1991; Strachan *et al.*, 2002). The cessation of tectonic activity associated with the Scandian Orogeny has been dated radiometrically to 435-425 Ma (van Breemen *et al.*, 1979; Johnson *et al.*, 1985; Freeman *et al.*, 1998).

The influence of the structural "grain" of the underlying geology is reflected in the SW-NE orientation of many lochs (parallel to the Moine Thrust) and the SE-NW orientation of many valleys and lochans (perpendicular to the thrust planes) (Fig. 2.2).

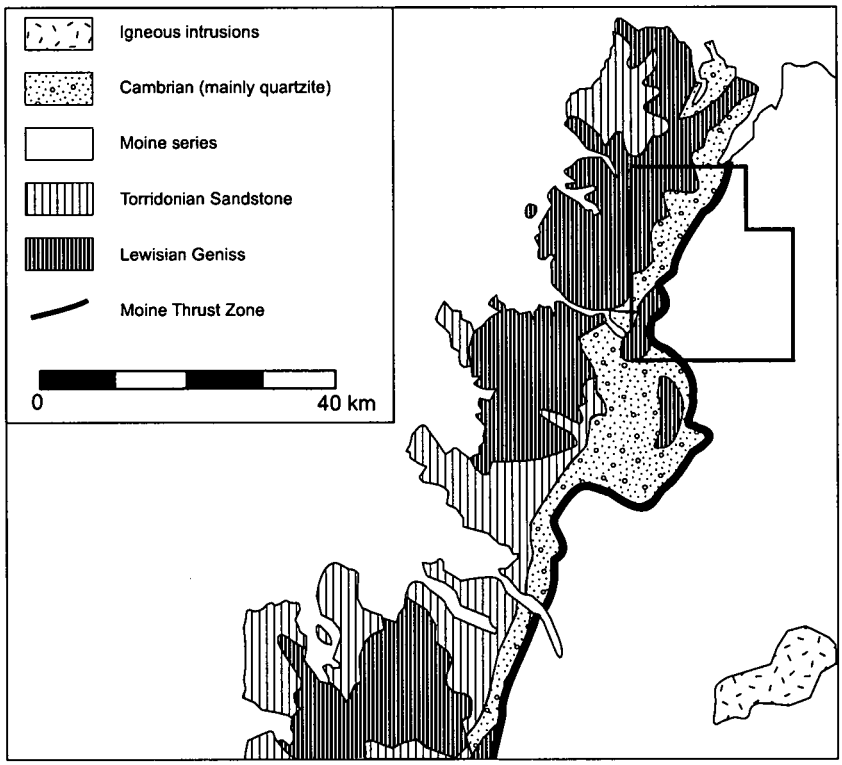


Fig. 2.3 Geological units and tectonic elements in parts of northern Scotland (modified from various sources). Location of the study area is shown by the black frame.

2.2.3 Quaternary geology

2.2.3.1 Introduction

Previous work carried out in the immediate study area is sparse, and the area can be regarded as a classical case of an under-researched area which is in stark contrast to other areas in Scotland that have repeatedly been investigated. As a result, the present thesis represents a necessary selection of possible research questions that could be followed in a number of PhD projects. Due to the scarcity of previous research within the central mountains, including for example Ben Hee and Carn Dearg,

research carried out in the wider surroundings, e.g. the western coastal strip including Canisp, Ben More Assynt and the Knock and Lochan terrain, will be included in the review below.

2.2.3.2 Early and Middle Quaternary

As discussed above, the erosive activity of the late Pleistocene ice sheets has left little sedimentological evidence of early glaciations. This is especially true in the North-West Highlands, where large areas are dominated by bedrock at or near the surface. Due to the presence of limestone caves west the study area around Inchnadamph, however, the timing of events pre-dating the Devensian glaciation of the area is comparatively well constrained. Temperate, ice-free episodes favourable for spelaeothem growth have been dated to 190-180 cal ka BP using the Thorium-Uranium disequilibrium technique (Atkinson *et al.*, 1987; Lawson and Atkinson, 1995). Apart from the spelaeothem record there is no evidence for events predating the Late Quaternary.

2.2.3.3 Late Quaternary

Early Late Quaternary

Additional spelaeothem dates indicate further periods around 140 cal ka BP, 122 ± 12 cal ka BP and 95-56 cal ka BP which also appear to have been ice-free (Lawson and Atkinson, 1995). The latter date is of particular importance as this limits the possibility of a prolonged built-up of the last ice sheet throughout Devensian times as advocated by Bowen *et al.* (2002) (cf. Hall *et al.*, 2003).

Lawson and Atkinson (1995) also reported four further dates that fall into the period between 38 and 26 cal ka BP and indicate groundwater recharge during a Middle Devensian ("Denekamp") Interstadial that is widely recognised in Scotland and elsewhere in Europe (e.g. Whittington and Hall, 2002). In addition, seven radiocarbon dates from reindeer antlers indicate that these animals roamed the surroundings of the cave between 48 - 43 and 32 - 22 ^{14}C ka BP which in turn suggests at least ice-free conditions favourable for reindeer calving in the caves (Murray *et al.*, 1993). Contrasting with the relatively long record contained within the caves, knowledge obtained from the surface terrestrial record is somewhat sparser and largely relates to the Late Devensian glaciation, renewed growth of glaciers during the Younger Dryas and palaeoenvironmental change during the Holocene, all of which will be reviewed below.

Late Devensian (Dimlington Stadial)

Although much of the earliest work in the NW Highlands by the officers of the British Geological Survey (Peach *et al.*, 1907) largely focuses on the solid geology of the Moine Thrust Zone, Read *et al.* (1926) and Read (1931) described and interpreted the Quaternary geology in more detail. According to these authors, and in agreement with the general perception of events at that time (cf. Barrow *et al.*, 1913; Charlesworth, 1955), four phases of glaciation were identified in Sutherland: (1) a period of

"maximum glaciation" when the whole landscape was covered by an extensive ice sheet; (2) a period of "waning glaciation", i.e. when local centres of ice dispersal such as the Ben More Assynt massif became more important and topography gained an increasing control on the patterns of ice flow; (3) a period when the ice became confined to the valleys as the main watersheds emerged from the ice and (4) a final episode when the last vestiges of ice retreated back into corries before melting completely. Using evidence provided by striae and boulder trains, Read *et al.* (1926) and Read (1931) placed the location of the former ice divide over the Ben More Assynt massif. From there, flow was initially radially outwards towards the WNW west of the ice divide, and towards the north and NNE east of the ice divide. Striae with a southerly orientation parallel to larger lochs (e.g. Loch Shin) were attributed to a later phase when topographic control on ice flow increased. An absolute chronology was not assigned to any of these phases.

It is noteworthy that Read (1931: 210) attributes the "tumultuous morainic mounds" to the period of valley-glaciation. He described their sedimentology (p. 213) as "a loose sandy or gravelly deposit typically seen in road-metal pits [...] where it shows roughly equal proportions of sandy matrix and angular boulders of local origin. Often there is a rude stratification, and lenticular masses of sand and gravel are common, whilst, at a few localities, thin beds of fine clay have been recorded."

Charlesworth (1955) mapped moraines and inferred glacier limits in the study area, and elsewhere in Scotland, to detect the accumulation centres of the last ice sheet as it disintegrated. According to his reconstruction, the ice sheet was sourced along the watershed and retreated into the corries of Ben Hope and the central mountain chain (i.e. north-east of Sabhal Beag and Sabhal Mor, Carn an Tionail, Carn Dearg, north and west of Ben Hee; p. 844) plus additional subcentres on Arkle, Foinaven and Meallan Liath Coire Mhic Dhughaill (p. 861f.) and around Ben Leoid and Ben More Assynt (pp. 851, 853). Unfortunately, he did not describe detailed field evidence so that his reconstruction appears rather descriptive and the basis of it untestable. The "50 and 100 foot raised beaches" that he used in his correlations of the larger readvance periods have never been reported in memoirs of the present study area either, thus making it difficult to test Charlesworth's (1955) reconstructions.

Lawson (1990) used the evidence provided by erratic boulder trains to infer the former flow directions of the last ice sheet west of the present study area around Ben More Assynt and Quinag. He later (1996) added the results of orientation measurements of striae in the same area to determine the direction of former ice flow more precisely. He found that most of the striae west of Ben More Assynt are aligned east-west with a few exceptions around the mountain's southern side where striae indicate ice movement towards the south. A second group of striae around Quinag shows cross-cutting relationships, the striae indicating ice flow towards the NNW, west and north while those SE of Glas Bheinn are orientated WSW and south to SSE. The latter orientation is explained by the location of striae in the "shadow" of the general westward ice flow that prevented erosion of this set of striae. As

the distribution of erratics is "unhelpful" in this location, Lawson (1996: 63) regards this set of striae as an inherited ice flow pattern derived from an early phase of glaciation. Lawson explained the pattern of striae around Quinag by arguing that during the build-up of the last ice sheet the ice was restricted to the corries on the eastern side of Quinag and that it flowed eastwards. Subsequently, the flow direction changed to a general westward ice flow, with the main ice source being located on Ben More Assynt. The striae showing ice movement towards the northwest and north are attributed to the LGM or some subsequent deglacial phase (Lawson, 1996).

In general, results obtained from striae and erratic boulder trains appear to agree well, indicating ice flow in a westerly direction, which corresponds well with the location of the ice divide that had been inferred by Peach and Horne (1892) to lie slightly east of the chain of mountains extending from Ben More Assynt to the north. Sutherland (1984a: Fig. 13) also suggested that the ice divide of the last ice sheet roughly follows the main watershed (Fig. 2.2). Lawson (1983), on the other hand, states that its location was always east of the main watershed, a reconstruction based on his extensive study of erratic trains and striae west of the main watershed.

The altitude of the Late Devensian ice sheet in the NW Highlands has recently been reconstructed using trimline evidence. Data for Sutherland are reported in McCarroll *et al.* (1995) and Ballantyne *et al.* (1998b). Table 2.3 gives the altitudes of periglacial trimlines on mountains within and in the immediate surroundings of the study area as reported in these papers. This reconstruction, based on a large amount of fieldwork supplemented by geochemical and geophysical laboratory analyses, however, has not gone undisputed. Bradwell and Krabbendam (2003a, b), using the same approach advocated by Ballantyne (1998a, b) on Stac Pollaidh, an isolated mountain on the west coast not investigated by Ballantyne and co-workers, discovered that a distinct weathering limit occurs below the previously-reconstructed regional ice sheet surface. Thus, Bradwell and Krabbendam (2003a, b) argued that Stac Pollaidh could represent a Late Devensian nunatak, and, if true, this would increase the local ice surface gradient significantly. Ballantyne (2003) argued that the observations of Bradwell and Krabbendam (2003a) are of purely morphological nature and contradict the findings of Ballantyne *et al.* (1998b) which are based on a larger data set involving a number of methods and spatially consistent. However, it should be noted that in their study of trimline altitudes, McCarroll *et al.* (1995) and Ballantyne *et al.* (1998b) visited selected, often isolated mountains in the present study area and that the intervening ground, which hosts a number of peaks and ridges oblique to former ice flow, was neglected.

Recent mapping in Assynt has resulted in the discovery of deep grooves in bedrock, interpreted as evidence for subglacial fluvial erosion related to channelised fast ice flow of a land-based ice stream of the last British-Irish Ice Sheet (Bradwell, 2005). This ice stream, termed the Minch Ice Stream, appears to have been a spatially consistent feature as the terrestrial evidence links well with geophysical marine data from the Minch off the NW coast of Scotland (Stoker and Bradwell,

2005) indicating that the last ice sheet was highly dynamic and similar to modern-day ice sheets in which ice streams are important elements of mass transfer (e.g. Stokes and Clark, 2001; Bennett, 2003). Undoubtedly, further work is needed to elucidate the dynamics and extent of the last ice sheet (cf. Evans *et al.*, 2005).

Table 2.3 Altitudes of weathering limits interpreted as periglacial trimlines on mountains within the study area (from McCarroll *et al.*, 1995 (M); Ballantyne *et al.*, 1998b (B)).

Mountain name	Grid reference (NC)	Trimline altitude	Source
Beinn Spionnaidh	362 573	565	M, B
Cranstackie	351 556	580	M, B
Ben Hope	478 502	575-670	B
Foinaven	317 508	620-650	B
Arkle	310 453	600-720	M, B
Meall Horn	353 449	630-690	B
Meallan Liath Coire Mhic Dhughaill	358 392	600-750	B
Ben Hee	426 341	680-750	B

Younger Dryas

Much of the work carried out in the 1970s and early 1980s focussed on the reconstruction of the Loch Lomond Stadial or Younger Dryas (e.g. Sissons, 1967, 1974, 1979a, b). As part of a wider programme of mapping the distribution of Younger Dryas glaciers, Sissons (1977) attempted to reconstruct their extent in the NW Highlands using geomorphological evidence (Fig. 2.4). He did not, however, investigate some of the features that Read *et al.* (1926) and Read (1931) had described earlier. He reconstructed 17 smaller valley and corrie glaciers in the present study area, which are between 0.4 and 7 km in length plus six smaller glaciers in Assynt, to the west of the study area. The distribution of these glaciers is shown in Fig. 2.4. It is noteworthy that the corries that source the reconstructed glaciers all have an easterly or north-easterly aspect apart from the glacier occupying Glen Oykel, which is, in comparison to the other reconstructed glaciers in the immediate vicinity, disproportionally large. One small niche glacier on Meallan Liath, a southern tributary ridge of the Ben Hee Massif, occurs in an unusual location on a uniform slope without, it seems, a suitable accumulation area. The reconstructed ELAs for the study area vary from as low as 225 m for part of one glacier (northern part of number 4) to much higher values of 570 m for one of the Ben Hope corrie glaciers (number 16) over very short distances (Sissons, 1977). The landforms were assigned to the Younger Dryas on the basis of radiocarbon dates obtained from basal layers of kettle holes and lochs outside the inferred glacial limits (see below). These indicate ice-free conditions from ca. 13 ^{14}C ka BP (e.g. Pennington *et al.*, 1972; Pennington, 1977; Birks, 1984).

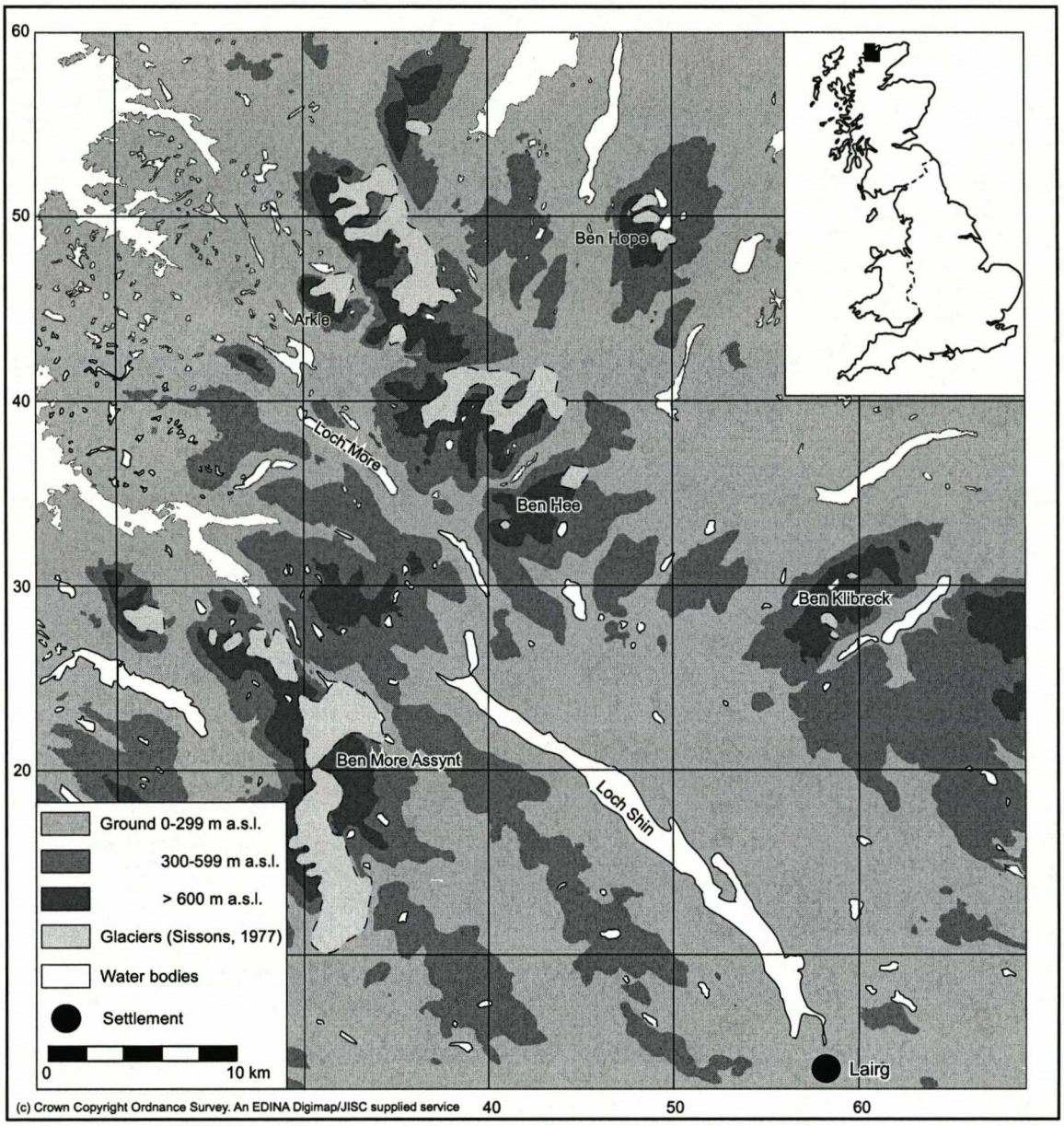


Fig. 2.4 Distribution of Younger Dryas glaciers in northern Scotland as reconstructed by Sissons (1977).

Following on from the work of Sissons (1977), Lawson (1986) amended Sissons' limits very slightly increasing the size of three former glaciers around Ben More Assynt and adding one on the eastern flank of Canisp. He used the same geomorphological approach as Sissons (1977), but additional dating techniques were not applied to confirm Younger Dryas ages of individual glaciers or indeed to reconstruct additional ice masses.

Lateglacial environmental change has been reconstructed using palynological, plant macrofossil and chemical evidence from sediment cores taken from lochs in the North-West Highlands. The lochs studied are mostly located along the west and north coast (Pennington *et al.*, 1972; Pennington, 1977; Birks, 1984, 2003) and Caithness (Charman, 1994). In the present study area,

two lochs were investigated, Lochan an Smuraich (NC 300 418) and Loch Stack (NC 298 423; Fig. 3.1; Pennington, 1977). Both these lochs were reported to contain full lateglacial chronologies, and in the case of Loch Stack, the sequence was referred to as being "similar to that found in Windermere" (Pennington, 1977: 122). Pennington (1977: 121) added that the sediments contained evidence of glacier presence in the Loch Stack catchment during the Younger Dryas, based on "laminated post-interstadial" glaciolacustrine sediments in the cores (p. 139). From the presence of these laminated "varved sediments", as opposed to sand and gravel devoid of pollen found in other lochs, Pennington (1977: 141) inferred that Loch Stack was "just outside the ice limits". The evidence presented by Pennington (1977) conflicts that of Sissons (1977) though, since he did not reconstruct glaciers in the Loch Stack catchment. Unfortunately, the Loch Stack data were never published in detail, and apart from the brief notes on Loch Stack and pollen data on Lochan an Smuraich, little is known about the timing of events within the present study area. However, the location of the lochs along the north and west coast and their containing full lateglacial sediment sequences indicates that deposition within these lochs commenced before the Windermere Interstadial and that these sites therefore remained ice-free during the Younger Dryas (Gray, 1997).

Holocene

The literature review in this section is kept brief and concentrates on those features and processes that are relevant to the mapping of the glacial geology and geomorphology. For further, more detailed information the reader is referred to the cited literature.

Following glacier retreat at the end of the Younger Dryas, glaciers in Scotland are thought to have melted away by, or shortly after, the beginning of the Holocene (Chapter 1.1.1.1), and no evidence has been reported from the present study area to suggest the contrary. Holocene geomorphological activity was thus characterised by high-level, small-scale microgelivation as well as frost-sorting and solifluction processes (Ballantyne, 1991; Ballantyne and Harris, 1994). The latter have been investigated in detail on Arkle (NC 312 452) by Mottershead and White (1969) and White and Mottershead (1972) who described former solifluction processes that were later dated to between 5441±55 to 3984±50 ^{14}C a BP (Mottershead, 1978). These authors interpreted the thick accumulation of mountaintop detritus near the summit as a relic from Lateglacial times and concluded that increased mid to late Holocene solifluction rates that were triggered by climate deterioration had modified its surface significantly. In comparison to these high solifluction rates Mottershead (1978) noted that modern processes appeared to have slowed down considerably. At other mountaintops, such as on Ben More Coigach (NC 094 042), windblown sands accumulated in leeside localities, due to either disturbance of the protective vegetation cover through animal grazing or, more likely, triggered by widespread climatic deterioration (Morrocco, 2003).

Those slopes that have partly been oversteepened during ice sheet glaciation responded by rock-slope failure (RSF) as an effect of debuttressing, five cases of which are known to occur in the present study area. One on Meall a' Chlerich (NC 409 366) represents an arrested slide and is currently being dated by cosmogenic radionuclide methods (C.K. Ballantyne, pers. comm., 2003) while three RSFs are recorded to occur on the Ben Hee Massif. The largest of these is located along the headwall rim of An Gorm Coire (NC 434 343) and records a multi-staged event (Jarman and Lukas, 2005). Another case study of a complex RSF has been reported by Sellier and Lawson (1998) from Assynt. In addition to these large-scale processes, paraglacial adjustment of oversteepened sediment covers on slopes, known to have occurred on a smaller scale elsewhere in the Scottish Highlands (e.g. Curry, 2000; Ballantyne, 2002b), are likely to have affected the study area as well.

Peat, which is ubiquitously found throughout the Scottish Highlands, is particularly thick on plateaux and interfluves in eastern Sutherland due to the base-poor Moine lithologies and the high precipitation (Lindsay *et al.*, 1988) and can provide important insights into Holocene climate change. Evidence for climate change obtained from spelaeothems in Assynt (Proctor *et al.*, 2000, 2002), lake sediments along the west coast (Moar, 1969; Pennington *et al.*, 1972; Pennington, 1977) and peat profiles in eastern Sutherland (Charman, 1994; Charman *et al.*, 2001) indicate wetter climate conditions favourable of peat growth during the mid and late Holocene.

CHAPTER 3 GLACIER RECONSTRUCTION

3.1 Introduction

As mentioned in the foregoing chapter, little previous work has been carried out in the present study area, and knowledge about past glaciations, their extent and palaeoclimatic significance is very limited. Recent advances in the reconstruction of palaeo-glaciers elsewhere in Scotland have led to a better understanding of the interactions between atmosphere and cryosphere in Lateglacial times (e.g. Ballantyne, 2002a; Benn and Ballantyne, 2005). Based on this recent work, detailed geomorphological mapping was utilised with the aim of reconstructing former glacial events in the present study area, to compare these results to former reconstructions in this area (Sissons, 1977) and to allow comparison with other sites in Scotland. This chapter will (a) discuss the problems encountered in establishing a chronological basis and the solutions that were found, (b) present the evidence used to reconstruct the extent of former glaciers in the far NW Highlands of Scotland, (c) address the palaeoclimatic implications of these findings and (d) compare this to previous findings elsewhere in Scotland.

3.2 Dating and chronology

3.2.1 Coring and radiocarbon dating

3.2.1.1 Introduction and coring site characteristics

Dating of glacier limits is crucial to understand the timing of individual glacier events and to establish links with the local, regional and global palaeoclimate. In Scotland, the dating of distinct glacier limits has traditionally been achieved by radiocarbon assay of organic layers obtained from kettle hole infills or lake basins inside and outside the limit by coring (e.g. Pennington, 1977; Walker *et al.*, 1988; Benn *et al.*, 1992). In this chapter, the attempts of constraining glacier limits and the problems encountered will be discussed and possible solutions suggested.

In contrast to other sites in the Scottish Highlands, the bedrock is at or near the surface in ca. 70-80% of the present study area, and enclosed, infilled basins are rare. The traditional approach of constraining glacier limits in a number of places is thus rendered almost impossible. Coring of lochs, of which there are plenty in the study area, was deemed a major logistical exercise and not feasible within the course of the present PhD project. In the whole study area, only one location suitable for coring could be found.

This site is located at the transition from Loch More to Loch Stack (NC 276 408; Fig. 3.1) and is colonised by sedges. To the south of the site, Loch Stack is separated from Loch More by a bedrock bar of Cambrian Quartzite. This rock bar is traversed by a moraine ridge which is ca. 2.5 m high and from which a bedrock bench overlain by peat slopes northwards towards the coring site (Fig. 3.1). South and east of this moraine, several recessional moraine fragments can be found trending obliquely

up the slopes, recording progressive glacier retreat southward along the axis of Loch More and southeast towards Meallan Liath Coire Mhic Dougaill (Appendix 2; Chapter 3.3.2.2). The southern part of the site is traversed by a river that meanders through the area that has silted up the connection between the two lochs. Together, these landform relationships suggest that, instead of an infilled, enclosed basin (a closed system), this basin is an open system. The presence of a terrace form sloping away from the moraine perched on top of the bedrock bar suggests the presence of a delta system at this site. If this was the case, then the thickness of individual lithofacies units would be expected to increase with distance from the moraine, and this will be tested below (Chapter 3.2.1.3).

3.2.1.2 Methods

Coring was undertaken by hand using a Russian corer. Initially, test cores with chamber diameters of 0.05 m and lengths of 0.5 m were brought down at location LST 1 (Fig. 3.1) to identify the stratigraphic sequence at this location and test for the presence of glaciolacustrine and intercalated, dateable layers. Once the sequence had been explored, a larger chamber (0.15 m width and 0.5 m length) was used for sampling. Consecutive segments with a 0.2 m overlap were recovered at depths of 3.46 m to 1.90 m to sample the transitions between minerogenic and organic layers found within the sequence. These segments were individually photographed, described, then extruded in the field and wrapped in plastic film and aluminium foil following the approach described by Birks (1984). The cores were then stored in segments of drainpipe in a cold room under constant temperature conditions at 4°C.

Sampling for accelerator mass spectrometry (AMS) radiocarbon dating was undertaken in a clean room to minimise contamination. The respective parts of the core were cleaned with a fresh razor blade to avoid down-core contamination and samples obtained from ca. 1 cm below the surface. Samples were placed on aluminium foil, weighed, wrapped, labelled and stored in airtight plastic bags before being sent away for dating.

3.2.1.3 Core lithostratigraphy

The lowermost 2.9 m of Core LST 1 consist of thixotrophic, blue-green-grey laminated silts with individual laminae-thicknesses between 1-5 mm followed by a transition zone of 0.24 m with a marked increase in thin organic layers between 1-2 mm thick (Fig. 3.2). This is overlain by an organic gyttja 0.25 m thick followed by grey-green silts and blue clay 0.74 m thick which contains a high proportion of plant macrofossils, identified as sedge (G.W. Whittington, pers. comm., 2003). The top of the core is marked by a gradual transition from silt and clay to an organic gyttja that in turn grades into humified peat. This last transition, which occurs at ca. 0.7 m depth, was not sampled.

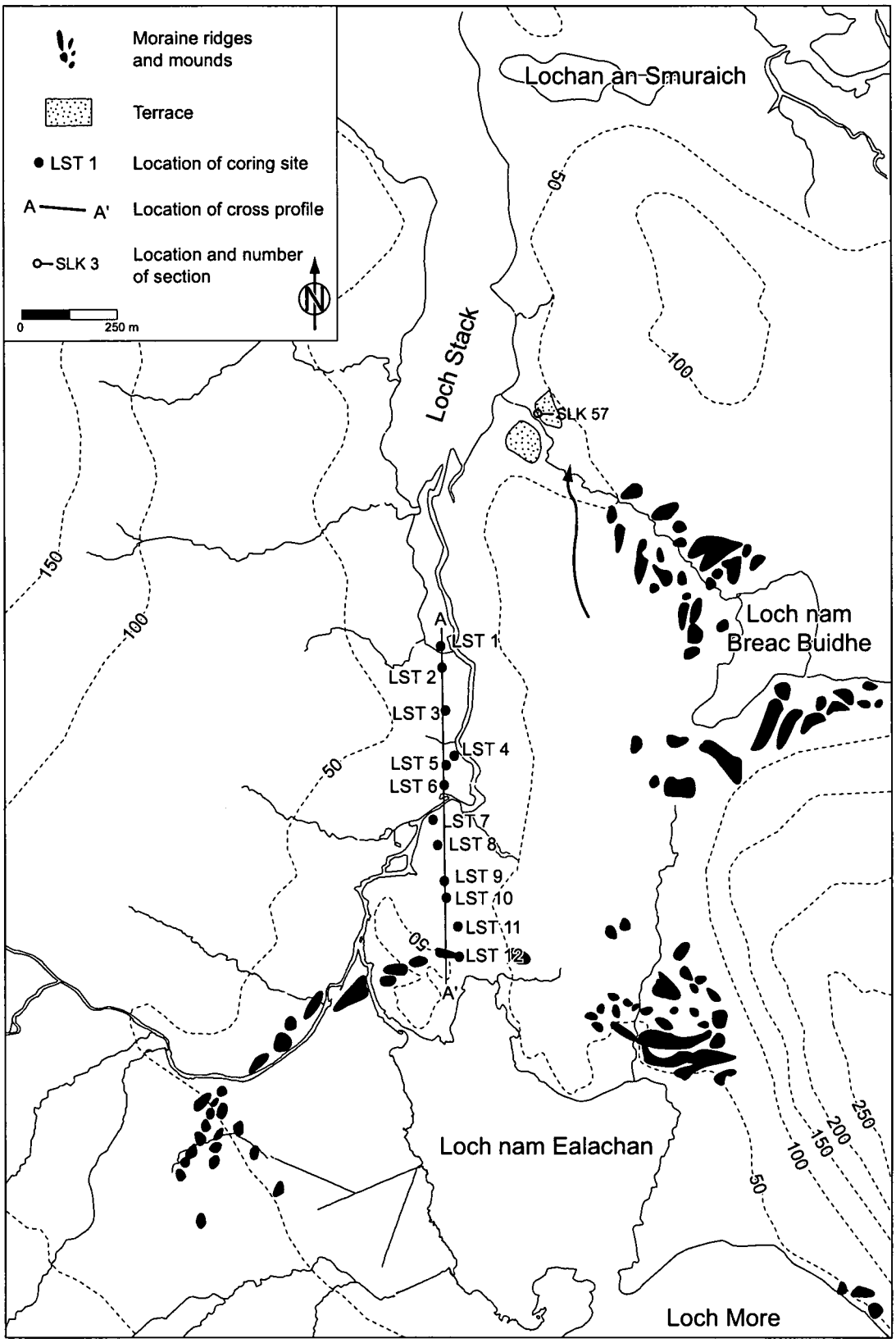


Fig. 3.1 Detailed map showing the location of cores and landscape elements of the open delta system between Loch More and Loch Stack. NB: A full lateglacial, tripartite sequence was recovered from Lochan an Smuraich, in the NE corner of the map, and Loch Stack by Pennington (1977). The location of section SLK 57 is also shown.

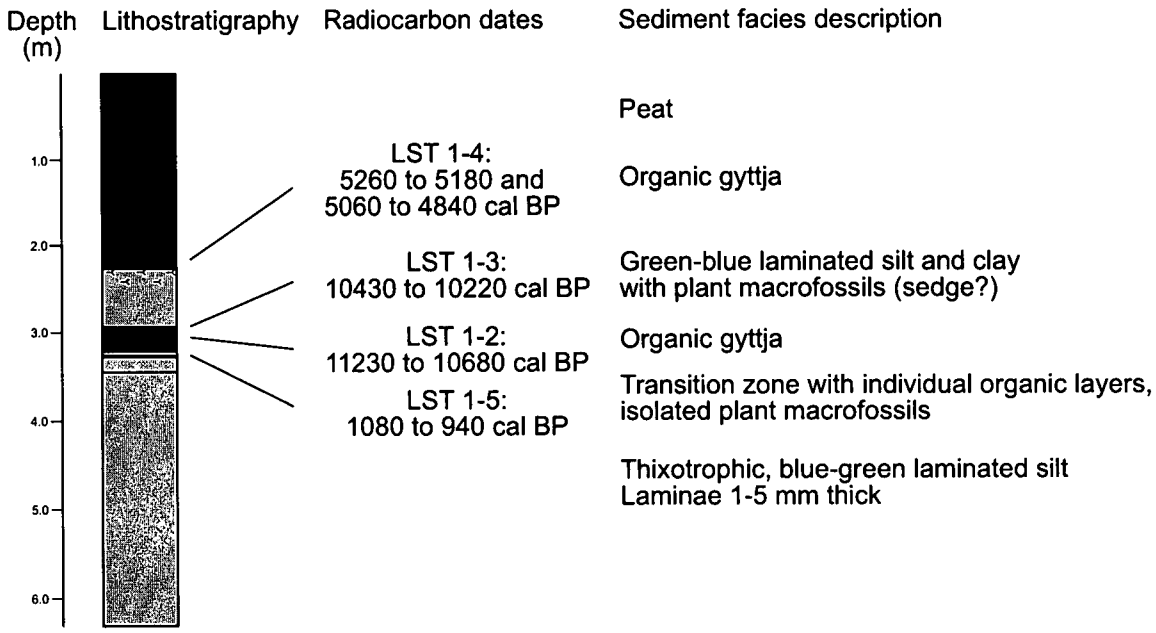


Fig. 3.2 Composite core log of LST 1 showing the location of ^{14}C-dates and description of lithofacies.

This lithostratigraphy was confirmed in a subsequent coring campaign in February 2004 when cores LST 2 to LST 12 were recovered to establish the lateral variation at the site (Figs. 3.1, 3.3). The lithofacies in cores LST 1 to LST 5 can be correlated across the basin. This relationship breaks down for the remaining cores (LST 6 to 12) where only thin ($\leq$ 2 cm) sand and gravel layers occur in peat; the transition to underlying gyttja could not be recovered in any of these cores. These occasional gravel layers are interpreted as overbank deposits formed during flood events of the nearby river. Two such layers also occur in LST 4 (Fig. 3.3). This indicates that, where the river enters the middle part of the basin and the sand and gravel horizons are most prominent, fluvial reworking has taken place. However, the basin stratigraphy also indicates that, although variations in unit thicknesses occur, an increasing trend in thickness with distance from the moraine can be observed. This is interpreted as a confirmation of the initial hypothesis of this locality in terms of an open system.

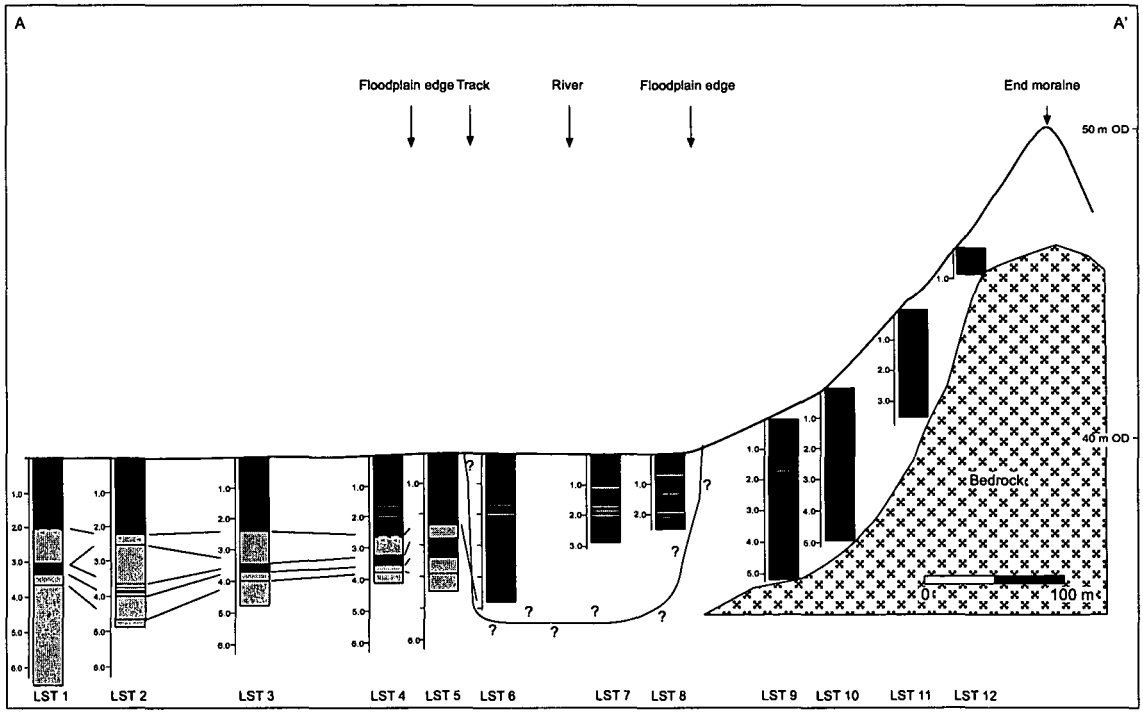


Fig. 3.3 Profile of sediment succession in multiple cores recovered along a transect (A-A', Fig. 3.1) showing the lateral variation of thicknesses of individual layers. Vertical exaggeration of profile is ca. 9 times. Grayscale codes of cores are the same as those employed in Fig. 3.2. For a full explanation see text.

3.2.1.4 Radiocarbon dates and discussion

Four AMS radiocarbon dates were obtained from core LST 1, details of which are listed in Table 3.1. The lowermost date, LST 1-1, taken from the lowermost plant macrofossil found in the core, yielded too little final carbon for analysis and was thus replaced by sample LST 1-5, which in turn gave a calibrated age of 1080 to 940 cal a BP. LST 1-2, obtained from the transition between thixotrophic silt and clay and overlying gyttja, was dated to 11230 to 10680 cal a BP, LST 1-3, from the middle of the lower gyttja, gave a calibrated age of 10430 to 10220 cal a BP and LST 1-4, taken from the bottom of the uppermost gyttja, intercepted the calibration curve twice and thus gave calibrated 2σ-ages of 5260 to 5180 and 5060 to 4940 cal a BP. These ages, apart from LST 1-5, are all in stratigraphic order. LST 1-2 closely resembles the date of end of the Younger Dryas as recorded elsewhere in Scotland and in the GRIP cores (cf. Björck *et al.*, 1998; Brooks and Birks, 2000). LST 1-1 yielded too little carbon in the first batch of samples submitted to the dating laboratory while LST 1-2 to 1-4 all yielded enough final carbon for AMS analysis. Therefore, new plant macrofossil material was required as a substitute for LST 1-1. However, the date obtained from this substituted material (LST 1-5) gave a date which causes a strong reversal of dates in the lowermost part of the core (Table 3.1; Fig. 3.2). Only this date which was obtained later from material that had been stored for some time (LST 1-5) is out of sequence, suggesting that contamination has occurred. Possible sources of such contamination could be by dragging younger material into the borehole whilst inserting the Russian corer or by

secondary growth of microorganisms or fungi on macrofossils in the stored samples (cf. Wohlfarth *et al.*, 1998). Consequently, this date is rejected.

Table 3.1 Details of radiocarbon samples obtained from core LST 1.

Sample ID	Laboratory ID	Sample depth (cm) and material description	$^{13}C/^{12}C$ ratio	Conventional age (a BP)	Calibrated 2σ age (cal a BP)
LST 1-4	Beta-189453	221; gyttja	-28.1‰	4370±60	5260 to 5180 and 5060 to 4940
LST 1-3	Beta-189452	308; gyttja	-23.6‰	9160±50	10430 to 10220
LST 1-2	Beta-189451	303-300; gyttja	-23.7‰	9660±110	11230 to 10680
LST 1-5*	Beta-190643	311-317; sedge stem	-25.7‰	1110±40	1080 to 940

*LST 1-5 replaced the original LST 1-1 (Beta-189450) as this yielded too little final carbon.

3.2.2 Optically-stimulated luminescence (OSL) dating

In order to overcome the shortcomings imposed by a lack of basins suitable for coring, optically-stimulated luminescence (OSL) dating of glacigenic sediments was attempted. Sampling, sample preparation and analysis followed procedures outlined by Aitken (1998), Murray and Wintle (2000, 2003) and Preusser (2003). Use of this method was initially hoped to alleviate the problems imposed by the paucity of sites suitable for dating with more traditional methods. The application of OSL techniques was particularly promising since the moraines in the study area contain numerous glaciofluvial layers that were deposited subaerially in shallow rills or sheetflows, thereby implying that zeroing issues would be less problematic, at least theoretically (cf. Wallinga, 2002). In addition, dating of glacigenic sediments from similar depositional settings had been previously successful (e.g. Richards *et al.*, 2000; Spencer and Owen, 2004), reducing limitations imposed by the sedimentary environment.

However, complex luminescence characteristics and poor bleaching of both quartz and K-rich feldspars were a significant barrier to the successful dating of these samples and neither larger single aliquots nor single-grain dating approaches could overcome these difficulties. Such substantial problems are rarely encountered at once in OSL dating (J.Q.G. Spencer, R.A.J. Robinson, A. Lang, F. Preusser, pers. comm., 2004-2005) and appear to be related to inherent sensitivity and inheritance problems in both quartz and K-rich feldspar grains. These are most likely associated with poor bleaching, bedrock lithology or a lack of re-working prior to deposition. As a result of aforementioned problems, dateable signals could not be obtained, and the samples have to be deemed un-dateable using present luminescence methods. A detailed description of the approaches and problems is given elsewhere (Lukas *et al.*, 2005b).

3.2.3 Synthesis of chronological data and their significance for the timing of glaciation

The large thickness of glaciolacustrine silt and clay (> 3 m) and the radiocarbon dates obtained from above this unit in core LST 1 (Fig. 3.2) strongly suggest that deposition at the site was directly influenced by glacial meltwaters at the maximum extent of local glaciers at this site and probably only for a short time during retreat. The basins of Loch Stack and Loch More are separated by a bedrock bar which would have prevented suspended sediment carried in glacial meltwater to enter the Loch Stack system once the glacier had retreated to south of this barrier (cf. Dahl *et al.*, 2003, Bakke *et al.*, 2005). Thus, the dates constrain the age of the glacier limit just south of the site to the maximum position of the Younger Dryas glacier in the Loch More basin. This interpretation is based on the lowermost date of 11230 to 10680 cal a BP which marks the onset of organic accumulation in the basin following glaciolacustrine conditions. This date closely resembles the end of the Younger Dryas at 11500 cal a BP as recognised elsewhere in Scotland (cf. Brooks and Birks, 2000). Loch Stack and Lochan an Smuraich, both just north of the coring site (Fig. 3.1), contain full lateglacial, tripartite sequences, and the core stratigraphy demonstrates that Younger Dryas glaciers influenced lake sedimentation in the catchment of Loch Stack (Pennington, 1977; Chapter 2.2.3.3) thereby giving further credibility to the glacier limit north of Loch More representing the Younger Dryas maximum in this area.

Based on the two dated sites from the literature and the data presented above, a Younger Dryas age of the glacial limit in the northern part of Loch More can be established with confidence. Because of the scarcity of additional sites however, extrapolation of this evidence is necessary. The reconstruction of glacier limits in the study area will be carried out using the morphostratigraphical approach developed and discussed in Chapter 2.1.3.2 from well-dated evidence for a Younger Dryas landsystem elsewhere in Scotland.

3.3 Glacier limits and reconstruction

3.3.1 Introduction

Glacier reconstruction usually deals with establishing the extent of ice masses at any given period for which a clear geomorphological record is preserved. As such, it forms the basis for the determination of palaeoglaciological and palaeoclimatic variables such as the equilibrium-line altitude (ELA) and palaeo-precipitation values which are needed for a meaningful understanding of the palaeo-glacier system (e.g. Sissons, 1974, 1976; Sutherland, 1984b; Ballantyne, 1989, 2002a; Sharp *et al.*, 1989; Benn *et al.*, 1992). Only if such sound knowledge has been extracted from field data can numerical modelling be tested against this evidence (e.g. Hubbard, 1999; Winkler and Haakensen, 1999; Golledge and Hubbard, 2005), and thus it is crucial to constrain the dimensions of a former ice mass as tightly as possible. For this purpose, the morphostratigraphic approach outlined in Chapter 2.1.3.2 will

be used. In addition, directional indicators will be used to reconstruct the former ice flowlines of individual glaciers of the ice cap and its surface contours, which are at right angles to the flowlines (cf. Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005).

Striae are rarely preserved on Moine lithologies, so the orientation of the lee faces of roches moutonnées was used as an additional source of information on ice flow direction, although data quality is influenced by bedrock structure, joint orientation, depth and spacing (Gordon, 1981; Rastas and Seppälä, 1981; Glasser and Bennett, 2004). It should be borne in mind that the orientation of roches moutonnées is an approximation of the true ice flow direction and that deviations from the measured direction can be as large as 15° (cf. Flint, 1971; Rea *et al.*, 1999). Striae are only well preserved on Cambrian Quartzite outcrops in the northern sector of the study area (Appendix 2); cross-cutting striae indicating a change in ice flow direction were not encountered in any of these locations.

Likewise, erratic trains have been used in other areas where clear tracer lithologies exist to reconstruct former flowlines (e.g. Benn, 1990, 1992a). Read (1931: 211f.) notes that only very few erratics are present within the area underlain by Moine lithologies, although he states that a few erratics of Lewisian Gneiss and Cambrian Quartzite have been recorded east of the Moine Thrust, indicating transport inland. However, some of the Lewisian "erratics" could indeed represent Lewisianoid lithologies derived from inliers farther inland (Chapter 2.2.2.1). Claims of his findings of erratics derived from west and north of the Moine Thrust that have been transported inland could not be substantiated during field mapping; furthermore, erratics showed a consistent pattern whereby Moine and Cambrian boulders were only found west and north of the Moine Thrust, but never east or south of it. Thus, it appears that the transport of boulders (and thus the inferred ice flow direction) was always radially outwards towards the coast during ice sheet glaciation and deglaciation. Erratic patterns could thus only be utilised in a narrow zone north and west of the Moine Thrust and striae and mainly roches moutonnées had to be relied on.

The reader is referred to Appendix 2, where all the evidence described below is displayed in detail. Appendix 3 contains place names to enable quick reference to individual sites discussed below; these could not be displayed in Appendix 2 for cartographic reasons. Specific reference to these appendices will not be made in the text.

3.3.2 Individual glacier basins

As the field evidence suggests the presence of one large, coherent mountain ice field and one independent corrie glacier complex, the limits will be described and reconstructed for the individual basins with names being given to the respective main glacier (cf. Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005). For the purpose of identification, the mountain ice field is divided into four transection complexes that are described and interpreted in turn. These are (1) the northern transection complex, including the ground between Strath Dionard and Glen Golly, (2) the central transection

complex, including the ground between Loch More and Gobernuisgach Lodge, (3) the southern transection complex, including the ground between Ben Leoid and Ben Hee, (4) the eastern transection complex, including the ground south and east of Ben Hee and (5) the independent corrie glacier complex east of Arkle.

3.3.2.1 The northern transection complex

Dionard Glacier

A conspicuous former glacial limit at the northern end of Strath Dionard is indicated by the presence of a large moraine ridge that alternates with ice-moulded bedrock outcrops and extends across the Strath from southwest to northeast at an altitude of ca. 90 m (NC 345 526). To the north, two to three distinct terrace levels are developed whereas south of the limit only one valley train level can be observed. From this limit, latero-terminal moraines ascend southeastwards along the Strath's eastern slope reaching a maximum altitude of 150 m. These moraines fade out after ca. 2 km and give rise to a valley train and thin, quartzitic scree slopes. The three corries around Foinaven, Glas-Coire Granda (NC 320 514), Bràighe a' Choire Leacaich (NC 325 502) and Coire na Lurgainn (NC 327 492) all contain very clear evidence of former glacier coverage. The ground between the former two is intensely ice-moulded, with the trend of grooves indicating that ice flow was diverted around Cnoc Dùail (NC 335 513). The top of the latter is frost-weathered and was thus most likely not overridden.

A distinct drift limit at the corrie headwall of Bràighe a' Choire Leacaich at ca. 720 m (NC 319 499) enables the delineation of the former glacier surface. The evidence for the maximum vertical extent of the former glaciers is equally clear in neighbouring Coire na Lurgainn, where a trimline at ca. 600 m (NC 329 490) and a drift limit that descends from ca. 480 to 450 m in the lower parts of the corrie (NC 340 489) give a good indication of the location of the former ice surface. The latter drift limit can be linked to lateral moraines on the southern slope of the corrie at its entrance to the Strath, allowing the ice surface to be constrained at ca. 250 m in this location (NC 346 500). Unfortunately, the eastern flank of the Strath does not contain any evidence to constrain the former ice surface, thus necessitating extrapolation over a large distance and correlation across the width of the Strath. However, the evidence in Coire na Lurgainn and that farther upvalley provides ample glacial evidence to ensure an accurate reconstruction of the former glacier dimensions.

Distinct trimlines, marked by a sharp transition from ice-moulded and plucked bedrock to solifluction lobes over a vertical distance of ca. 10 to 20 m, occur on steep slopes north and northeast of Meall Horn (NC 352 453). They descend from ca. 650 m on Meall Horn to about 400 m on the southern flanks of Plàt Reidh (NC 352 468) and eastwards to an altitude of about 600-610 m at the Bealach Eadar na Sabhal (NC 365 439). Just south of the Bealach, the trimline altitude culminates at ca. 620 m, descending again southwards thereby suggesting a connection between the northern and central transection complex (see Seilich Glacier below). East of the Bealach, the trimlines are equally

sharply defined and descend to about 500 m at the junction with the corrie northeast of Sabhal Beag (NC 380 443). Here, frost-weathered roches moutonnées indicate a uniform northeasterly flow direction and show a clear lateral boundary to much less weathered roches moutonnées and ice-moulded bedrock, the orientation of which follows the valley axes (NC 386 435). The former glacier surface is defined by a trimline at ca. 670 m in the northeastern corrie.

The pattern of ice flow can be well reconstructed across the whole area due to the presence of flutes in the valley bottoms and corries and roches moutonnées and ice-moulded bedrock in source areas where bedrock prevails at the surface, respectively. From this evidence it would appear that, in the western parts of the central transection complex, the ice flowed northwards from Meall Horn into Strath Dionard, supplemented by small amounts of ice from Plàt Reidh and being confluent with larger glaciers from the Foinaven corries near the maximum extent. In the eastern parts, ice flowed eastwards towards Glen Golly where the ice split into a northern and southern lobe (see below). It would appear likely that part of Plàt Reidh contributed ice to the Dionard Glacier since the ice surface encroaches onto an outlet bedrock gully of the plateau at an altitude of ca. 300 m (NC 358 487).

Golly and Staonsaid Glaciers

The southeastern margin of the northern transection complex is defined by a distinct terminal moraine arc that obliquely crosses Glen Golly at an altitude of ca. 120 m (NC 428 428). Downvalley of this moraine, three distinct terraces occupy the valley floor whereas upvalley only one river terrace is observed. The terminal moraine grades into pronounced lateral moraines on the southern valley side where they can be traced upvalley to an altitude of ca. 200 m (NC 413 437) where they fade out. No further information from which the former glacier surface could be determined is found on this valley side, and glacier limits have to be interpolated with those at the junction with Coire Beith a' Mheadhoin (NC 399 445). On the northern slopes of Glen Golly, latero-frontal moraines are absent and the glacier limit has to be interpolated to the upslope limit of ice-moulded bedrock at an altitude of ca. 230-250 m (NC 408 458). This limit coincides with the upper boundary of latero-frontal moraines on the slope; however, the glacier most likely extended into the broad valley entrance occupied by the Allt Beith (NC 405 464) as dictated by the surface slope that can be well constrained on the southern slopes of Coire Beith a' Mheadhoin immediately to the west (NC 396 445).

To the north, these limits can be linked to the clear lateral moraines of the Staonsaid Glacier, which is defined by a small but distinct moraine arc that crosses the valley at an altitude of ca. 180 m (NC 384 482). This moraine can be linked to clear lateral moraines that ascend southwards for a distance of ca. 500 m (NC 383 382) before fading out. Again, interpolation between this and the limits on the northern slopes of Coire Beith a' Mheadhoin is required. Some control on the height of the former glacier surface is provided by frost-weathered bedrock on Creag Staonsaid (NC 383 470) but this location arguably represents the most uncertain situation for correlations of glacial and periglacial

evidence across the whole study area. This part of the glacier margin should thus be regarded as tentative at present.

3.3.2.2 The central transection complex

More Glacier

The terminus region around the northern end of Loch More is defined by a low moraine ridge perched on a bedrock bar between Lochs More and Stack from which a bedrock bench slopes northwards. From this limit, lateral moraines ascend on the eastern side of Loch More towards the south, but are obscured by forest and then fade out on the steep slopes of Ben Screavie (NC 316 384), and the first clear evidence of an ice margin is given around NC 303 394 by moraines surrounded by outwash. Evidence for the maximum extent on the western side of Loch More is largely obscured by forests, and only a few lateral moraines could be traced west of Achfary (NC 292 397) connecting up to the aforementioned low ridge to delineate a broad glacier lobe. Only farther south, around NC 343 365, do lateral moraines on either side of Loch More allow a precise control on the glacier surface at its maximum extent to be gained.

The More Glacier lobe was, according to the geomorphological evidence, not confluent with ice flowing northwards out of the valley drained by the Allt a Chuilinn immediately to the east of Loch nam Breac Buidhe (NC 305 406). The present Breac Buidhe basin contains about eight distinct moraine arcs that can be joined to delineate a broad lobe that extended almost down to the shores of Loch Stack (NC 299 412) where the glacier appears to have terminated in a water body. Evidence for this comes from an exposure (SLK 57; Fig. 3.1) that displays conformably northward sloping (~10°) clinoforms with fining upwards sequences and sharp contacts between individual units. Identified lithofacies range from parallel-bedded granules in a medium to coarse-sandy matrix with outsized coarse gravel clasts to parallel-bedded fine to coarse sand with few outsized clasts of granule to medium gravel and thinner layers and discontinuous lenses of granules to smaller pebbles and very fine to coarse sand. The lower part of the section is composed of a ca. 0.4 m-thick clinoform of openwork, rounded pebbles and isolated cobbles. Imbrication was not observed in any of these units. This sedimentary succession is interpreted as a Gilbert-type delta with the clinoforms representing foreset beds (Fyfe, 1990; Benn, 1992b, 1996; Benn and Evans, 1993, 1996; Ashley, 1995; Lukas, 2004b). Section SLK 57 is located in a terrace just downvalley of distinct moraines that indicate ice retreat upvalley towards Loch na Mucnaich (NC 325 390; Fig. 3.1). On the SW side of the small stream draining this valley, a flat-topped sediment accumulation reaches a similar height of ca. 50 m OD (NC 2995 4095) indicating that the delta might be a continuous feature across the lower part of the valley. Unfortunately, the geomorphological context of the terrace containing SLK 57 cannot be established with confidence since much of the original landform shape, especially the surface and former ice-proximal slope, has been altered during gravel extraction. However, due to the

delta being only 150 m downvalley of clear recessional moraines, it is quite likely to represent an ice-contact delta.

To the east of Loch nam Breac Buidhe, the maximum position of the glacier is marked by an arc of distinct moraine ridges (NC 310 414). Within the narrow Eas an Aighe to the south, several smaller moraines and meltwater channels occur (NC 316 401).

Seilich Glacier

To the north of the More Glacier, a distinct glacier limit in the northwestern part of Strath Luib na Seilich (NC 325 413) is defined by exceptionally large moraine ridges that attain heights of 15 m, widths of up to 100 m and lengths of up to 200 m. These can be linked to meltwater channels and moraines in Eas an Aighe (described above). At the maximum extent of this glacier, therefore, Meall a Cheardaich (NC 316 405) acted as a barrier for northward ice flow, thereby splitting a coherent ice mass into one that funnelled through Eas an Aighe and a larger piedmont glacier that spread out over the plain west of Meallan Liath Coire Mhic Dhughaill. The maximum extent of the glaciers in this part is marked by four converging lines of evidence: (a) the clear and large moraines that form an arcuate pattern across Strath and can be linked to evidence farther west; (b) these moraines in turn join up with clear lateral moraines and a marked upslope termination of the sediment cover on the southern slopes of Creachan Thormaid around NC 340 424; (c) the downvalley side of the terminal moraines coincides with a sudden transition from one small river terrace east of NC 324 413 to two distinct higher terrace levels west of that point; and (d) the thickness of the sediment cover in the valley bottoms changes dramatically. Whereas east of the maximum moraines the river has incised into up to 25 m of sediment and thick accumulations of till have been recorded there (SLK 55), sediment thickness is < 5 m west of these.

The vertical limits of the Seilich lobe are marked by the upslope termination of the glacigenic sediment blankets and the transition from ice-moulded bedrock around NC 339 428 to weathered bedrock. This boundary, which occurs at ca. 300 m, can be traced into Coire Gràndа (NC 351 435) where it rises northwards to ca. 450 m. In neighbouring Bealach Eadar da Sabhal (NC 366 439; 563 m), there is a sharp transition between the upper limit of ice-moulded and plucked bedrock and the lower limit of solifluction lobes. These two lines of evidence bracket the altitude of the periglacial trimline in this location to 600-610 m. This trimline rises northwards where it culminates at an altitude of ca. 620 m just south of the Bealach, providing evidence of the junction with the northern transection complex described above. It is at the same altitude on the eastern slopes of Bealach Eadar da Sabhal and can be traced southwards around the southern slopes of Sabhal Beag where it descends to an altitude of ca. 550 m (NC 370 421). The trimline is obscured by rockfall deposits and thick peat in an easterly direction, but can be traced across the col of Bealach na Fèithe (NC 372 419; 452 m)

suggesting that a connection existed between the west-flowing Seilich Glacier and the east-flowing Easaidh Glacier (below).

A similarly-clear vertical limit can be traced around the northern slopes of Meall Garbh (NC 370 412) where solifluction lobes descend to ice-moulded bedrock and roches moutonnées at an altitude of 540 m. This trimline ascends into Blaoch Choire to about 570 m (NC 363 402) before the upper limit of ice-moulded bedrock descends to ca. 270 m on the northern gentler slopes of Sàil Rac around NC 358 412 over a distance of 2.5 km. Along the steep western cliff faces of the latter, a thick talus body has accumulated below the free face, indicating severe frost weathering above the ice surface at this site.

A clear trimline rises eastward from this scree slope into Coire Mhic Dhughaill where it reaches a maximum altitude of ca. 580 m (NC 360 396). On the southern flank of the corrie, however, this limit becomes obscured by scree and from ca. NC 350 392 the glacier limits cannot be traced southwards due to the absence of constructional or erosional glacial landforms that could be used as indicators of ice marginal positions. However, the limit can be extrapolated southeastwards to east of Leac a'Ghobhainn (NC 347 378) where a large meltwater channel, ca. 10 m deep, 15 m wide and ca. 1 km long is cut into bedrock at an altitude of ca. 300 m. The long profile of this channel is regular, it runs parallel to the axis of Loch More and to the ice flow direction as inferred from the orientation of roches moutonnées in this area and is thus interpreted as a lateral meltwater channel marking the eastern extent of the More Glacier in this part of the trough. This limit corresponds to the upslope termination of ice-moulded bedrock on the southern loch shore around NC 331 358. These findings imply a confluence of the glaciers sourced in Coire Mhic Dhughaill and that occupying the Loch More trough around NC 3370382. Overall, both lateral and vertical glacial limits are well defined in the Loch More and Seilich basins, providing clear morphostratigraphic evidence of a change in landsystem.

Easaidh Glacier

The downvalley limit in Srath Coir an Easaidh is defined by a well-developed terminal moraine, which rises ca. 10 m above the valley floor and is continuous for 100 m and ca. 40 m wide (NC 424 415). A number of indications can be used to argue for this limit representing the Younger Dryas limit: (a) Outside (downvalley) of this moraine, only sparse subdued moraines can be found whereas inside of the limit, clear moraine ridges and mounds, fragmented by meltwater channels, and farther upvalley, west of NC 401 414, "hummocky moraines" occur. South of the river, on Heathery Face Wood, clear lateral moraines descend towards the terminal moraine, showing crestline bifurcations. Outside the limit, elongated and streamlined ridges occur, which parallel the ice flow direction recorded by roches moutonnées (showing an azimuth of ca. 20-25°). These streamlined and subdued moraines are interpreted as small drumlins that formed during the later stages of ice sheet retreat as their direction

differs from the one recorded by moraines in Srath Coir an Easaidh. (b) A clear succession of three different levels of large outwash terrace fragments occur just downvalley of this limit, sloping eastwards towards Gobernuisgach Lodge. Upvalley of the moraine, only one low river terrace occurs ca. 1 m above the present floodplain. (c) The sediment thickness decreases suddenly and markedly outside the limit, bedrock being at the surface in large areas of the valley bottoms whereas bedrock does not crop out in the vicinity of the valley bottom inside the limit. For example, thick till sheets blanket Heathery Face Wood, just inside the limit (NC 415 411). These are locally covered by thin spreads of talus around NC 401 410 farther upvalley where steep rock faces overlook the valley bottom; elsewhere outside the limit, such talus sheets are thicker and occur under less steep faces as well (e.g. NC 424 408).

This limit, marked by the clear terminal moraine, can be traced upvalley for about 1 km using the gradient of lateral moraines, but is then replaced by rock faces to the south and obscured by dense forest plantations in the northern part of the valley. On the northern flanks of the valley, however, the lateral moraines marking the eastern termination of the valley glacier can be extrapolated to join up with the trimline in the northern part of Bealach na Fèithe (see above). Extrapolation on the southern flanks is less straightforward, as no clear clues exist that would allow the lateral margin of the glacier to be constrained. This might be because ice masses tributary to the valley glacier were possibly sourced on the gentle plateau surface of Bruach an Fhraoich (NC 405 405) and in Coirein Easach (NC 403 397). On the gently-northward sloping plateau of Bruach an Fhraoich, a cover of frost-weathered soliflucted debris ("mountaintop detritus") extends from Loch na Mang (NC 391 397) to the plateau edge, with no signs of modification by ice movement. However, this periglacially-modified surface is bounded laterally by the ice-scoured tributary trough of Coirein Easach, and this boundary is sharp and well-defined. This juxtaposition of glacially-sculpted terrain along a topographic trough on the one hand and unaltered terrain on the other is interpreted here as evidence of warm-based, erosive ice in the trough and cold-based, protective ice on the plateau with a subglacial thermal boundary between the two (Sugden, 1968; Sugden and Watts, 1977; Kleman, 1994) and provides evidence of the existence of palaeo-plateau ice (Rea *et al.*, 1998, 1999; McDougall, 2001; Rea and Evans, 2003).

The plateau ice on Bruach an Fhraoich was replaced by warm-based ice in Coire Mhic Fàide (NC 381 393) as evidenced by the presence of roches moutonnées and ice-moulded bedrock that indicate a northerly flow direction. Furthermore, moraines at NC 385 386 indicate northward retreat up the Bealach Lochan a'Bhealaich and suggest that the former ice divide was located here with a transition from cold-based plateau ice to thicker, southward-flowing, warm-based ice just north of the aforementioned lochan. The only indication for a vertical limit in the Bealach is at NC 382 387 on the southeastern slopes of Càrn Dearg where the top of a small rockslope failure grades into a periglacial trimline at ca. 580 m to the west. To the east, this trimline is lost on the steep, scree-covered slopes above the Bealach and Coire Mhic Fàide, but can clearly be recognised in the corrie headwalls

overlooking Coire Loch (NC 369 396) at an altitude of 620 m. This trimline is marked by a transition from blockfields and solifluction lobes at higher altitudes to ice-moulded and plucked bedrock faces which can be defined with a vertical accuracy of ca. 10 m. This limit in turn can be linked up with the trimline on Meall Garbh (NC 369 404) described above.

In the valley bottom of Coire na Phris (NC 382 403) numerous flutes can be found that trend parallel to the ice flow direction indicated by roches moutonnées at the lip of Coire Mhic Fàide. Together these indicate that flow accelerated as a result of the steep drop of over 200 m between the source basins of Coire Loch and Coire Mhic Fàide and the valley bottom of Coire na Phris. This lends further support to the interpretation of the evidence in terms of the juxtaposition of cold- and warm-based ice where plateau ice contributed to ice masses in the valleys. As indicated above, the Easaidh and Ulbhach Glaciers were connected through the narrow Bealach Lochan a'Bhealaich, and, as evidenced by diverging ice flow directions north and south of this pass, the former ice divide is likely to have been located there.

Evidence for a glacial limit in Coire Granda is given by a terminal moraine that crosses the valley at an altitude of ca. 240 m (NC 420 398). This limit extends upvalley on both valley sides as lateral moraines that rise to an altitude of ca. 400 m. Roches moutonnées just west of Beinn Direach (NC 410 384) indicate that ice flow was diverted north of the col with Coire a' Glaise (NC 401 385) into a westerly flow towards Coire an Easaidh and an easterly flow into Coire Granda, therefore linking the evidence in the latter corrie with that on the plateaux described above. Furthermore, the flow directional evidence strongly suggests that the Easaidh Glacier was also connected to the southern transection glacier complex via this col.

3.3.2.3 The southern transection complex

Ulbhach basin

As discussed above, the southern and central transection complexes were connected via the Bealach Lochan a'Bhealaich. A distinct drift limit occurs at an altitude of ca. 590 m along the southern slopes of Carn Dearg (NC 378 386), and this can be traced to the north and joined up with the evidence for the Easaidh Glacier as described above. Southwards, this limit descends to an altitude of ca. 450 m, being defined by a distinct drift limit on the eastern side. A diffuse upper limit of glacially transported boulders that show limited signs of surface weathering can be traced southward along the eastern slope of Meallan Liath Beag from ca. 530 m (NC 370 374) to ca. 440 m (NC 363 357). Meall Gillie Pheadair (NC 361 356, 387 m) shows no signs of frost weathering and displays a rounded top, implying, together with the evidence from the surroundings, that it was overridden when the glaciers were at their maximum. In the southwestern part, particularly in the valley connecting Lochs More and Merkland, the ice surface has to be interpolated entirely between the limits in the Loch More trough

and those in the Ulbhach, Feur Loch (NC 372 312) and lower Bealach nam Meirleach basins where they are much more distinct (see below).

A' Glaise basin

Distinct trimlines occur in the upper parts of Coire a'Glaise east of Carn an Tionail at an altitude of ca. 570 m (NC 401 386). Ice-scoured bedrock and large roches moutonnées below this trimline give way to solifluction lobes and mountaintop detritus on either side of a pass which is at an altitude of ca. 530 m. The ice flow direction as indicated by roches moutonnée lee face orientations changes from southwesterly and southerly south of this pass to easterly and northeasterly north of it. On the east side of Coire a' Glaise, a conspicuous upslope limit of subangular, facetted boulders coincides with the lower limit of solifluction lobes descending downslope from 550 west of Beinn Direach (NC 407 380) to ca. 500 m west of Meall a' Chleirich (NC 407 3650). This boulder limit is inferred to mark the glacier surface in this valley, and it is consistent with the surrounding evidence. This limit is obscured by a large rockslope failure on the south side of Meall a' Chleirich, but interpolation along the northern slope of the northern Bealach nam Meirleach allows it to be joined to the northern upslope limit of the Meirleach Glacier which is defined by lateral moraines (see below).

In the western parts of the Meirleach basin, distinct trimlines culminate at 480 m over a pass on the west side of Ben Hee (NC 401 355). These can be correlated with a distinct arcuate moraine that crosses the valley of Coir' a Chruteir to the east (NC 402 342; 450 m) and interpolated with lateral moraines on the northern side of Ben Hee (see below). In general the trimlines rapidly descend to altitudes of 450 m to the northeast and south, respectively, and southwards they can be linked with further evidence for glacier ice in the Loch Merkland trough.

Meirleach Glacier

In the northern parts of Bealach nam Meirleach, a distinct glacier limit is defined by the distribution of "hummocky moraines" that terminate in an arcuate shape around NC 447 396. This termination defines a boundary between two different landsystems: to the north, the valley floor is occupied by a sequence of outwash terraces in a deeply-incised meltwater channel, and higher slopes are mantled with frost-weathered bedrock; whereas south of this limit, "hummocky moraines" and only a low outwash terrace occur. Appendix 2 only shows the deeply-incised meltwater channel as the terraces cannot be displayed at the print scale. On the northern slopes of the Bealach, lateral moraines can be traced from the terminal moraines to an altitude of ca. 350 m before they fade out into an indistinct and discontinuous upslope termination of the sediment blanket. However, as indicated above, this limit can be joined with the distinct upslope limit of boulders in Coire a' Glaise. On the southern slopes, lateral moraines and the upslope termination of subangular, facetted boulders can be used to constrain the limit. These can be traced to an altitude of ca. 300 m (NC 457 386) before being obscured by thick

peat on Druim nam Bad (NC 460 372). However, ice-moulded bedrock on top of Meall Carr nan Ruadhag (NC 458 369; 364 m) indicates that this was not affected by frost-weathering to the degree that can be observed elsewhere at this altitude outside or above distinct glacial limits, thus providing a minimum estimate of the ice surface at this point. Further information on the maximum elevation of the ice surface is provided by large moraine ridges that trend downslope towards the east in Coire na Saidhe Duibhe and reach a maximum elevation at ca. 280 m (NC 466 368), ca. 700 m east of the overridden Meall Carr nan Ruadhag. Thus, interpolation between these marker points allows the ice surface to be constrained with relative confidence in this location. The limits of the Mudale Glacier to the east will be discussed below. The northern spur of the Ben Hee massif (NC 442 366) is marked by a well-defined ridge that consists of angular boulders with a-axes of up to 5 m. From the spur it extends towards the NNE. This ridge is interpreted as medial moraine that marks the confluence of the former Mudale and Meirleach glaciers in this area.

Farther south, on the southwestern slopes of Ben Hee, around NC 435 365, lateral bouldery moraines are banked against bedrock and can be used to constrain the upper surface of the glacier. These indicate a rise of the glacier surface to ca. 480 m at NC 401 355 where they can be linked with evidence for the glacier surface in Coire a' Glaise and the southern part of Bealach nam Meirleach (see above).

Shin Glacier lobe

The southern limit of the Shin Glacier is defined by large moraine ridges on either side of the loch that attain maximum heights of 10 m, widths of 100 m and lengths of 300 m. These moraines are located at an altitude of ca. 100 m (NC 405 225) and can be linked to distinct lateral moraines that rise gently northwards where they continue intermittently on the eastern flank of Lochs a'Ghriama and Merkland, delineating the former ice surface. Further lateral moraines rise westwards at the northwestern end of Loch Shin where they trend obliquely across the slopes around Corriekinloch (NC 362 263). Their upslope termination is marked by a sudden decrease in boulder frequency and an abrupt termination of the thick sediment cover. Particularly good examples occur on the eastern slopes between Lochs a'Ghriama and Merkland at an altitude of ca. 270 m (NC 401 282). Using these lateral moraines and the evidence from drift limits, the former ice surface can be established with great confidence, even along the axis of Loch Merkland. Likewise, drift limits can be traced around Sron na Garbh Uidh (NC 373 268; 393 m) and Grianan a'Choire (NC 356 274; 468 m) and provide clear evidence for the former glacier surface. This compensates for the problems encountered in adjacent areas where evidence is sparser, e.g. on the western slopes above Creag nan Suibheag (NC 382 295) where the lower limit of frost-weathered bedrock indicates a southward-sloping glacier surface between ca. 400 to 350 m. Similarly, the limit farther north and west, in the Feur Loch basin (NC 371 312) is defined by discontinuous trimlines around 450 m that make interpolation over larger distances necessary.

Around Meallan a'Chuail (NC 345 293) these trimlines rise westwards to a maximum altitude of ca. 650 m, indicating that the latter mountain and the Feur Loch basin formed part of the source area for the Shin Glacier. In addition, these trimlines, and lateral moraines farther north, allow a link between the More Glacier and the Shin Glacier to be established. One recognisable problem for constraining the Shin Glacier is the absence of flow directional indicators due to the valley axes being occupied by lochs. Thus, the consistent slope provided by the lateral moraines is the only solid line of evidence that can be used.

3.3.2.4 The eastern transection complex

Mudale Glacier

Evidence for a glacial limit east of Ben Hee is given by numerous terminal moraine fragments that form a broad arcuate ridge across the River Mudale (NC 545 360). The outermost moraine ridge can be traced upslope for ca. 500 m on the northern valley side where it rises gently towards the west. On the southern side, the limit is more continuous and after ca. 1600 m grades into a discontinuous boundary of sediments banked against bedrock. This limit ascends obliquely towards the south and southwest where it is replaced by a sudden change in sediment thickness at altitudes around 200 to 300 m. Similar limits can be observed in Srath a'Dhuibh (NC 490 339) and Coire na Saidhe Duibhe (NC 485 358). All these limits rise westward and southwards until they are largely obscured by thick peat near the plateau edges on either side of An Glas-Loch (NC 495 312) and Meall na Teanga (NC 480 345). On the plateaux themselves, however, frost-shattered roches moutonnées showing northeasterly ice flow directions can be found that bear no relation to the underlying topography. Similar lee face orientations of roches moutonnées can be found across higher plateaux in this area and those farther north in the area between the junctions of Glen Golly and Strath na Easaidh and in areas nearer the coast. As they are unrelated to radial flow within the valleys, these are interpreted as indications of the flow direction of the last ice sheet across the area. This information can thus be used to delineate three valley glaciers in this area that occupied Coire na Saidhe Duibhe, Srath a'Dhuibh and the valley containing An Glas-Loch. The upvalley limits of these glaciers are well constrained by distinct upslope boulder limits (west sides of Creag Dhubh Mhór, ca. 470 m, NC 455 335) and upslope boundaries of thick sediment accumulations (valley containing An Glas-Loch; ca. 320 m; NC 469 298). Contrasting with these clear terminations, the top of Bad a' Bhacaidh (NC 454 352) is characterised by ice-moulded bedrock, and its eastern side contains fluted moraines indicating ice flow into the northern part of Coire an Saidhe Duibhe. This evidence suggests that the top of Bad a' Bhacaidh, in contrast to Creag Dhubh Mhór (553 m), was overridden.

On the east side of Ben Hee, evidence of former glacial limits is restricted to distinct trimlines on the higher slopes where the lower limit of solifluction lobes and the upper limit of ice-moulded bedrock allow delineation with an accuracy of 10-50 m. In An Gorm Coire (NC 448 341) a large

accumulation of sediment is banked against the northern corrie flank. Its surface is fluted, and ice-moulded bedrock extends to an altitude of ca. 670 m near the headwall. The extent of ice-moulded bedrock farther north up to 650 m (NC 442 350) and up to 600 m farther south in Coire nam Mang (NC 430 330) suggests that these altitudes delineate the former glacier surface well, indicating that the corries on the east side of Ben Hee were the main source areas for these glaciers.

The pass southwest of Coire nam Mang (NC 407 316, ca. 445 m) contains ice-moulded bedrock slabs at its top, partly obscured by peat. Together with lateral moraines that rise westwards in Coire na Claise Móire (NC 425 319, ca. 410 m) this suggests that this col was overridden and that the glacier surface descended southeastwards towards the Fiag basin. Thus, the southern and eastern transection complex appear to have been connected during the maximum glacier extent.

Fiag Glacier

On the eastern side of Meallan Liath Beag, large lateral moraines and a distinct upslope limit to a thick sediment cover allow the delineation of a glacier limit that slopes southeastwards from an altitude of ca. 400 m at the corrie head to ca. 300 m near the western flanks around Loch Fiag (NC 440 290) from which it descends further downvalley. On the opposite valley flank, a distinct boulder limit descends from a similar altitude (NC 460 290) towards the south. This boulder line demonstrates that glaciers in Glen Fiag and the valley containing An Glas-Loch to the east were not confluent, but more importantly it allows the glacier surface slope to be constrained very accurately. The maximum extent of the Fiag Glacier is marked by a prominent moraine arc that crosses the valley at an altitude of 170 m (NC 458 248). On the eastern side, this ridge grades into an indistinct drift limit which can be extrapolated to the lateral moraines and boulder limit east of Loch Fiag. On the western side, lateral and "hummocky moraines" extend into Strath Duchally, ascending from ca. 240 m at the southern end of the valley to ca. 300 m at its western end. The col that delimits this relatively narrow and short valley at the western end (NC 412 287) contains numerous subangular, facetted boulders and ice-moulded bedrock slabs, also indicative of overriding and providing evidence for a further confluence of the Shin and Fiag Glaciers.

3.3.2.5 Arkle corrie complex

Compared to the evidence for larger glacier lobes within transection complexes presented above, the evidence for a former distinct period of glaciation on Arkle (NC 310 454) is spatially more restricted. Evidence for a former larger glacier lobe is provided by up to 200 m-long, bouldery moraine ridges at the eastern side of the two corries. These moraines are separated by Loch an Easain Uaine (NC 323 464), but their trend, and that of moraines inside the terminal moraine, suggests that they mark the same, arcuate limit. Further evidence for a confluence of two corrie glaciers is provided by a small medial moraine composed of angular quartzite boulders at the spur protruding between the two

corries (NC 318 459). Flutes in the corries give a good indication of the former ice flow direction and further help constraining the glacier dimensions. The terminal moraines grade into lateral moraines and then the upslope termination of thick scree slopes that are entirely composed of quartzite debris. These limits can be followed through the whole corrie perimeter.

3.3.3 Ice cap reconstruction

The evidence for distinct glacier limits in the individual basins as described above can be correlated across the study area using the methods described in Chapter 2.1.3.2 to delimit a large coeval mountain ice field (Fig. 3.4). To facilitate reading, the ice mass will be referred to as "ice cap" rather than mountain ice field below.

From the geomorphological evidence it becomes apparent that the northern lobes of the More and Seilich glaciers were coeval and most likely responsible for the input of glaciolacustrine sediment into the Loch Stack basin as indicated by the data presented by Pennington (1977) and above. Thus, a Younger Dryas age (12.7-11.5 cal ka BP) of this ice cap and the independent corrie glacier complex on Arkle can be established.

Ice surface contours shown in Fig. 3.4 have been reconstructed at 50 m intervals applying methods outlined earlier (Chapter 2.1.3.3). The reconstructed ice cap is quite substantial, measuring ca. 31.4 km from north to south (Dionard to Shin Glacier termini) and a minimum of 22 km E-W (Mudale terminus to western map boundary), occupying a total area of 211 km^2 (Table 3.2). The ice cap was sourced in the central mountains where it reached its maximum altitude in the corries east of Foinaven (650-770 m), northeast of Meall Horn (640 m) and around Ben Hee, east of Meallan a' Chuail and northeast of Sabhal Beag (ca. 670-680 m). The ice surface was somewhat lower above the passes between Carn an Tionail and Carn Dearg, Sabhal Beag and Sabhal Mor (620 m) and Carn an Tionail and Beinn Direach (570 m). Ice divides have been reconstructed from ice-flow directional indicators and the shape of contours on the ice surface. According to this evidence, the reconstructed ice cap had a complicated flow pattern in the central parts of the mountains, with local ice divides splitting feeder glaciers into two lobes downglacier of passes and valley junctions, for example.

While the ice surface contours imply relatively gentle and regular surfaces over most of the ice cap, relatively steep and irregular portions are encountered in the corries around Foinaven (Fig. 3.4). These correspond to steeper sections in the present valley floor topography and might thus be interpreted as an indication of localised ice falls that fed the glaciers in Strath Dionard. The long, very gently sloping glacier segment in the northern part of Strath Dionard deserves attention as a similar situation cannot be found elsewhere in the study area. Two factors are likely to contribute to the relatively gentle ice surface slope in this location. Firstly, the valley floor is similarly gentle and does not show any large undulations; bedrock obstacles are absent from this part of the valley. Secondly, the gentle part coincides with the confluence of the main glacier in the Strath with those sourced in the

corries east of Foinaven. This confluence is characterised by steep glacier sections, indicating an ice fall at this point. Since there is no geomorphological evidence to suggest that the glacier surface bulged up as a result of the confluence, the input of additional glacier ice cannot have been sufficient to cause substantial thickening of the main glacier, however. Hence it would appear that the Foinaven corrie glaciers and the main Dionard glacier would have just been confluent at this point, making the glacier atypically long compared to other glaciers (cf. Benn and Evans, 1998: Fig. 9.57).

Fig. 3.4 Extent of the reconstructed transection glacier complex and isolated corrie glacier complex of Younger Dryas age in the study area. The southwestern and western limits of this complex are outside the present study area but will be joined up once this information is available from BGS mapping.

3.4 Palaeoclimatic significance

3.4.1 Equilibrium-line altitudes

Due to the fact that the southwestern and southern limits of the ice cap were not encountered in the study area, the reconstruction of palaeoclimatic variables is unfortunately restricted at present. However, the missing area has been mapped by T. Bradwell, and this information will be linked to the evidence presented here once these maps have been finalised and produced at the British Geological Survey (T. Bradwell, pers. comm., 2005). Thus, the variables calculated from the currently available evidence should be regarded as preliminary.

Applying the approach outlined in Chapter 2.1.3.3 to the individual glacier basins delineated above yields a range of ELAs across the ice cap (Table 3.2). Due to the nature of the individual methods these results differ slightly for each glacier. Earlier work, recently reviewed by Osmaston (2005), has demonstrated that the Area-Altitude Balance Ratio (AABR) is most reliable in approximating the true climatic ELA, and a balance ratio of 2.0 has been suggested to closely represent conditions at mid-latitude glaciers in general (Benn and Gemmell, 1997). This suggestion is followed here, and ELA3 (Table 3.2) will be used for all further calculations and comparisons.

The ELA is calculated to have been 324 m for the ice cap as a whole and 393 m for the independent corrie glacier complex on Arkle. The ELAs in individual glacier basins are relatively consistent across the ice cap, ranging from 268 to 429 m, which is comparable to the ranges of ELAs calculated for ice caps and transection glacier complexes elsewhere in Scotland using similar approaches (e.g. Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005). However, the low values of the Mudale Glacier and the comparatively high values of the Easaidh and Meirleach Glaciers merit further discussion. The distribution of glaciers in the study area is somewhat asymmetric with the longest glaciers in the east and the north (Fig. 3.4). This is characteristic of many Scottish Younger Dryas glaciers and has been interpreted as evidence of dominantly westerly and southwesterly snow-bearing winds that have led to an accumulation of glaciers in the lee of mountains or on plateaux (Sissons, 1979a, 1980; Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005).

A dominant wind-direction from the southwestern quadrant has recently been confirmed by independent mapping of the orientation of snow- and windpolished surfaces across Scotland (Christiansen, 2004). The distribution and orientation of glaciers in the study area thus probably reflects the overall palaeo-windfield during the Younger Dryas in the far NW Highlands, which would appear to have also been dominated by westerly and southwesterly winds. Therefore, the low ELA of the Mudale Glacier in the east might be attributed to snow being preferentially accumulated in the lee of Ben Hee, i.e. its eastern side. The three corries along this eastern flank would have acted as major snow traps which might help explaining the existence of a very large glacier in this location. The southern part of the Mudale Glacier appears to have been fed by cold-based ice from surrounding

plateaux since this feeder glacier is below the general ELA and could thus not exist without additional inputs from surrounding slopes and plateaux. However, due to thick peat cover on these plateaux, the exact extent of cold-based plateau ice cannot be established.

Table 3.2 Equilibrium-line altitudes calculated for individual glaciers in the study area. The different methods used are explained in Chapter 2.1.3.3.

	Size (km^2)	ELA1 BR=1.67 (m)	ELA2 BR=1.8 (m)	**ELA3 BR=2.0 (m)**	ELA4 AAR=0.5 (m)	ELA5 AAR=0.6 (m)	ELA6 (Sissons, 1974) (m)
Arkle corrie complex	*2*	*401*	*397*	***393***	*394*	*359*	*424*
Northern corrie	1.1	422	418	**412**	433	369	450
Southern corrie	0.9	376	373	**370**	380	337	394
NW Highlands ice cap	*211*	*332*	*328*	***324***	*335*	*296*	*331*
Dionard Glacier	35.3	334	331	**325**	334	290	363
More and Seilich Glaciers	44.8	334	331	**326**	386	341	358
Easaidh Glacier	13.8	437	434	**429**	473	453	459
Meirleach Glacier	15.1	381	379	**375**	397	369	400
Fiag and Shin Glaciers	65.5	331	328	**324**	360	325	352
Mudale Glacier	36.5	276	273	**268**	294	260	299

Contrasting to this comparatively low ELA, that of the Meirleach and Easaidh Glaciers is higher than the overall value for the whole ice cap. In the case of the Easaidh Glacier, a larger plateau area at altitudes between 600 and 450 m contributes to the accumulation area, making up ca. 15% of the overall glacier surface (Fig. 3.4). Although this plateau was covered by ice, it was thin and presumably cold-based as evident from the juxtaposition of unmodified mountaintop detritus and ice-moulded and plucked bedrock in adjacent topographic troughs where ice thickness would have been greater, promoting basal sliding (e.g. Sugden, 1968; Hall and Glasser, 2003). The presence of plateau ice caps tends to raise the ELA compared to situations where valley glaciers accumulated in cirques (e.g. Rea *et al.*, 1998, 1999; McDougall, 2001; Rea and Evans, 2003), and this appears to be the most likely explanation for the anomalously high ELA of the Easaidh Glacier. The Meirleach Glacier, on the other hand, would appear to have been relatively unsheltered from winds that could have blown snow off its surface in the lower parts; apart from the immediate, relatively sheltered source area east of Carn an Tionail only two smaller tributary corries on the north-western side of Ben Hee would have contributed to its overall mass balance (Fig. 3.4). The ELA of the small independent glacier complex on Arkle is somewhat higher than average and might have been caused by the restricted nature of lee-side basins in which snow could have accumulated. When calculated independently, the ELA of the southern corrie glacier is 42 m lower than that of the northern one which is probably due to the different altitude of the corrie rim.

3.4.2 Palaeo-precipitation

As discussed in Chapter 2.1.3.3 the ELA marks the place where accumulation and ablation are exactly equal. While the former is represented by annual precipitation, the latter can be equated by mean summer temperature (e.g. Ohmura *et al.*, 1992; Benn and Evans, 1998; Kerschner *et al.*, 2000). Because accumulation is correlated with annual precipitation and ablation is correlated with summer air temperature, a positive correlation between summer air temperature and annual accumulation exists at the ELA (Ohmura *et al.*, 1992). Numerous empirical studies have found relationships between these variables for glaciers in different geographical settings (e.g. Sutherland, 1984b; Leonard, 1989; Dahl *et al.*, 1997). One difficulty in applying these local or regional data sets is that the relationships between energy balance and air temperature varies with geographic location (Kaser and Osmaston, 2001; Benn *et al.*, 2005). Hence a global data set (Ohmura *et al.*, 1992), which removes such difficulties, is more reliable for the reconstruction of palaeoclimatic data from a location for which no modern glacier data are available (Benn and Ballantyne, 2005). The data set presented by Ohmura *et al.* (1992) defines the relationship between summer temperature and precipitation at the ELA as follows:

$$P_a = 645 + 296T_3 + 9T_3^2 \qquad (3.1)$$

where P_a is the annual precipitation in mm a^{-1} (water equivalent) and T_3 is the mean summer temperature (°C) in the free atmosphere at the ELA of the months June, July and August for the northern hemisphere.

Quantitative estimates of air temperatures during the Younger Dryas can be obtained from subfossil assemblages of coleoptera (e.g. Coope *et al.*, 1998), pollen (e.g. Isarin and Bohncke, 1999), chironomids (Brooks and Birks, 2001) and the distribution of periglacial phenomena (Ballantyne and Harris, 1994). A compilation of data by Isarin *et al.* (1998) and Isarin and Renssen (1999) indicates mean sea-level temperatures of ca. 10°C for the warmest month, <-25°C for the coldest month, and mean annual temperatures of <-8°C for the coldest part of the Younger Dryas in Scotland. Slightly lower mean July temperature estimates during the Younger Dryas have recently been obtained by Brooks and Birks (2000, 2001) using chironomid assemblages. The only published site for Scotland is Whitrig Bog, ca. 350 km to the south of the present study area. Mean July air temperatures at this site (125 m OD) are 7.5°C for the coldest part of the Younger Dryas. Unpublished mean July temperatures from Abernethy Forest (220 m OD), ca. 150 km to the south of the present study area, indicate July temperatures of 7.3°C for the same time interval (Benn and Ballantyne, 2005). Assuming typical wet adiabatic lapse rates of 0.006 to 0.007°C m^{-1} (Häckel, 1999), these values would correspond to 8.2 to 8.3°C and 8.6 to 8.8°C at sea-level. Following Benn and Ballantyne (2005), and in the absence of a closer site with mean July temperature estimates, a mean Younger Dryas sea-level temperature of

8.5±0.3°C is used as a basis for all subsequent calculations. As the data set by Ohmura *et al.* (1992) requires mean summer temperature (T_3), this has to be calculated from the available mean July temperatures (T_J). Benn and Ballantyne (2005) state that the two are related as follows:

$$T_3 = 0.97\ T_J \tag{3.2}$$

Using a sea-level temperature of 8.5±0.3°C and this relationship, and assuming typical lapse rates, the resulting mean summer temperature at the ELA of the main ice cap (324 m) is 6.2±0.5°C. Applying these values to equation (3.1) by Ohmura *et al.* (1992) yields an annual precipitation of 2828±404 mm a^{-1} at the ELA. The error range reported here includes cumulative uncertainties introduced by temperature, varying lapse rates and a standard error of ±200 mm a^{-1} that reflects regional variations in the relationship between air temperature and ice melt (Ohmura *et al.*, 1992). Mean modern annual precipitation from Loch Merkland (NC 401 297, 118 m) is 1990.8 mm a^{-1} for a 15-year period between 1969-1989 for which the rainfall record was complete (British Atmospheric Data Centre). Since precipitation increases non-linearly with altitude (e.g. Ballantyne, 1983; Ward and Robinson, 1990), these values have to be adjusted to those at the Younger Dryas ELA to allow a direct comparison. In order to achieve this, the relationship

$$P_{Z1} = P_{Z2}/(1+P^*)^{0.01\,(Z2-Z1)} \tag{3.3}$$

will be used (Ballantyne, 2002a). P_{Z1} and P_{Z2} represent the precipitation at altitudes Z1 and Z2 and P* is the proportional increase in precipitation per 100 m increase in elevation. Using the example of Ben Nevis, Ballantyne (2002a) has shown that P* = 0.0578, and this value is used here. This yields modern precipitation values of 1863 mm a^{-1} at sea-level and 2235.1 mm a^{-1}at the ELA (324 m).

3.4.3 Comparison and discussion

In order to compare Younger Dryas precipitation values between the NW Highlands and Mull ice caps, the same ELA values (AABR = 2.0) and mean July temperatures have been used. Inserting these values into equation (3.3) yields mean annual Younger Dryas precipitation at sea-level of 2606±145 mm a^{-1} for the Mull ice cap and 2358±337 mm a^{-1} for the NW Highlands. The values presented here for the NW Highlands are thus only slightly lower than those obtained for the Isle of Mull farther south, the only other site along the Scottish west coast that allows a direct comparison. Benn and Ballantyne (2005) recently found that, using the same approach as that above for the ice cap on the Isle of Mull (Ballantyne, 2002a), Younger Dryas precipitation was ca. 25% higher compared to the present day. The data presented above indicate that precipitation was ca. 26% higher than at present in the NW Highland. Together, these data suggest that Younger Dryas precipitation totals

might have been higher than at present along the Scottish west coast with only a slight north-south gradient. This contrasts with findings from the Central Grampian Highlands (Benn and Ballantyne, 2005) and Cairngorm Mountains where precipitation was equal and less than at present, respectively (D.I. Benn, pers. comm., 2005). The conclusion that snowfall was high in western Scotland during the Younger Dryas is consistent with the modelling results of Isarin *et al.* (1998) and Renssen *et al.* (2001), which indicate strong zonal circulation with increased frequency of depressions, especially in winter. However, since the values calculated here only reflect about two thirds of the NW Highlands ice cap (T. Bradwell, pers. comm., 2005), they have to be regarded as preliminary, and comparisons with other work are tentative at this stage.

The ELA of the NW Highland ice cap is also similar to that on Mull, being ca. about 80 m higher than at the latter site. This appears somewhat counterintuitive since ELAs would be expected to lower with increasing latitude due to a decrease in incoming solar radiation (Benn and Evans, 1998; Nesje and Dahl, 2000). A possible cause for a higher ELA in the present study area might be the fact that the main accumulation area is more removed from the coastline than the higher ground on Mull. Scavenging of moist, southwesterly air masses by the coastal mountains (e.g. Quinag, Ben More Assynt; Fig. 2.2) might have played a role in raising the ELA somewhat. However, in general the values are in good agreement with independent findings from the Scottish west coast.

Sissons (1977) reconstructed glaciers with a total size of 36 km^2 in the present study area. Comparison with the present reconstruction shows that he underestimated the ice extent during the Younger Dryas by almost six times. Apart from the corrie glacier complex on Arkle and the southwestern segment of the Dionard Glacier, the shape of the glaciers reconstructed by Sissons (1977) does not largely correspond to that depicted in Fig. 3.4.

Likewise, and in contrast to the previous work, the ELA values calculated here are much lower and spatially more consistent than the mean value of 520 m calculated by Sissons (1977, 1980) for all glaciers in the NW Highlands. Contrasting to previous suggestions that drier air masses and differences in storm tracks were responsible for such anomalously high ELAs in the far NW Highlands, the consistency of the present work and good agreement with other independent and more recent studies implies that the previous reconstructions were in error.

To conclude, this chapter has presented new evidence for the extent of a transection glacier complex and an independent corrie glacier complex in the study area and has provided a framework within which the detailed geomorphological and sedimentological work on "hummocky moraine" will be presented in the next chapter.

CHAPTER 4 MORAINE MORPHOLOGY, SEDIMENTOLOGY AND SIGNIFICANCE

4.1 Introduction

As shown in the previous chapter, ice-marginal moraines define the extent of glacier ice at a given time and thus form an important part in glacier reconstructions (e.g. Sissons, 1967, 1974, 1976, 1979a; Gray, 1982; Ballantyne, 1989, 2002a; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b; McDougall, 2001; Benn and Ballantyne, 2005; Golledge and Hubbard, 2005). If mapped in detail and if an ice-marginal origin of all the moraines can be established, then they can be used to establish the pattern and chronology of ice retreat (cf. Benn, 1990, 1992a; Bennett, 1990; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003, 2004a; Golledge and Hubbard, 2005). Thus, in order to reconstruct the distribution of glaciers and their dynamics, identifying the exact location and planform patterns of moraines is important. It is equally important that the mode or modes of moraine formation can be demonstrated to avoid misinterpretations.

In the context of Scottish moraines, a number of detailed studies have been carried out that provide good examples of the use of concepts of sediment-landform assemblages to reconstruct the dynamics of palaeo-glaciers (e.g. Benn, 1990, 1992a; Benn and Evans, 1993, 1996, 1998; Benn *et al.*, 2004; Golledge and Hubbard, 2005). Their contributions to the understanding of past glacier dynamics contrast markedly with the majority of purely geomorphological studies that are hampered by limited exposure conditions (e.g. Bennett and Boulton, 1993a, b; Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Lukas, 2003).

As discussed in Chapter 1.1.2, two conflicting and mutually-exclusive models for the formation of "hummocky moraine" have been proposed: (1) the incremental retreat model in which glacier stillstands and readvances produce distinct recessional moraine arcs (e.g. Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Benn and Evans, 1993; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003) and (2) the englacial thrusting model, which proposes that moraines with clear proximal rectilinear slopes represent englacial material melted out during glacier downwasting. In model (1) one moraine arc corresponds to a former ice-marginal position (Fig. 4.1a-c) whereas in model (2) several moraines could be formed in a single stage (Fig. 4.1d-e).

To resolve the problems arising from these conflicting interpretations, this chapter aims to describe the morphology and distribution of moraines in the far NW Highlands. This will serve as a basis for detailed sedimentological case studies from "hummocky moraines" with the particular aim to test which of the two models, if any, are applicable to "hummocky moraine" in the study area. These data are then used to arrive at a well-informed interpretation of moraines in the study area. Further implications for Younger Dryas glacier dynamics will be discussed based on this interpretation.

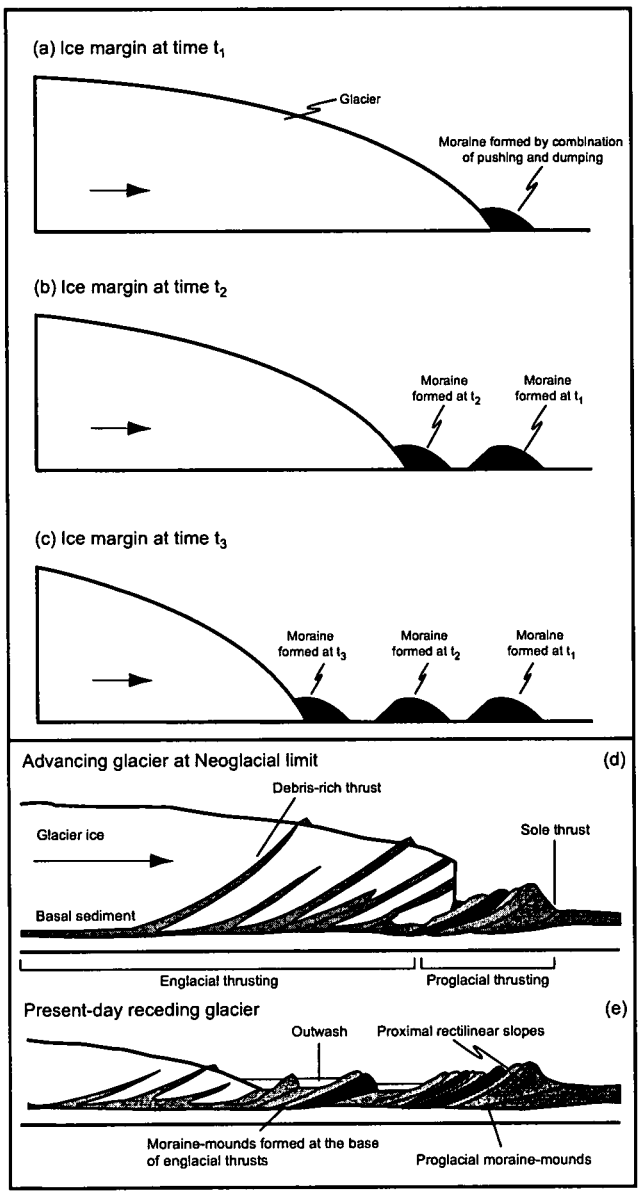


Fig. 4.1 (a) to (c) Conceptual sketch of moraine formation according to the incremental 'active' retreat model without incorporation of dead ice bodies or significant stagnation (for explanation see text). (d) Sketch of englacial and proglacial thrusting as deduced for polythermal glaciers on Svalbard. Stresses are propagated into the foreland along a basal decollèment surface or sole thrust. (e) Lowering of englacial debris after ice retreat to form a sequence of stacked moraines with characteristic proximal rectilinear slopes that dip upglacier. Note the inferred lack of alteration in the mounds due to dead ice meltout ((d) and (e) redrawn from Bennett *et al.*, 1998).

4.2 Moraine types and distribution

In the study area, moraines can be grouped into three categories. (1) those that have previously been described as "hummocky moraine"; (2) lateral moraines that occur higher up on the hillsides and (3) flutes or fluted moraines. Characteristic features and the distribution of each type will be described below.

4.2.1 "Hummocky moraine"

The majority of moraines found in the study area consist of mounds and short ridge fragments, giving areas of moraine a hummocky appearance from the ground (cf. Fig. 4.2), justifying the use of the traditional term "hummocky moraine" for this landform assemblage (cf. Sissons, 1977).

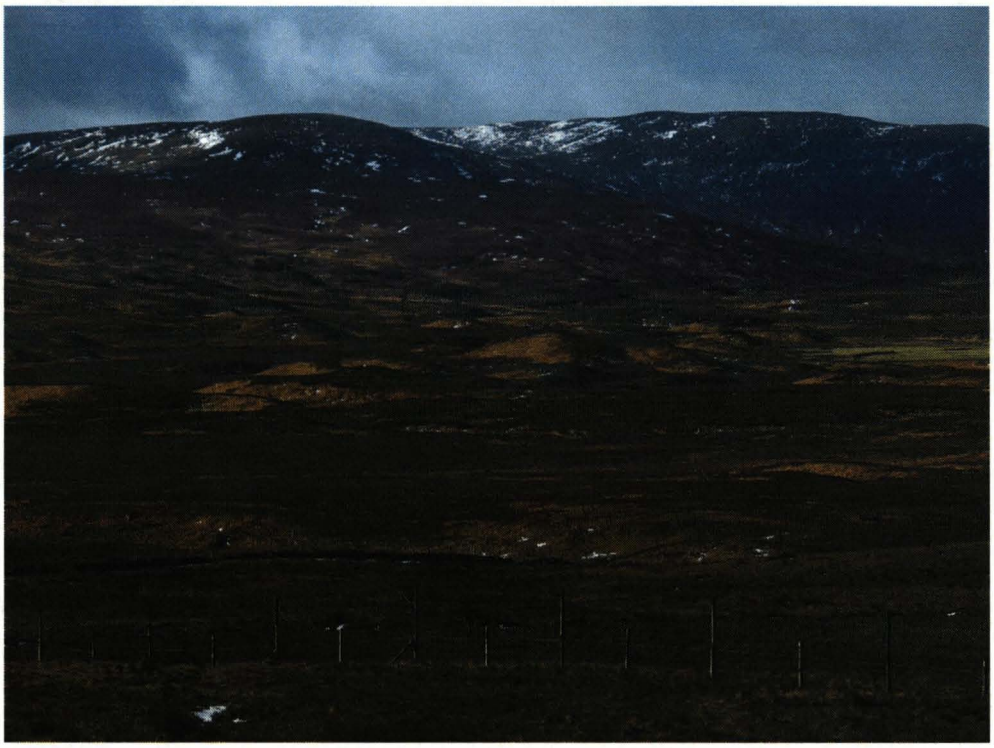


Fig. 4.2 Overview of typical "hummocky moraines" around Mudale Farm (NC 519 354) as viewed from the northeast. Note the seemingly unordered appearance of moraine ridges and mounds as noted by Sissons (e.g. 1967, 1977) elsewhere in Scotland.

Detailed geomorphological mapping, presented in Appendix 2 for the whole study area, shows that this type of moraines is widespread throughout the glens of the far NW Highlands. They usually consist of moraine ridges and mounds that reach heights of between 5 to 15 m, lengths of 5 to 200 m and widths of up to 50 m. In most cases, ridges and mounds are aligned forming chains that trend obliquely downvalley across the slopes towards the valley axis, and their occurrence is usually restricted to the lowermost ca. 50-100 m above the valley floor, a distribution that also appears to be typical of these moraines elsewhere in Scotland (Read, 1931; Charlesworth, 1955; Sissons, 1967, 1977; Benn, 1990, 1992a; Bennett and Boulton, 1993a, b; Lukas, 2003; Golledge and Hubbard, 2005). These arcs in turn have close spacings of only a few metres to tens of metres. Crestline bifurcations, i.e. where one moraine splits into two, occur throughout the study area (Fig. 4.3).

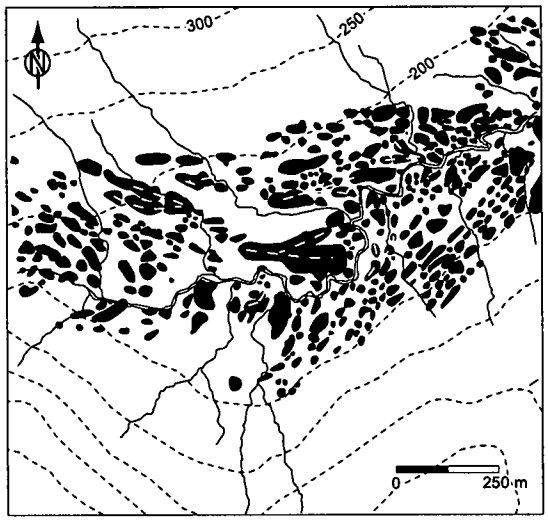


Fig. 4.3 Close-up of an area of "hummocky moraines" SW of Mudale (NC 479 340) showing crestline bifurcations (white stippled lines). For key see Appendix 2.

A survey of the form and symmetry patterns of moraines shows that the majority of moraine ridges and mounds displays a clear asymmetry with a clear steeper proximal and gentler distal slope. Due to blanket peat cover, however, the exact planform of individual moraines (e.g. fan or ridge-shape) cannot be reconstructed with confidence in every case. Because of the importance attributed to the angles of proximal rectilinear slopes to deduce the presence of former englacial thrust planes (Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001, Graham, 2002), and in order to enable direct comparison with the results of the aforementioned model, only the proximal slopes were surveyed using a compass clinometer. Of 118 moraines surveyed, 86.4% contain clear proximal rectilinear slopes. The distribution of slope angles (Fig. 4.4) is very similar to those reported from Svalbard and other isolated sites throughout upland Britain with 44.2% of slopes in the classes 28-33° (cf. Bennett *et al.*, 1998; Graham and Midgley, 2000a; Graham, 2002).

Below, three representative examples of characteristic assemblages of "hummocky moraines" will be given. The reader is referred to Appendix 2 for a full areal representation of the extent, location and planform patterns found in "hummocky moraines" in the study area.

4.2.1.1 Coire na Phris

A large number of boulder-strewn moraine ridges and mounds occur at the junction of the Coire na Phris with the Strath Coir an Easaidh (NC 389 412). In this location, the moraines reach heights of up to 15 m, widths of up to 50 m and lengths of up to 250 m. When viewed under low illumination they appear hummocky, yet individual moraine fragments are clearly aligned in chains (Fig. 4.5). Ridges and mounds are frequently asymmetrical in cross-section with steeper sides pointing at ca. 25 to 35° upvalley and gentler sides dipping at ca. 15 to 20° downvalley. These moraine ridges are often closely associated with meltwater channels that run parallel to the crestlines of the moraines or cross several

arcs, arguably dissecting formerly continuous moraine ridges. The spacing of individual chains of moraines is very dense, between 5 and 20 m. The size of individual moraines varies randomly with a very high (15 m high) chain of mounds and ridges occurring around NC 390 411. In general, moraines are most frequent along the lower 50-100 m of the valley flanks while they are only sparsely encountered in the immediate valley bottom where outwash dominates.

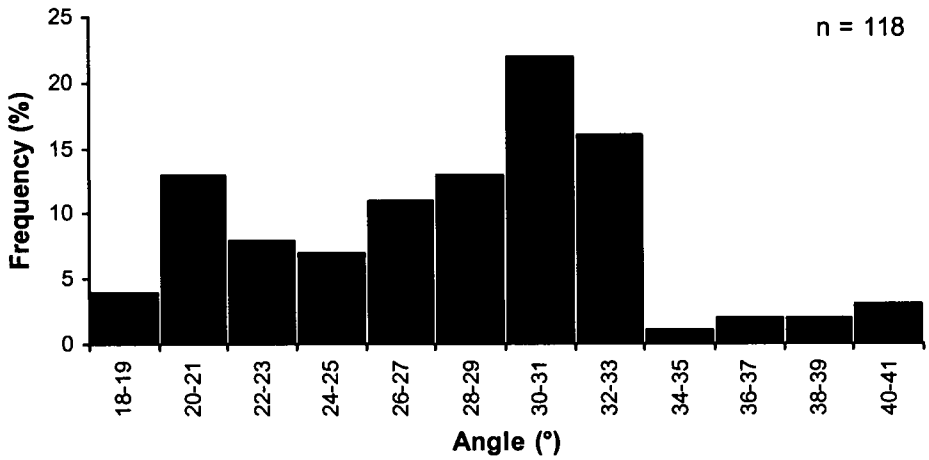


Fig. 4.4 Distribution of slope angles of proximal rectilinear slopes of 118 moraines in the NW Highlands. Only 16 of the 118 moraines (=14.6%) measured displayed an irregular, undulating slope. Note the peak of slope angles around 30°. The classes follow the approach by Bennett *et al.* (1998) and Graham and Midgley (2000a).

4.2.1.2 Mudale Farm

In the lowland area just west of Mudale Farm (NC 517 352), pronounced moraine ridges and mounds with a maximum height of 11 m, width of 20 m and lengths of up to 70 m occur. Connecting the crestlines of these moraines yields arcuate lines. While the moraines in the lowland area around Mudale Farm are relatively large and continuous, those in the more confined valleys to the west are less continuous, contain a higher proportion of mounds and are smaller, reaching a maximum height of 5 m (Fig. 4.6). In addition, although the moraines are more continuous in the lowland area around Mudale, they are also dissected by larger tracts of outwash. Additional evidence for an abundance of meltwater comes from numerous channels that run parallel to the outermost chains of moraines (Fig. 4.6).

Fig. 4.5 Overview of "hummocky moraine" in Coire na Phris seen from the north (ca. NC 392 419). Note how lateral moraines in the centre grade into "hummocky moraine" in a downslope direction.

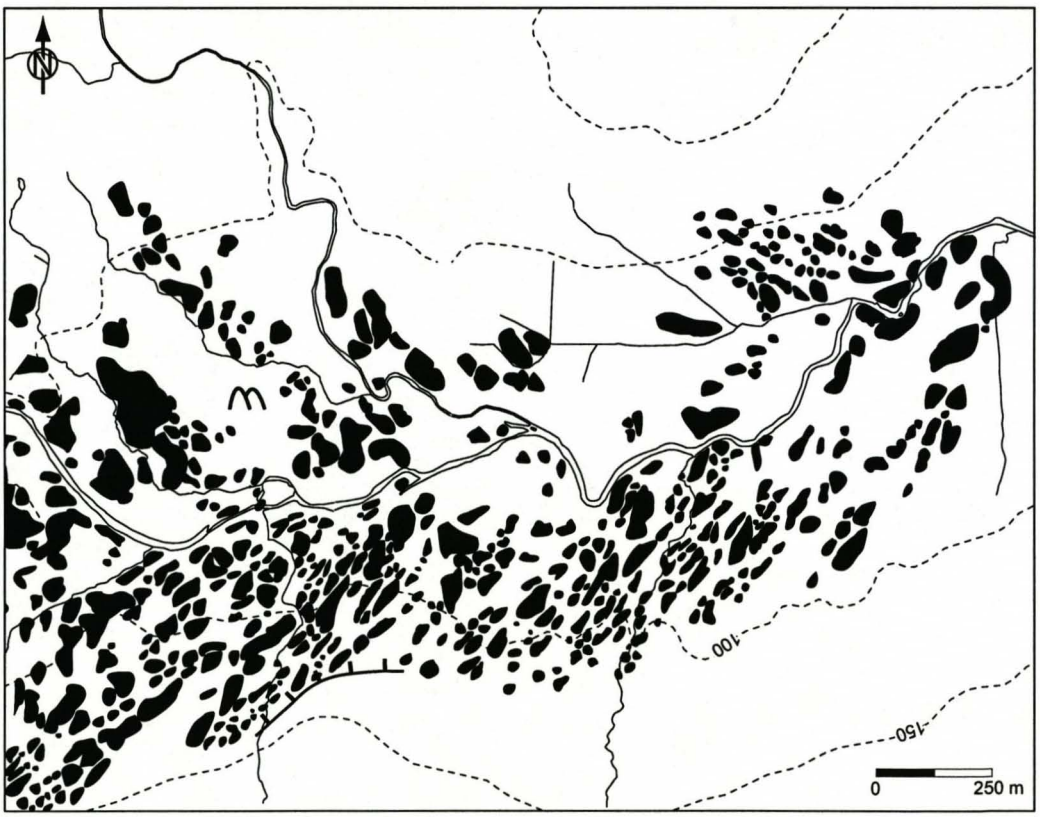


Fig. 4.6 Geomorphological overview map of "hummocky moraine" near Mudale Farm (cf. Fig. 4.2). For key see Appendix 2.

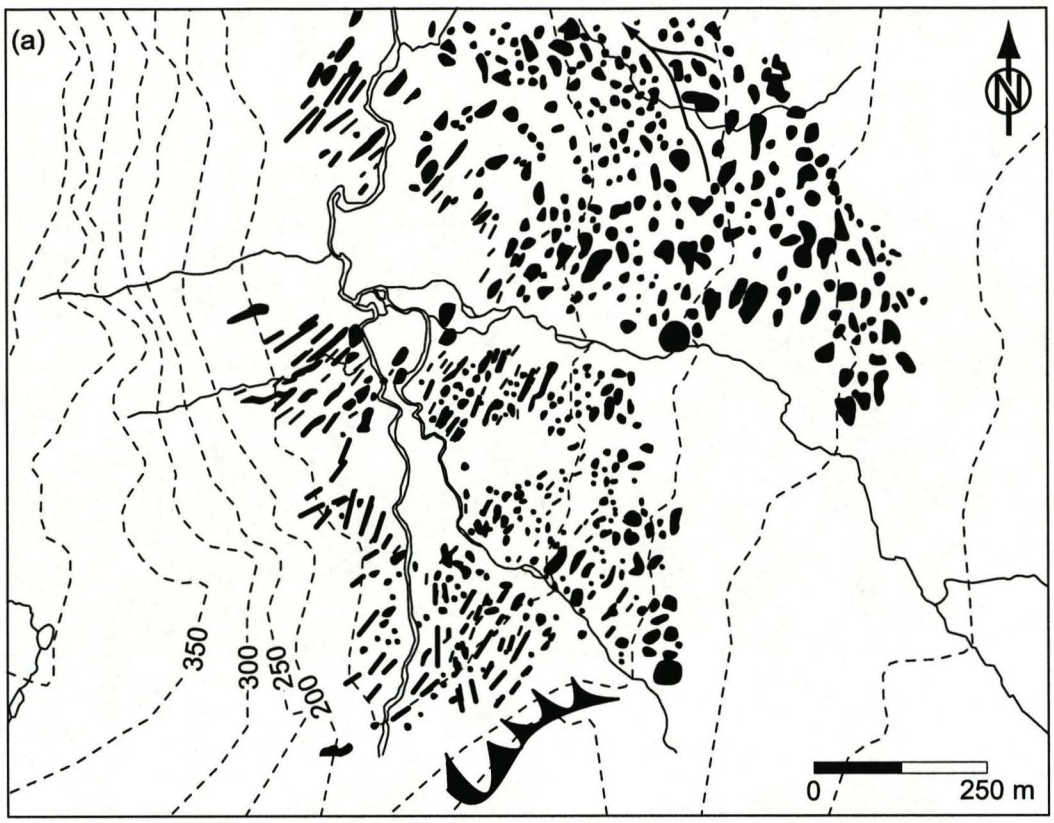


Fig. 4.7 (a) Geomorphological map showing the distribution and planform patterns of "hummocky moraines" at the southern end of Strath Dionard. For key see Appendix 2. (b) Overview photograph of "hummocky moraines" and cross-cutting flutes at the southern end of Strath Dionard viewed from the south.

4.2.1.3 Strath Dionard

One of the best examples of "hummocky moraine" occurs at the southern end of Strath Dionard where the moraines take the form of sharp-crested ridges and pointed mounds, are densely spaced and intimately associated with meltwater channels (Fig. 4.7). When the crestlines of these are combined, a large number of arcuate lines can be reconstructed. Several crestline bifurcations are evident within this field of moraines. The picture given by these moraines is complicated by the fact that they cross-cut with flutes (Fig. 4.7b, Chapter 4.2.3.2).

4.2.2 Lateral moraines

A type of moraines distinctly different from the fragmented moraines confined to the valley bottoms are more continuous ridges, which trend obliquely across the slopes with a low gradient and are typically found at higher elevations above the valley floors, perched on the hillsides. These can in some locations be linked to chains of moraines in the valley bottoms, indicating that they represent the lateral continuation of some "hummocky moraines". They are thus interpreted as lateral moraines that, by analogy with modern glaciers, mark the location of the ice margin. The close association with "hummocky moraines" confirms the point made by Boulton (1992) that the approach of using the upper limit of "hummocky moraine" to reconstruct the ice surface where lateral moraines might be absent (e.g. Lowe and Walker, 1997) is not valid. The value of lateral moraines in glacier reconstructions is two-fold: Firstly, their gradient can be used to reconstruct the slope of the ice surface where they occur, enabling a tight control on the surface configuration to be gained. This represents a significant advantage over other areas where only "hummocky moraines" could be utilised as extrapolation is minimised in such cases (e.g. Bennett and Boulton, 1993a, b; Benn and Ballantyne, 2005). Secondly, the maximum elevation of lateral moraines can be used in ELA reconstructions to constrain the extent of the former ablation area (Meierding, 1982; Nesje, 1992; Torsnes *et al.*, 1993; Benn and Evans, 1998; Nesje and Dahl, 2000). Thus, lateral moraines can be used as an additional control and as a tool for testing whether the reconstructed ELAs are realistic.

As they occur along the sides of most glens in the study area, three representative examples are described below.

4.2.2.1 Loch More Basin

Along the valley flanks surrounding Loch More (NC 330 373), ridges and, less commonly, mounds are aligned in chains that trend obliquely across the slopes. In some cases, ridges grade laterally into terraces or benches. Different fragments are usually cut obliquely by gullies that trend directly downslope whereas the upslope side of some of the lower ridges is marked by meltwater channels that can be up to 2 m deep, 3 m wide and 200 m long. Dense forest obscures these moraines in large areas of the basin, allowing only a partial representation of the true number of lateral moraine fragments to

be mapped (Fig. 4.8). Combining the crestlines, and, in the case of terraces or benches their valley-side break of slope, enables the reconstruction of straight lines with consistently low gradients in the order of 5 m per 100 m.

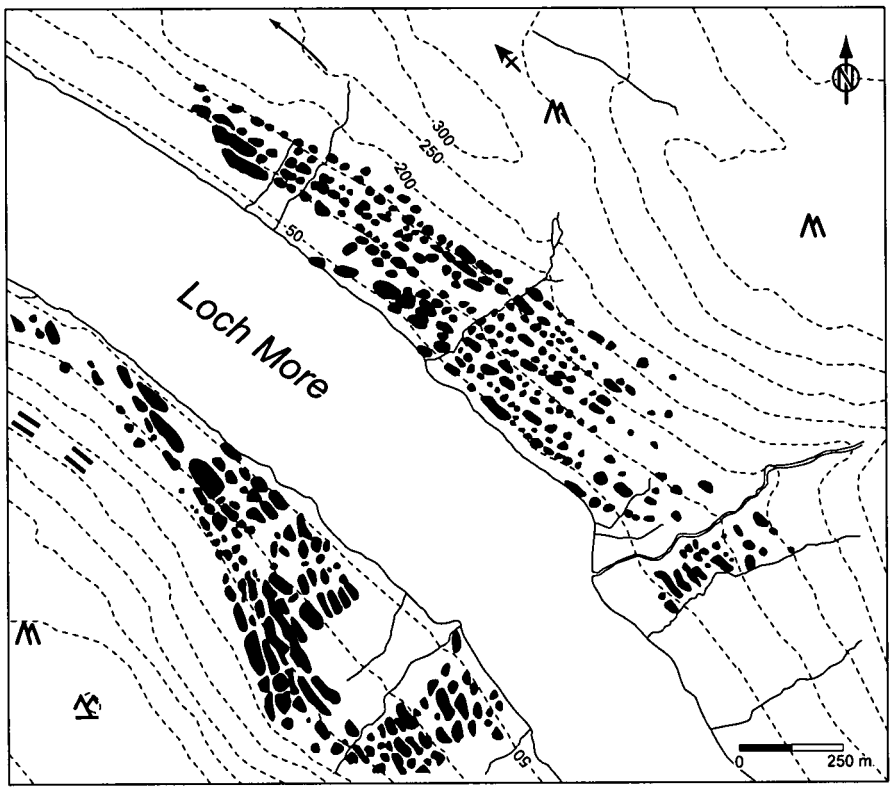


Fig. 4.8 Geomorphological map of lateral moraines along Loch More. For key see Appendix 2.

4.2.2.2 Northern end of Loch Shin

Along the eastern flank of Lochs Merkland, a'Ghriama and the northern part of Loch Shin, superb examples of dissected lateral moraines up to altitudes of 310 m (NC 382 268), very similar in character to the ones described above, can be observed. In this case, their upslope termination is marked by a sudden transition from an irregular surface with abundant subangular boulders of Moine lithologies with a-axes of up to 3 m to a subdued, gently undulating surface with broad solifluction lobes and only isolated, weathered and partly disintegrated, subrounded boulders. Again, this area of lateral moraines consists of several bench and ridge fragments cut by gullies (perpendicular) and meltwater channels (parallel to oblique). These lateral moraines slope gently southwards at about 6 m per 100 m and can be linked to larger moraine ridges of the "hummocky" type (e.g. NC 399 257; Fig. 4.9). The upper transition from lateral, boulder-strewn moraines to solifluction lobes and sheets descends in altitude towards the south, linking up with a large terminal moraine around NC 413 244.

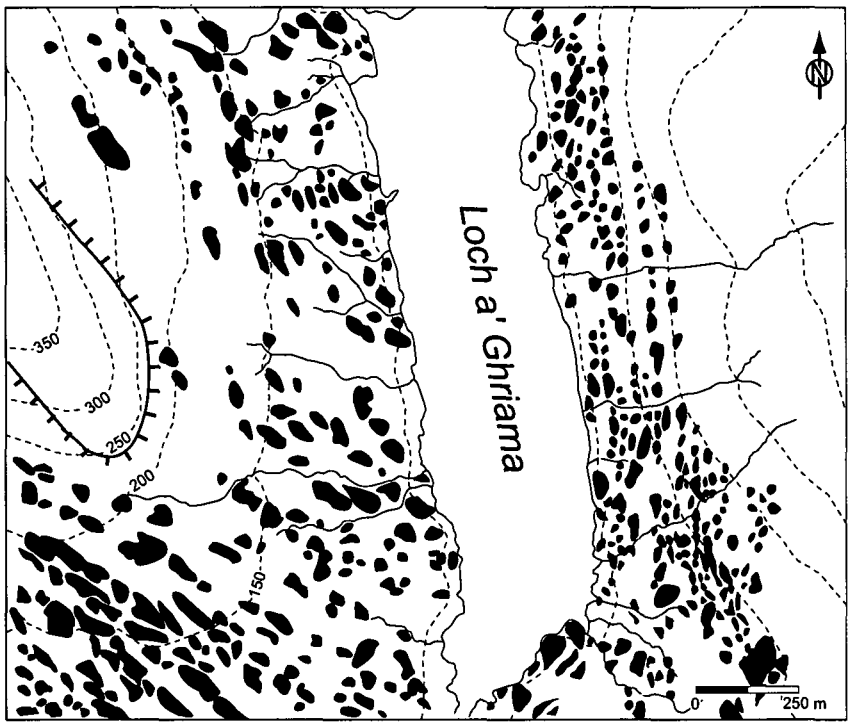


Fig. 4.9 Geomorphological detail of lateral moraines around the south-eastern side of Loch a'Ghriama. For key see Appendix 2.

4.2.2.3 Northern part of Bealach nam Meirleach

Around the northern part of Bealach nam Meirleach, lateral moraines on the southern slopes of Suil a'Bhadain Duinn take the form of boulder-strewn ridges and benches of sediment perched against weathered, ice-moulded bedrock (NC 440 387). They reach heights of up to 10 m, lengths of up to 100 m and widths of up to 20 m (Fig. 4.10). The uppermost of these can be traced to the valley floor where it joins up with a large, arcuate moraine ridge that can in turn be linked to lateral moraines on the northern side of Clach-bhuaile (NC 450 383). The latter consist of sediment accumulations of similar proportions to those described above. However, these are replaced southwestwards by lower bouldery ridges and benches that attain heights of ca. 3 m, widths of up to 20 m and maximum lengths of ca. 50 m. The maximum altitude of moraines reached in this area is ca. 400 m and the average gradient ca. 5 m per 100 m.

Fig. 4.10 Photograph of lateral moraines around the northern part of Bealach nam Meirleach as seen from the northeast (ca. NC 436 383).

4.2.3 Fluted moraines

A different type of moraine can be found on many corrie floors and downvalley of larger bedrock steps. This type of moraine is elongated, sharp-crested and linear and is restricted to the central part of the valley or corrie bottoms, trending parallel or subparallel to the valley axis. It differs markedly in morphological appearance from the two types above in that moraines in this category are not connected to either "hummocky" or lateral moraines. By analogy with previous studies (e.g. Peacock, 1970; Sissons, 1967, 1972, 1976, 1977; Hodgson, 1986; Ballantyne, 1989, 2002a; Benn, 1990, 1992a; Bennett, 1995; Wilson and Evans, 2000) these moraines are referred to as fluted moraines or flutes.

Flutes in modern terrestrial glacial environments are relatively small forms, commonly being between 0.5 and 3 m in height, < 3 m in width and up to hundred metres in length, and are found on the surface of recently deglaciated terrain, especially at temperate glacier margins (e.g. Gordon *et al.*, 1992; Benn, 1994; Krüger, 1994; Hart, 1995; Eklund and Hart, 1996; Evans and Twigg, 2002). The formation of these small flutes has been linked to subglacial till deformation that allows liquefied sediment to be squeezed into subglacial cavities initiated in the lee of boulders lodged in the ground surface. As these cavities propagate downglacier, the flutes can grow, forming long ridges on the surface which record the former ice flow direction (Gordon *et al.*, 1992; Benn, 1994; Hart, 1995;

Eklund and Hart, 1996; Benn and Evans, 1998; Evans and Twigg, 2002). This small type of flutes can be superimposed on larger, streamlined, subglacial landforms (e.g. van der Meer, 1983).

This type of modern flute differs from those found in Scotland since the latter can be several metres high and several hundred metres long. To distinguish them from the smaller ones, Benn (1992a) and Bennett (1995) have termed them megaflutes. Their mode of formation has never been investigated in detail, largely due to the absence of larger natural or man-made exposures (Hodgson, 1986; Benn, 1992a; Bennett, 1995). Benn (1992a) has suggested that streamlined landforms with a downvalley orientation on the Isle of Skye are likely to form part of a continuum ranging from small flutes to megaflutes and small drumlins (cf. Rose, 1987). The presence of flutes indicates basal sliding and thus the presence of temperate ice (Gordon *et al.*, 1992; Benn, 1994; Hart, 1995; Eklund and Hart, 1996; Benn and Evans, 1998; Evans and Twigg, 2002).

Below, a description of the morphological characteristics of fluted moraines in the study area will be given using the best examples available. The full distribution of flutes in the study area is displayed in Appendix 2.

4.2.3.1 Arkle

The clearest and best-developed flutes have been observed in the two corries east of Arkle (NC 311 454). Both corries contain ridges up to 250 m long, 2-5 m wide and up to 3 m high. In planform these flutes converge towards the corrie centre in a ca. 400 m-wide zone near the headwall, run parallel to each other in the middle part of both corries over a distance of 600 m and terminate at bouldery terminal moraines in the lower part of the corrie (Fig. 4.11). These flutes have already been mapped by Sissons (1977), but were not described in detail. They are significant since they record the flow direction of a small corrie glacier and can be used to constrain the former ice flow pattern with confidence.

4.2.3.2 Head of Strath Dionard

At the head of Strath Dionard (NC 362 472) numerous parallel flutes occur east of the eastern edge of Plàt Reidh. These attain heights of ca. 4 m, widths of ca. 10 m, lengths of ca. 100 m and indicate a northeasterly ice flow direction (Fig. 4.7a). To the north, however, the trend of these flutes progressively changes to a direction parallel to the valley axis over a distance of about 1 km indicating alignment of the flow direction with the trend of the Strath. On the eastern side of Strath Dionard, on the other hand, flutes are superimposed by moraines to give an apparently chaotic picture of numerous steep-sided, pointed mounds (Fig. 4.7b).

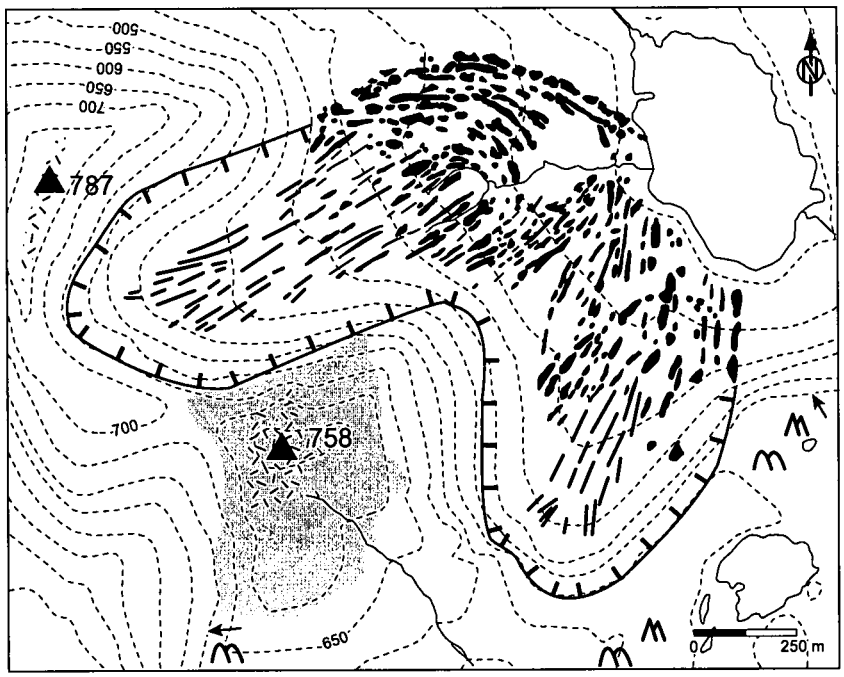


Fig. 4.11 Geomorphological map showing the distribution and orientation of flutes and moraines in the corrie complex east of Arkle. For key see Appendix 2.

4.2.3.3 Coire na Phris

In the upper parts of Coire na Phris (NC 382 393) flutes with heights of up to 5 m, widths of up to 13 m and lengths of 300 m converge towards the valley axis south of a large bedrock step with a vertical drop of 200 m over a horizontal distance of 400 m forming a fan-shape (Fig. 4.12). The long axis of these flutes is parallel to the ice flow direction as inferred from roches moutonnées and ice-moulded bedrock outcrops amongst them. The coincidence of flutes and steep bedrock slopes is particularly striking in this area.

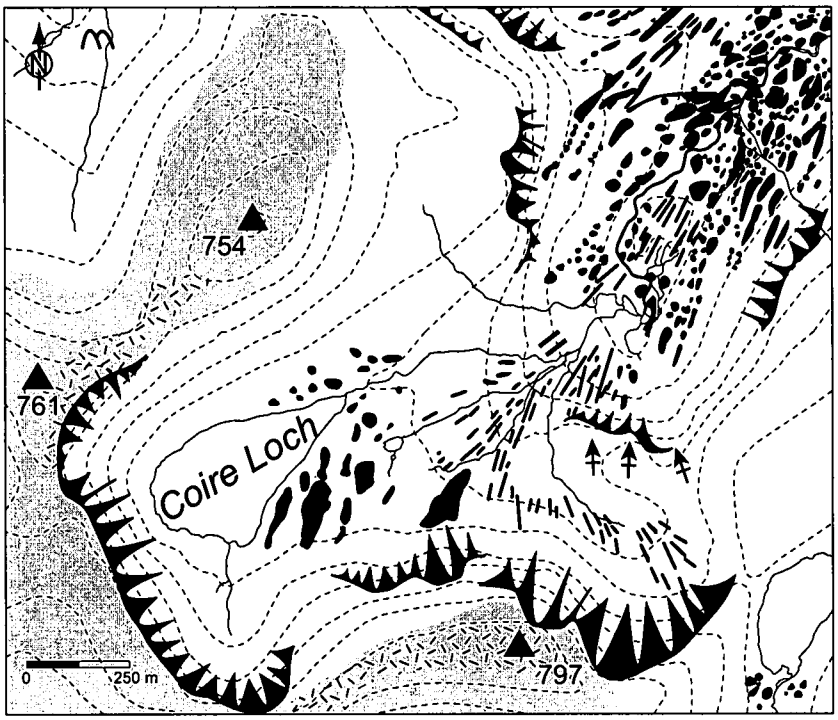


Fig. 4.12 Geomorphological detail of fluted moraines around Coire na Phris. For key see Appendix 2.

4.3 Zooming in on "hummocky moraine" formation: sedimentological data

4.3.1 Sedimentological case studies

In order to solve the problems of model applicability and associated questions regarding the significance of "hummocky moraines" in glacier reconstructions (Chapter 1), all available sections were logged in detail. In contrast with most other study areas, numerous exposures have been created for gravel extraction purposes along estate tracks, providing ideal exposure conditions through the whole widths of individual moraine ridges. In addition, landslides, shallow surface exposures and river cuttings were enlarged with a trenching tool to increase the data set and to enhance the accuracy of the interpretations made from these data. Fifty-two exposures in moraines were logged in detail. To avoid unnecessary repetition and to emphasise characteristic lithofacies, lithofacies associations and sedimentary structures that occur in "hummocky moraines" in the study area, a representative sub-sample of seven case studies will be described and interpreted in detail. The remainder of the data is presented in Appendix 1 to allow the most important information to be readily accessed by the reader.

4.3.1.1 Case study 1

Description

Moraine ridge SLK 1 (NC 393 344; 215 m a.s.l.) is located in at the foot of the western slopes of Ben Hee where it forms part of a larger series of moraines stacked on the mountainside. In the valley bottom a series of "hummocky moraines" can be found trending obliquely across the slope in a

downvalley direction. Both suites of moraines are partly breached and/or bounded by glacial meltwater channels (Fig. 4.13a). Moraine SLK 1 is ca. 6 m high, 10 m wide, 30 m long and displays an asymmetrical cross-profile with a steeper ice-proximal rectilinear slope (ca. 32°) pointing NNW and a gentler distal rectilinear face (ca. 16°) dipping SSE. This latter gentle surface grades into the steep ice-proximal slope of the next moraine upslope that shows a similar cross-section and is of similar dimensions. The overall direction of ice retreat is reconstructed to have been towards the N and NNW in this area (Figs. 4.13a, b, c).

A trackside pit provides excellent exposure through the whole moraine perpendicular to its crestline. Sediment characteristics allow grouping into two facies associations, those on the ice-proximal (left, NW) and those on the ice-distal (right, SE) side. The latter consists of alternating units of matrix- to clast-supported stratified diamictons that reach thicknesses up to 1.1 m with the thickest units towards the top near the crestline (Fig. 4.13c). These diamict units generally taper out over a few decimetres in a distinct wedge shape. Individual units are typically separated by fine to medium sand units that are commonly between 0.01 and 0.1 m (Fig. 4.13d) but can reach thicknesses of 0.7 m, especially nearer the crestline of the moraine. Few interspersed fine to coarse gravel and occasional pebble-sized particles occur throughout the sequence near the crestline. Thinner beds of sand generally bend around the underside of embedded larger clasts. The sorting within the sand units increases while their thickness decreases towards the right (SSE) (cf. Fig. 4.13c, d). In a similar fashion the sand units show fining from predominantly medium and coarse sand near the crestline to silty very fine sand towards the right. On the far right-hand side of the section, internal stratification within the finer-grained units is apparent and reveals a sequence of gently dipping, subparallel, 2-5 millimetre-thick beds of silt and silty very fine sand unconformably overlain by horizontal units of fine to medium sand and clast-supported diamicton (Fig. 4.13d). All units are inclined subparallel to the surface slope of the moraine (ca. 16°).

Below the moraine crestline, the dip and direction of units changes abruptly over a distance of 0.1 m. Below the ice-proximal side of the moraine which dips at ca. 32°, units parallel the surface slope and have an almost constant thickness of ca. 1.2 m. This side of the section, in contrast to the right, consists mainly of inclined layers of stratified clast-supported diamicton that are intercalated with very thin continuous sand layers which taper out in sharp wedges in a downslope direction (Fig. 4.13c).

Interpretation

The alternation of diamict and sorted units on the right side of the moraine together with their uniform and gentle dip is consistent with an interpretation as supraglacial debris flow units and fluvial "wash" horizons, respectively. Distinct wedge shapes at the downslope ends of diamict units are interpreted as flow noses indicative of the deposition in slow cohesive debris flows (cf. Lawson, 1982, 1988; Benn,

1992a). The alternating deposition of diamictic and sorted sediment units can be explained by a combination of two factors, namely a stationary ice margin and variations in water content, the latter of which determine whether gravitational or fluvial processes dominate in fan formation at any given time (cf. Krzyszkowski and Zielinski, 2002). Laminations within the sand units and their laterally discontinuous fashion are compatible with deposition in sheets and shallow channels or rills, respectively (cf. Lawson, 1988). Thin, inclined beds of silt and silty very fine sand on the far right side are interpreted as a puddle fill which was then capped by slightly coarser sand during a subsequent sheet wash event. The fine internal stratification of the sand units, preservation of the underlying delicate silt layers and limited bed thickness suggests slow sedimentation rates and is compatible with deposition in an ice-marginal setting on a relatively gentle fan surface (cf. Gripp, 1975; Krüger, 1997; Krzyszkowski and Zielinski, 2002). The decrease in thickness and grain size within the sand units can be explained by waning flow velocity leading to downfan fining, a phenomenon commonly observed in ice-marginal fans (Zielinski and van Loon, 2000). Down-warping of thinner sand beds underneath clasts is interpreted as evidence of clasts sliding and falling from the ice surface onto the former fan surface causing localised deformation.

The steeper slope on the left side of the moraine is interpreted as the former ice-contact face that formed during collapse of the proximal part of the units as support by the ice was withdrawn during ice retreat. Limited compaction of the diamict units, clear stratification and interdigitation of sand and diamict units within a narrow, well-defined zone all support this interpretation. The lack of collapse features in this zone suggests that no dead ice bodies were present at the time of glacier retreat and that this ice-contact face is merely the result of withdrawal of support by the ice margin.

This moraine is thus interpreted as a *terrestrial ice-contact fan* following Benn (1992a), Benn and Evans (1998), Zielinski and van Loon (2000) and Krzyszkowski and Zielinski (2002). This form is a scaled version of those classed as Type A (mass flow deposits-dominated ice marginal fans) by Krzyszkowski and Zielinski (2002) in that it bears striking similarities in its sedimentological composition and characteristics, but is about an order of magnitude smaller. The morphological asymmetry thus reflects a steeper proximal ice-contact slope with material at the angle of repose and a much gentler fan surface representing the depositional slope.

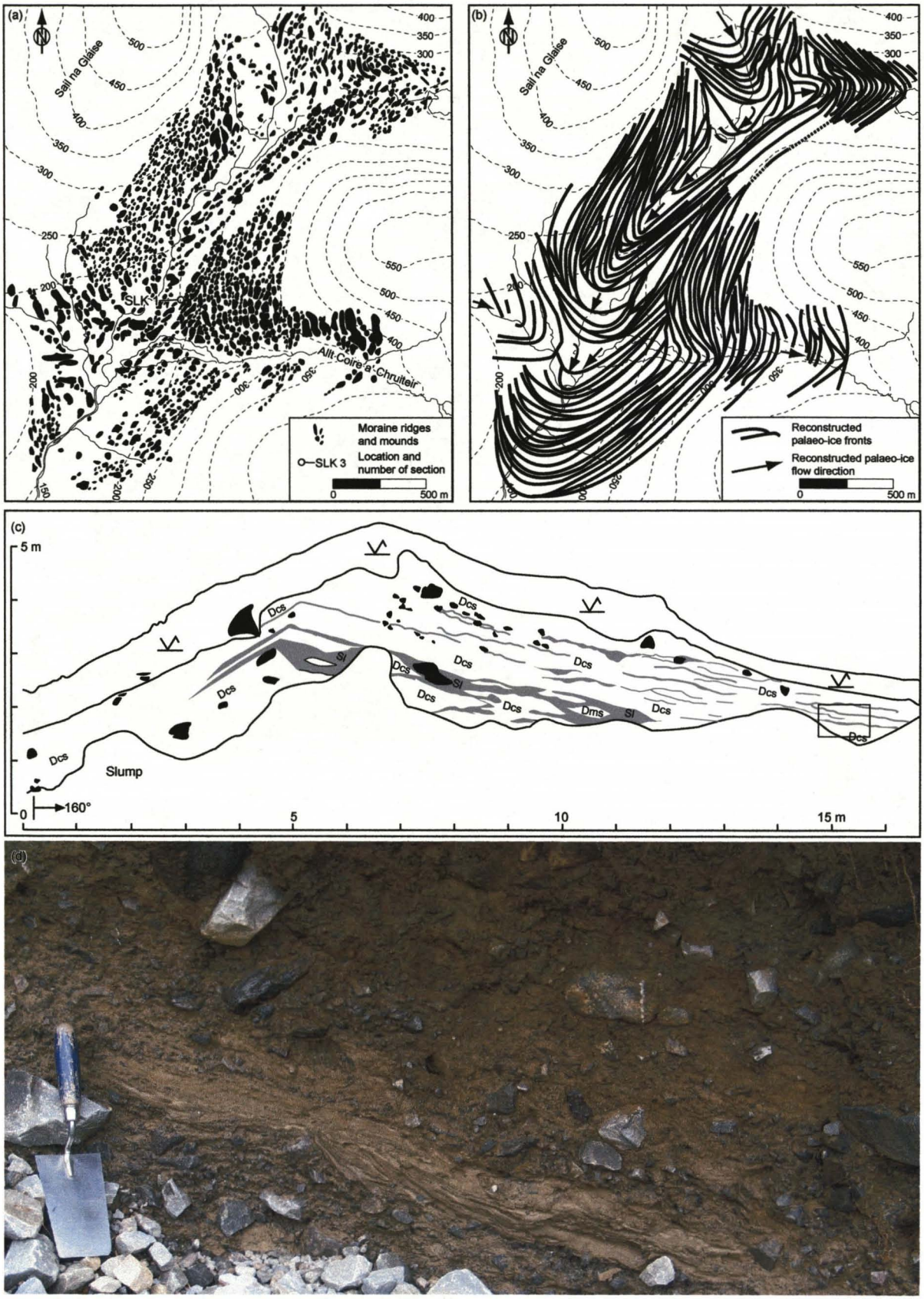


Fig. 4.13 (a) Geomorphological map of the southern part of Bealach nam Meirleach showing "stacked" moraines on the south-western slopes of Ben Hee and location of SLK 1. (b) Reconstruction of palaeo-ice fronts and ice flow directions. (c) Section log of section SLK 1; for key see Fig. 2.1. (d) Close-up photograph of units at the distal side shown by frame in (c).

4.3.1.2 Case study 2

In the lower reaches of Coire Eas na Maoile (Fig. 4.14a, b), two asymmetrical moraine ridges are exposed through almost their whole width and height. The setting and quality of the exposures presents the rare opportunity to study the sedimentology of two adjacent moraines and test whether they formed in succession or during one event.

Description of moraine SLK 13

The northernmost moraine, which is located downvalley of the example presented below (NC 3575 3477; 72 m a.s.l.; SLK 13) is asymmetrical and bounded by rectilinear slopes with gentle (15°) northern distal and steeper (32°) southern proximal gradients. An enlarged exposure reveals alternating units of clast-supported, stratified diamictons (A) and interbedded units of (B) fine sand to granule-sized gravel, (C) openwork granule to coarse gravel and (D) well-sorted very fine to fine sand with occasional silt layers (Fig. 4.14c). The right-most 3 m of the moraine are not exposed and the top is partly obliterated by vegetation and peat. In general, these sediments are very similar to those described in Case study 1. Facies (A) contains subrounded, facetted clasts with a-axes up to 0.8 m. Layers of this facies attain thicknesses of 0.5 m and are particularly well developed in the left (northern) part of the exposure where they dip at ca. 15° subparallel to the moraine surface. Interbeds of sorted sediments (B-D) are rare and only reach thicknesses of 1-2 cm. A discontinuous layer of alternating facies B and D is exposed in the lower parts of the sequence and shows signs of gentle folding, particularly at the contacts with facies A and B (Fig. 4.14c). Folded lenses and layers of facies B and D occur throughout the sequence.

Interpretation of moraine SLK 13

The succession of stratified diamict and sorted finer-grained units is compatible with an interpretation as a terrestrial ice-contact fan formed at a temporarily stationary ice margin during overall southward retreat. Gravel and granule units suggest that deposition occasionally occurred at higher flow velocities than those described for the finer-grained units identified for SLK 1 above. These gravel lenses might represent lag bodies created by winnowing of fines during clean water runoff off the glacier surface (Lawson, 1988). Deformation structures within the finer-grained unit (D) are indicative of proglacial push from the right (south). This episode of lateral compression is inferred to have occurred after fan formation and initial retreat during a small-scale readvance of the glacier. This is supported by the largely unaltered fan morphology with a clear proximal rectilinear ice-contact face and a distal fan surface, but also by the fact that the sedimentary units are largely undisturbed. Similar relationships between fan formation along a stationary ice margin and proglacial deformation during a subsequent readvance were noted by Benn (1992a) in terrestrial ice-contact fans on the Isle of Skye and for much larger ice-contact fans in Poland by Krzyszkowski (2002).

Description of moraine SLK 12

This moraine (NC 3582 3465; 75 m a.s.l.; SLK 12) is located upvalley of SLK 13 and merges with the proximal (right) slope of the latter (Fig. 4.14a). SLK 12 too displays a clear asymmetry with a gentle distal rectilinear face dipping to the left (north) at ca. 20° and a steeper proximal rectilinear face dipping at ca. 27° to the right (south). Three lithofacies can be distinguished (Fig. 4.14d):

(1) A lower sandy-gravelly, clast-supported, stratified diamicton with clast a-axes ≤0.4 m. This facies is of medium compactness due to interlocking clasts within clast clusters, voids in which are sometimes infilled by finer material. The lower diamict dips with ca. 10° towards the left (north). Its stratified appearance is enhanced by laterally discontinuous lenses of sorted silty to very fine sandy, matrix-supported, stratified diamicton with most clasts in the granule and fine gravel fraction. Individual lenses reach maximum thicknesses of 0.1 m.

(2) Discontinuous layers of up to 0.1 m-thick, massive very fine to medium sand with occasional coarse sand grains occur within facies (1). This facies displays numerous deformation structures, e.g. small-scale overturned folds and undulating contacts with the surrounding diamicton (Figs. 4.14d, 4.15).

(3) A unit of massive very fine sandy silt can be traced through much of the lower part of the exposure. It reaches a maximum thickness of 0.6 m and contains numerous signs of pervasive deformation on a variety of scales (Figs. 4.14d, 4.15). The smallest examples are 5 cm-wide dykes of sand extending upwards into the surrounding diamicton by as much as 0.1 m. Internally, such dykes show widespread wavy laminations that are frequently disrupted, particularly around clasts. Small-scale reverse faults with a maximum displacement of 2 cm are frequent at the contact with the lower diamict. At the larger scale, the silt layer is partly folded and bifurcates towards the lower left side (north) of the section. The upper branch is heavily disturbed in an overturned fold.

The lower part of the moraine is overlain unconformably by a sandy clast-supported, stratified diamicton (facies 4 in Fig. 4.14d) which occupies a series of channel and half-channel structures. This erosive unconformity is most obvious where the upper diamicton truncates the branches of facies (3). Part of the overturned fold described above is folded back on itself along this contact. The upper diamict unit dips at ca. 8-10° towards the left (north); deviations from the angle of dip of the underlying diamict are apparent.

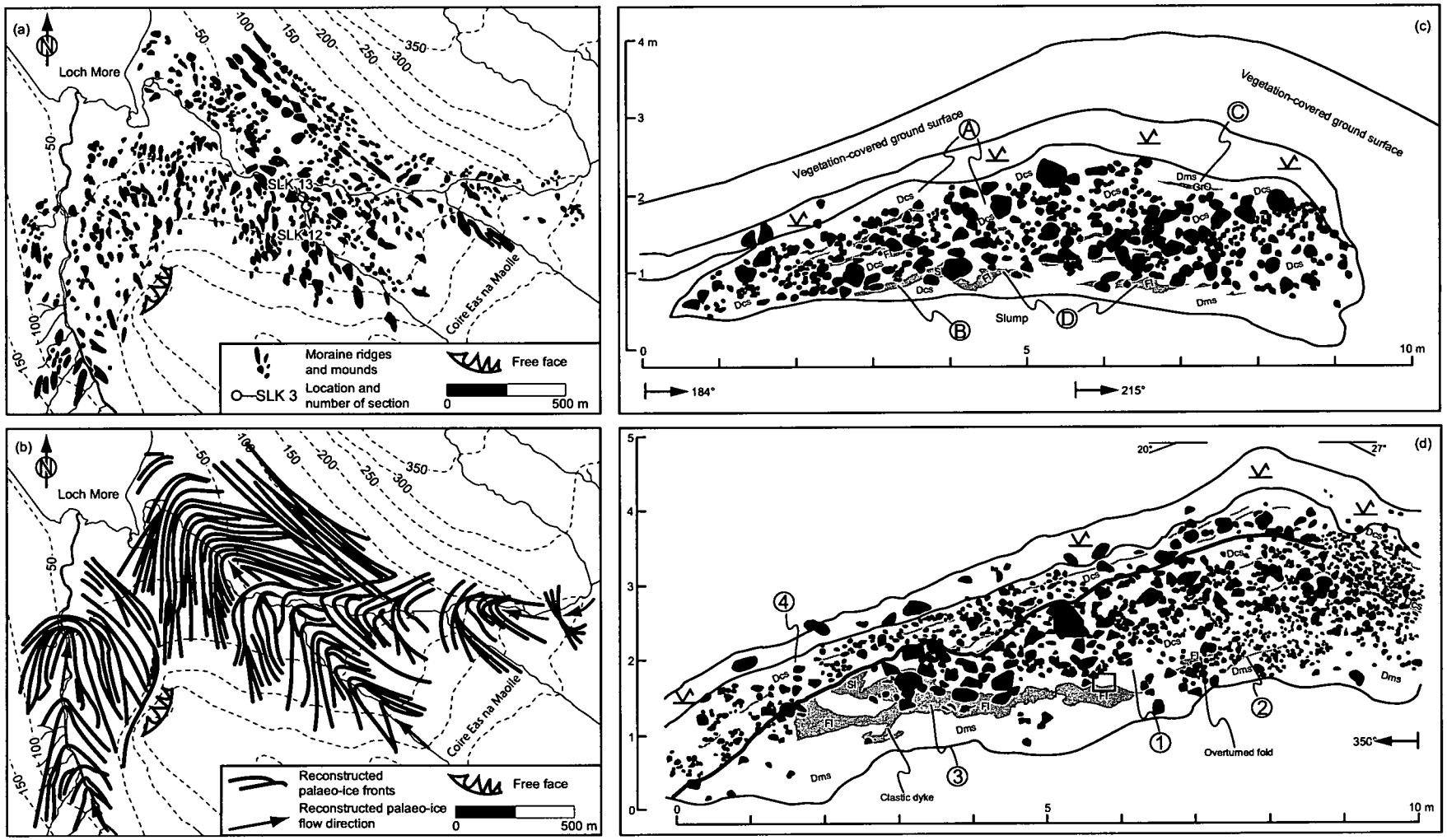


Fig. 4.14 (a) Geomorphological map of Coire Eas na Maoile showing moraine mounds and ridges and the locations of SLK 13 and SLK 12. (b) Reconstruction of palaeo-ice fronts and ice flow directions. (c) Sedimentary log of section SLK 13; for key see Fig. 2.1. (d) Sedimentary log of section SLK 12; for key see Fig. 2.1. Circled numbers and letters refer to units mentioned in the text.

Interpretation of moraine SLK 12

The facies in the lower part of this moraine are interpreted as supraglacial debris flows and fluvial deposits and are indicative of a terrestrial ice-contact fan formed at a temporarily stationary ice margin. Interlocking of boulders is inferred to result from blocks sliding and toppling off the ice front and successive blocks becoming trapped on their stoss side. The infill by relatively coarse sediments (no silt) is interpreted as a result of trapping of sediment transported in shallow flows in the lee of such boulders.

The deformation structures are indicative of proglacial deformation and are interpreted as evidence for a glacier readvance into this fan following formation and initial retreat. The continuous silt layer has probably acted as a decollèment surface during push as inferred from the concentrations of deformation structures within, and in the near vicinity of, this layer. In particular, the numerous injected layers of sand into the surrounding diamict units are indicative of liquefaction and hydrofracturing and are interpreted as water escape structures. They form as a result of compression under impeded drainage conditions (cf. van der Meer *et al.,* 2001; Rijsdijk, 2001) or due to differential loading, such as during deposition of a debris flow on top of a thin aquifer (e.g. Lawson, 1988). It appears that the surrounding diamict and silt was sufficiently dense to over-pressurise water contained within thin sand lenses in their vicinity to cause widespread hydrofracturing. The water escape structures are comparatively small and support an interpretation of localised stress transmission during a short-lived readvance.

The upper diamicton that truncates the top left of this prominent silt layer is interpreted as a later debris flow event that also partly eroded the existing fan surface. Basal traction gravels and erosive scours are sedimentary features compatible with such an interpretation (cf. Lawson, 1981, 1982, 1988; Benn, 1992a; Krzyszkowski and Zielinski, 2002). This addition of new material is likely to have caused some of the surficial folding of the upper parts of the silt layers as a response to drag forces exerted by cohesive debris flows.

This terrestrial ice-contact fan thus formed by a similar two-stage process to SLK 13 described above. In this case, however, the readvance was longer-lived, causing severe disruption of the sediment pile along a basal decollèment surface. Additional debris flows were also able to modify the pre-existing fan surface.

Significance of SLK 12 and SLK 13

The sedimentary architecture of both moraines demonstrates that each moraine represents a separate terrestrial ice-contact fan and is the product of a two-staged process of (1) fan formation and (2) subsequent deformation during a readvance. This indicates an active, incremental and oscillatory retreat mode. Both moraines also display a marked asymmetry with steep proximal and gentle distal slopes, the proximal slope reflecting the former ice-contact face with material at the angle of repose.

Thus, these moraines demonstrate that they formed sequentially in two separate episodes of fan formation and deformation. This evidence is clearly incompatible with the englacial trusting model, which interprets rectilinear ice-proximal slopes as thrust planes rather than angle of repose debris slopes and argues that multiple moraines form contemporaneously during one glacial event.

Fig. 4.15 Close-up photograph of overturned fold at the deformed boundary between diamicton (Dcs) and underlying laminated sand (Sl) and silt (Fl) layers and smaller faults within Sl. Note the occurrence of thin silt and fine sand lenses in the lower diamict. Location is given by frame in Fig. 4.14d. The blade of the scraper is 7 cm long.

4.3.1.3 Case study 3

Description

This moraine complex at NC 4385 3827 in the northern part of Bealach nam Meirleach (230 m a.s.l.; SLK 3) consists of three individual moraine ridges with rounded crestlines separated by peat-infilled basins (Fig. 4.16a). The complex is ca. 7 m high at its highest point, up to 10 m wide and 12 m long. Most of the moraine complex has been disturbed by quarrying activity so that the original cross-sectional shape (e.g. asymmetry) cannot be determined. A section oriented subparallel to the crestline exposes the uppermost 2 m of the main mound.

Main section. Three lithofacies can be distinguished (Fig. 4.16c). (1) A succession of sandy clast-supported, stratified diamicton units attain maximum thicknesses of ca. 1.5 m and dip at up to 52° towards the left (west to WNW). Contacts between individual beds are largely conformable.

Intercalated within the diamictons is unit (2), a ca. 0.4 m thick very fine to coarse sand that commonly contains lenses of very fine sand to silt with occasional fine gravel and few outsized clasts. A sharp basal contact striking 198° and dipping towards the WNW by 68° separates it from a unit of clast-supported, stratified diamicton. An isolated lens of horizontally stratified granule gravel is embedded within this unit and shows signs of soft sediment deformation (Fig. 4.16c). Deformation structures occur in a ca. 15 cm-wide zone, visible in attenuated lenses of silt. Unit (3) is a loose, clast-supported, massive diamicton with a maximum thickness of 1.3 m. It is overlain by a subhorizontal loose, sandy, clast-supported, stratified diamicton that attains maximum thicknesses of 0.7 m, contains subrounded and subangular clasts with a-axes up to 0.5 m and forms a continuous carapace on top of the sediments exposed in the pit described below.

Pit section. In a smaller pit dug both parallel with and at right angles to the crestline, four units can be distinguished (Fig. 4.16c). Unit (A) corresponds to (1) described above and caps the whole frontal wall. (B) is a matrix-supported stratified diamicton with a maximum thickness of 0.8 m and few clasts with a-axes <3 cm interspersed in a medium to coarse sandy matrix. Subhorizontal stratification is picked out by discrete layers of medium sand-sized particles. Few deformation structures, best described as flexures, occur in the upper left hand part of this unit. The upper and lower contacts are sharp, and the lower is shown to be erosional by the presence of channel structures and drag features in silt lenses just above the contact. (C) is a 0.6 m thick body of planar cross-bedded fine to medium sand that has been affected by high-angle reverse faults with displacements of up to 2 cm. Unit (C) has a sharp contact with the underlying unit (D), a sandy, clast-supported, stratified diamicton. These four units are also exposed in the sidewalls where they dip more steeply towards the NNE (Fig. 4.16c). Here, unit (A) tapers out after ca. 0.1 m from the left side of the right (southern) wall, the remainder being occupied by units (B) and (C) which both exhibit deformation structures. The upper contact of unit (B) with (A) is sharp and relatively straight; the lower contact by contrast is highly undulating and contains several smaller fold traces. Steeply-inclined planar cross beds occur within this unit, and the lower contact strikes 290° and dips at 55° to the NE. Unit (C) exhibits both low-angle normal and reverse faults and high-angle reverse fault sets subparallel to the cross beds with displacements between 1 cm and 4 cm. *En echelon* high-angle normal fault sets strike 328° and are inclined between 40° and 60° towards the NE and show similar displacement. Their spacing between 0.5 and 2 cm gives a fragmented appearance to individual inter-fault blocks.

Interpretation

The sedimentological composition of this moraine is compatible with an interpretation as a terrestrial ice-contact fan. However, the steep inclination of the units (in the left part of the section to the WNW, in the pit towards the NE) cannot be explained in terms of the primary depositional slope. The wealth

of deformational information contained within unit (C) and the sharp contact between sandy and diamicton units in the main section interpreted as a high-angle normal fault (Fig. 4.16c) indicate ice pushing into the sediment pile resulting in a dislocation of the entire fan. Different scales of deformation can be distinguished, small-scale faults and folds as recognised in unit (C) and large-scale steepening of the entire landform.

This evidence suggests a readvance of the glacier into the previously formed terrestrial ice-contact fan which was probably more sustained than in the cases reported above. The deformation history contained within the sediments indicates a two-stage process. First, the smaller-scale compressional reverse fault sets and large-scale steepening of the units were formed by forward push by the glacier readvancing from a southern direction. Secondly, a subjugate set of normal faults developed in both parts of the section and is interpreted as relaxation response following steepening of the ice-contact fan (cf. Humlum, 1985). It appears to mark a stage during which slippage occurred in the sediment pile as a response to the considerable oversteepening of a much shallower depositional slope of ca. 10-20° to the present ca. 60°. The exact direction of ice push cannot be reconstructed due to the limited extent of the exposure but ice flow from the south-eastern quadrant as suggested by structural data (Fig. 4.16c) is fully compatible with the available geomorphological evidence from the surroundings (Fig. 4.16b).

4.3.1.4 Case study 4

Description

A moraine ridge (NC 4422 3871; 220 m a.s.l.; SLK 4), ca. 4 m high, 15 m wide and 45 m long with a straight crestline orientated NW-SE (126°) has been excavated for gravel extraction along a width of 6 m. Due to removal of the upper metre of sediment and peat, the original asymmetrical cross-profile with a steeper side to the left (SW) and a gentler slope on the right (NE) of this moraine is not discernible on the sedimentary log. The sediments exposed in this ridge can be grouped into two distinct associations, one in the left third and one occupying the right-hand two-thirds (Fig. 4.16d), the junction between which coincides with the crestline. The units on the right consist of alternating loose clast-supported, stratified diamicts and sorted sediments, mostly sand and very fine silt, that both dip at about 11° to the right (NE). The diamict units reach thicknesses up to 1.0 m with only occasional thin stringers of silt or very fine sand while larger boulders with a-axes ≤0.5 m are frequently found within the diamict units. Finer-grained sediment units in the right-hand part of the section display numerous small-scale overturned folds with a concentration of the most severe deformation structures near the bottom of the exposure (Fig. 4.16d). Here, a formerly subhorizontal bed of gravel adjacent to a clast has been rotated and now dips at about 40° to the right (NE). Near the right-hand surface, a sand bed with a distinct flatiron shape can be found underneath a clast with a flat underside, and this in

turn is underlain by a continuous 10 cm-thick layer of very fine sandy silt that shows localised folding in the vicinity of the clast.

In the left third of the section, similar lithofacies are capped by 30 cm of matrix-supported, stratified diamicton (Fig. 4.16d). Clasts in the diamict units are dislocated by reverse faults filled with silty very fine sand. Unlike those in the right-hand two-thirds, all lithofacies units on the left take the form of elongated, attenuated and boudin-shaped lenses and contain flame-structures throughout, undulate across the width of the left-hand side and dip by about 29° to the left (SSW). One isolated sand body contains a massive silt bed with overturned folds that display small reverse faults in their hinge-lines (Fig. 4.16d).

Interpretation

The alternation of the three lithofacies units described above is compatible with an interpretation as terrestrial ice-contact fan deposits. The left-hand side of the exposure contains evidence for widespread and severe deformation that is not easily reconcilable with proglacial compression. The presence of boudins and flame structures, fractured rock with intervening silt layers and a gently-undulating appearance in this part of the moraine is interpreted as evidence of subglacial shearing. Stresses transmitted into the distal (right) part of the fan were large enough to create the deformation structures observed there. These in turn are compatible with proglacial deformation caused by ice pushing into, and partly overriding, this fan from the left (SW) during a readvance, causing widespread folding. This lateral compression also led to the rotation of previously gently-inclined units, i.e. the gravel accumulation in the lower part. Rotation of one clast near the right-hand surface ploughed material up in front of it to form the flatiron shaped sand body. This appears to be associated with localised decollèment within the underlying silt layer. Close to the crestline, a variety of both ductile and brittle deformation structures indicates the location of the transition zone between subglacial and proglacial deformation.

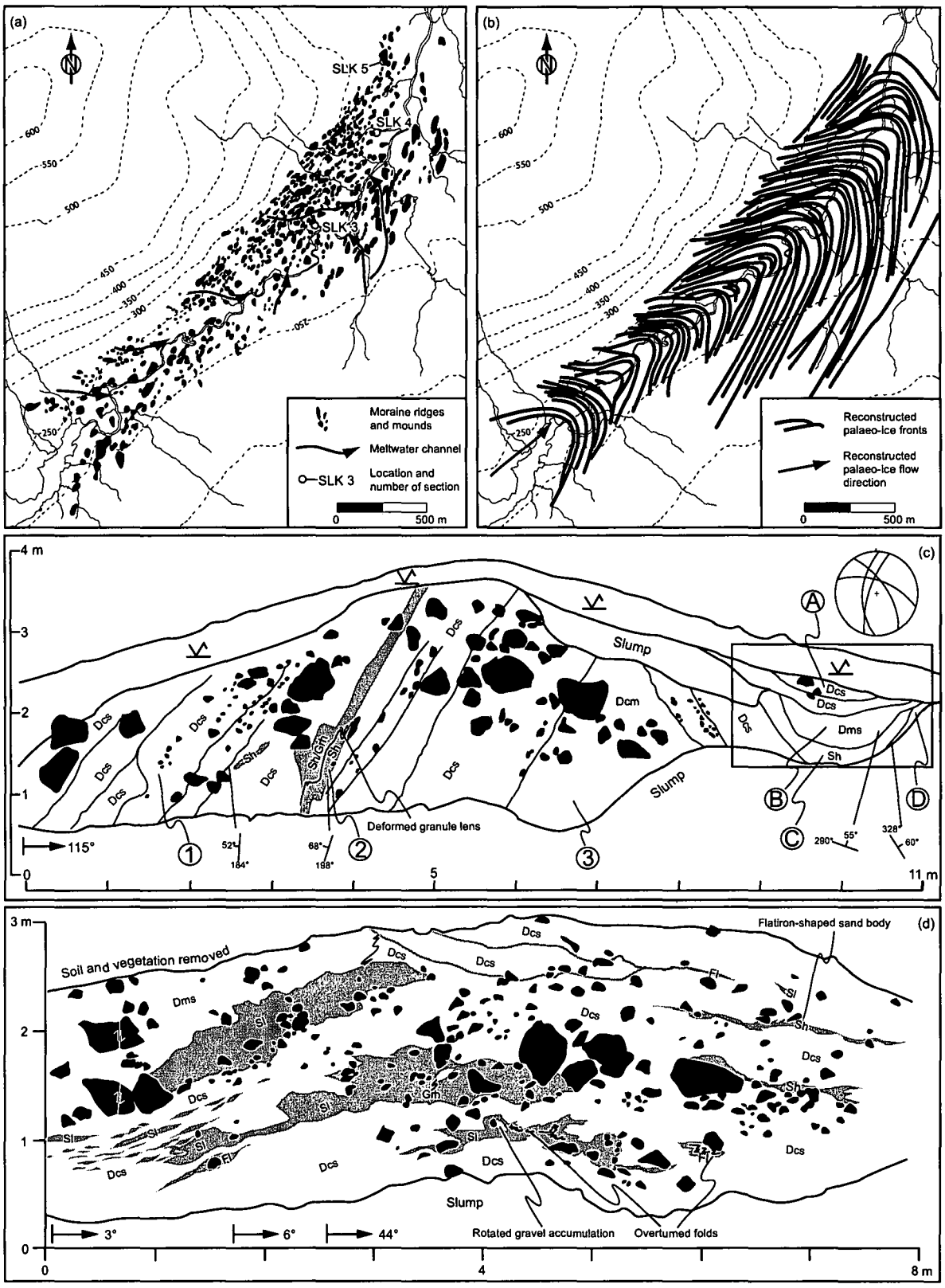


Fig. 4.16 (a) Geomorphological map of the northern part of Bealach nam Meirleach showing moraine mounds and ridges and the locations of SLK 3 to SLK 5. (b) Reconstruction of palaeo-ice fronts and ice flow directions. (c) Sedimentary log of section SLK 3; for key see Fig. 2.1. The location of the frame shows the location of the pit section described in the text. (d) Sedimentary log of section SLK 4; for key see Fig. 2.1. Circled numbers and letters refer to units mentioned in the text.

4.3.1.5 Case study 5

Description

Moraines along the eastern shore of Loch Shin differ from the general character of those described above in that they are only between 2-3 m high, up to 100 m long and 30 m wide and show an almost symmetrical profile with maximum slope angles of 12° giving them a subdued or smoothed appearance. They form discrete moraines that are somewhat wider than moraines elsewhere in the area and appear to be much more continuous features transverse to former ice flow. Depth to bedrock cannot be established due the presence of beach gravel and the water level of the loch. Unfortunately, this area is largely forested hindering mapping of smaller moraines, both on the ground and from aerial photographs (Fig. 4.17a, b).

Exposures in the moraines along the loch shore (NC 3918 2495; 100 m a.s.l.; SLK 17) reveal three alternating lithofacies which are exposed along a width of ca. 8 m (Fig. 4.17c). Facies (1) is a loose, stratified, clast-supported diamict (Dcs) that reaches a maximum thickness of 0.2 m and contains numerous clasts with maximum a-axis-lengths of 0.4 m. Facies (2) is a slightly-silty fine sand (Sl, Fl) that contains occasional outsized granules, exhibits wavy laminations and reaches a maximum thickness of 0.1 m. Both facies exhibit frequent lateral thickness changes and take the general shape of elongated or streaked-out lenses or pods. Beds of facies (2) commonly bifurcate around clasts or lenses of facies (1), showing partitioning and rejoining, while individual beds appear attenuated, tapering out within the space of a few millimetres on either side. Although discontinuous, individual lenses of the same facies are laterally aligned while the different facies themselves are subparallel (Fig. 4.18). These units form a series of long-wavelength, low-amplitude anticlines and synclines, which do not reflect the surface of the moraines (Fig. 4.17c). Fold axis measurements shown in the inset of Fig. 4.17c indicate an east-west orientation, roughly perpendicular to the section wall.

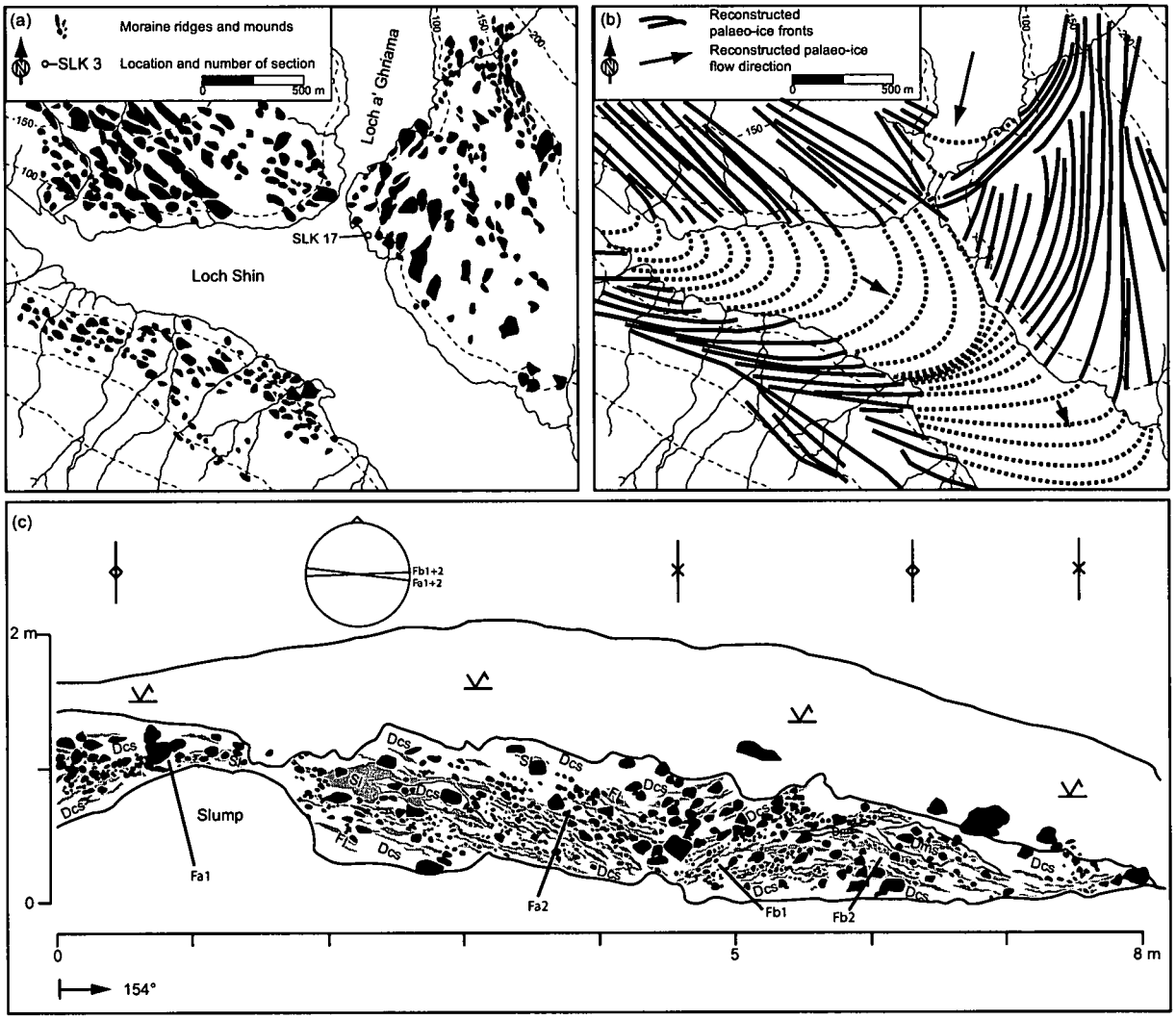


Fig. 4.17 (a) Geomorphological map of the northern part of Loch Shin showing subdued moraine mounds and ridges and the location of SLK 17. (b) Reconstruction of palaeo-ice fronts and ice flow directions. (c) Sedimentary log of section SLK 17; for key see Fig. 2.1. Approximate position of synclinal and anticlinal fold axial planes is shown as viewed.

Interpretation

The sedimentary composition of the units and their pattern of alternation suggests that they formed as a series of terrestrial ice-contact fans during glacier retreat. The wealth of deformation structures can be best interpreted as the result of two processes. The long-wavelength, low-amplitude folding is indicative of larger-scale proglacial push as recognised in larger exposures elsewhere (e.g. van der Wateren, 1999). Proglacial push leading to this deformation was directed from the north (Fig. 4.17c). This implies glaciers sourced in the mountains and fits the evidence in the surrounding area well (Fig. 4.17a, b). The units themselves contain ubiquitous features indicative of subglacial simple shear and can hence clearly be distinguished from the lateral compression typical of proglacial environments (e.g. Hart and Boulton, 1991; van der Wateren, 1995, 1999; Benn and Evans, 1996, 1998; Benn and Clapperton, 2000; van der Wateren *et al.,* 2000; Golledge, 2002; McCarroll and Rijsdijk, 2003). In

particular, units that bifurcate around clasts or inclusions have also been observed in glaciotectonised sediments by Benn and Evans (1996). In summary, the subdued character of the moraine ridges and the out-of-phase folding of the underlying sediment is due to a three-stage process of successive fan formation, proglacial folding and subsequent overriding, either during the same, or a later, readvance.

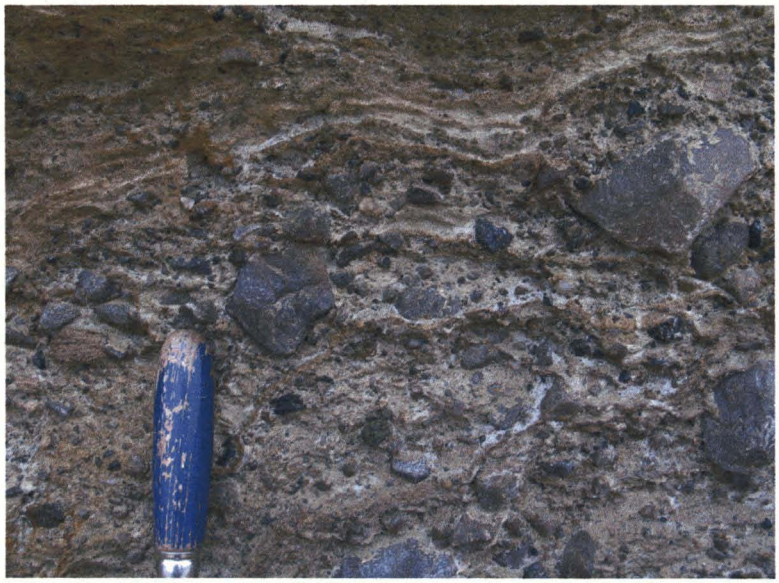


Fig. 4.18 Close-up photograph of glaciotectonised debris flow and fluvial wash units exposed along the eastern shore of Loch Shin. The wooden handle of the trowel is 10 cm long.

4.3.1.6 Case study 6

Description

This moraine (NC 44317 39171; 217 m a.s.l.; SLK 5) is ca. 5 m high, 20 m long, 40 m wide and contains an exposure in a subsidiary ridge in its left-hand part (Fig. 4.16a, b). The ice-proximal slope (on the left, SE) is irregular while the distal slope dips by about 25° to the right (NW). Three lithofacies can be distinguished (Fig. 4.19). (1) This facies consists of alternating very fine and fine sand layers which are horizontally stratified and contain numerous lenses of fine to coarse sand with interspersed fine gravel and very fine sand and silt couplets. The couplets typically reach thicknesses between 1-5 mm while the sand layers are up to 5 cm thick. High- and low-angle reverse and – to a lesser extent – normal faults are common, with the silt and very fine sand couplets acting as marker horizons (Fig. 4.19). The mean strike of the fault sets is towards the ESE (111°) with a mean dip of 35° towards the SW (Fig. 4.19). Displacement of beds is commonly between 2-9 cm with a maximum of 25 cm. Unit (1) is overlain by a clast-supported, stratified diamicton (2) with a fine to medium sandy matrix and clasts ≤0.4 m. Stratification in this unit is subparallel to that of unit (1). The contact between the two units is unconformable and undulating, forming an overturned fold in the right hand part of the exposure (Fig. 4.19). In the fold core, the sand units are deformed in a brittle fashion. The

contact with unit (3), a compact, clast-supported massive fine to medium gravel, which extends into unit (1) as a gently folded wedge, is partly obscured by soil formation (Fig. 4.19).

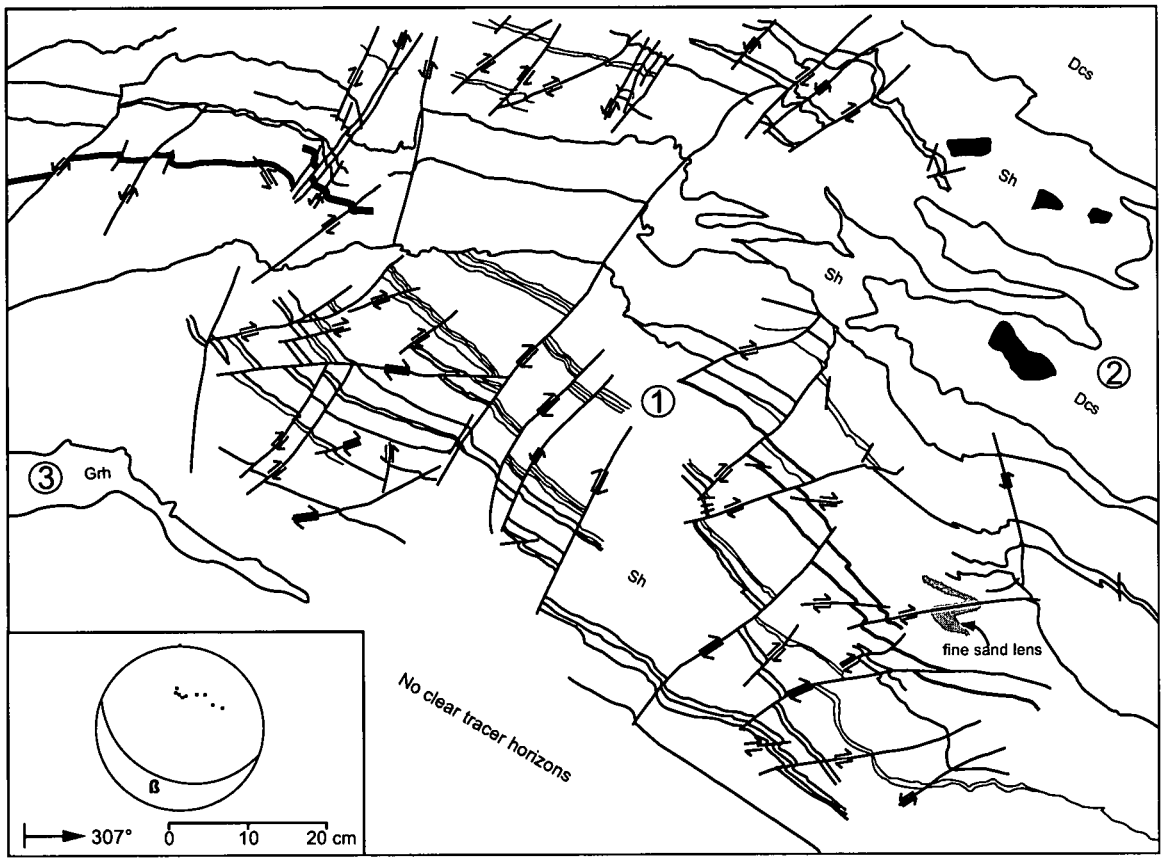


Fig. 4.19 Sedimentary log of section SLK 5; for key see Fig. 2.1. The polar diagram represents a lower hemisphere Schmidt net (for explanation of structural data see text). Circled numbers refer to lithofacies mentioned in the text.

Interpretation

Sedimentary units (1) and (2) can be interpreted as proglacial outwash capped by (3) a debris flow deposit probably deposited shortly after the outwash. The fault sets and their uniform strike and dip indicate that deformation was induced by compression from the south-west (cf. Chapters 4.3.3, 4.3.4). The deformational evidence is compatible with proglacial deformation caused by lateral compression in which the more competent sand and fine sediment units (Sh and Fl) responded by brittle deformation while the diamict units (Dcs) were folded (cf. Aber *et al.,* 1989; Hart and Boulton, 1991; Benn and Clapperton, 2000; Lukas and Merritt, 2004). This is interpreted as evidence for a readvance after deposition of a valley train the presence of which is indicated by outwash terraces in the surroundings. Similar observations of deformed ("shunted") outwash sediments in "hummocky moraines" were reported by Golledge and Hubbard (2005) from the SW Highlands.

4.3.1.7 Synthesis of sedimentary processes

Geomorphologically, moraine mounds and ridges are organised in discrete chains in all areas reported here. They are thus very similar to those observed throughout large areas of the Scottish Highlands (e.g. Benn, 1992a; Bennett and Boulton, 1993a; Lukas, 2003; Golledge and Hubbard, 2005).

Sedimentologically, six of the seven examples reported here represent terrestrial ice-contact fans (Table 4.1). Exposures through 52 moraines in the study area have all yielded similar lithofacies associations and thus confirm that this is characteristic of the moraines in the NW Highlands. In addition, the most detailed studies on Scottish "hummocky moraine" to date (Benn, 1990, 1992a; Benn *et al.,* 1992) recognise ice-contact fan deposits in the majority of exposures in moraines, and similar evidence elsewhere (Bennett and Boulton, 1993a; Lukas, 2003; Mitchell and Lukas, 2004; Golledge and Hubbard, 2005) indicates that most of the sediment in "hummocky moraine" was originally deposited in terrestrial ice-contact fans. The deformed proglacial outwash sediments described above, however, indicate that there are exceptions, and local conditions such as the existence of ice-dammed lakes might add further variants yet to be discovered.

The frequent occurrence of deformation structures of varying intensities within the fans shows that their formation involves at least a two-staged process (Fig. 4.20): (a) the formation of the fan at a temporarily stationary ice margin followed by retreat and formation of a proximal rectilinear ice-contact slope (b). During a readvance (c), the fan units can be proglacially deformed and, when followed by a phase of retreat, a proximal rectilinear ice-contact slope is formed (d). If the readvance is more sustained than at stage (d), partial (e) or complete overriding (f) can take place, resulting in subglacial shearing of units. Complete overriding was also found to lead to the reduction of relief and glaciotectonisation. The formation of Scottish ice-marginal "hummocky moraines" can thus best be viewed as a continuum of undeformed to completely overridden and glaciotectonised terrestrial ice-contact fans with a variety of scales in between, allowing grouping into seven distinct classes at present (Table 4.1).

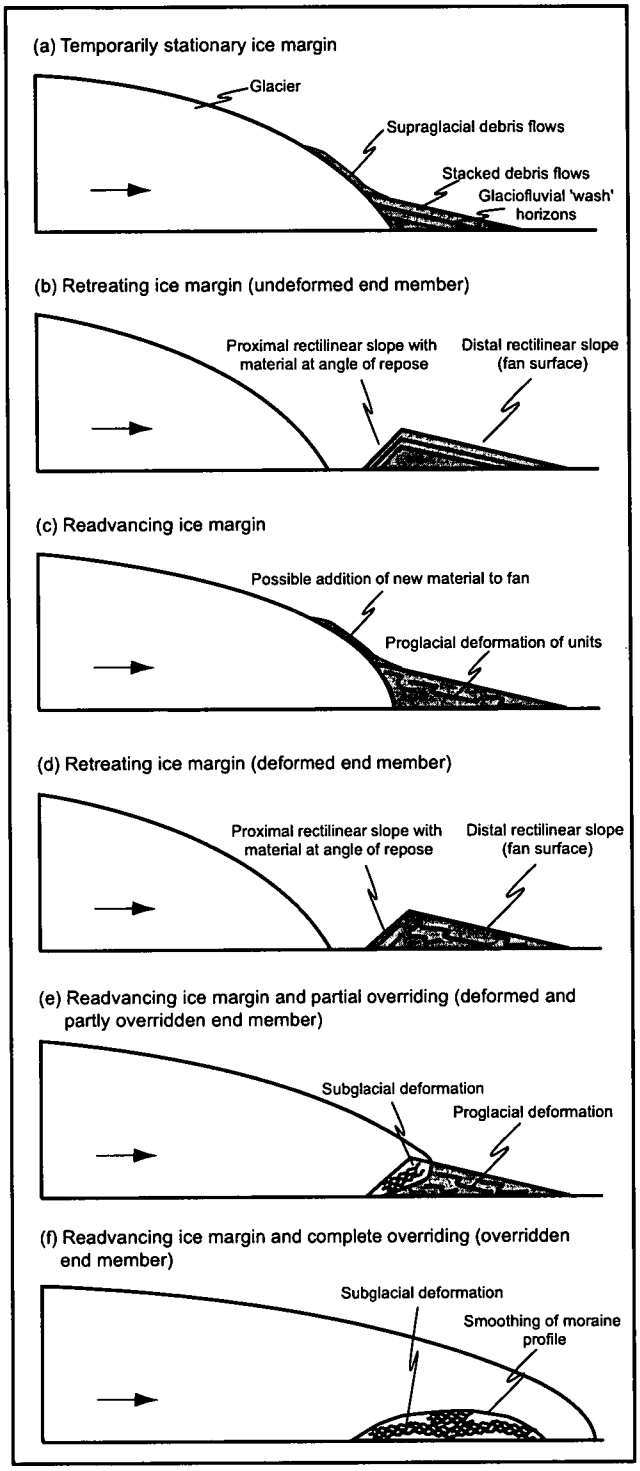


Fig. 4.20 Schematic representation of sequence of events involved in the formation of terrestrial ice-contact fans as described in this paper (cf. Table 4.1). (a) Fan formation along a temporarily stationary ice margin by stacking of supraglacial debris flows and glaciofluvial sediments. (b) Formation of the rectilinear ice-contact face where material is at the angle of repose as a result of partial collapse following withdrawal of ice support. (c) Short-lived readvance of the ice margin (with a possible annual signature in some areas) causing widespread deformation within the fan and occasionally addition of new material. (d) Formation of a new ice-contact face and abandonment of the fan. (e) Partial overriding of the proximal part of a moraine leading to partial glaciotectonisation. (f) Larger-scale overriding leading to smoothing and alteration of the original moraine asymmetry and complete glaciotectonisation. This sequence does not represent an evolutionary sequence, and stages (b), (d), (e) and (f) are discrete end members of a continuum of deformation.

Table 4.1 Classification of Younger Dryas "hummocky moraines" in the NW Highlands and their implications for palaeo-glacier dynamics based on sedimentological composition, deformation structures and style.

Sedimentological interpretation	Deformation structures	Style of deformation	Implications for palaeo-glacier dynamics	Example[1]
Terrestrial ice-contact fan	Absent	None	Formation at temporarily stationary ice margin during overall retreat	SLK 1 (4.20a, b)
Deformed terrestrial ice-contact fan	Small-scale folds, low-angle reverse faults	Proglacial lateral compression	Short-lived readvance after formation	SLK 13 (4.20a-d)
Heavily deformed terrestrial ice-contact fan	Overturned folds, low- and high-angle reverse faults, decollèment	Proglacial lateral compression	Prolonged readvance after formation	SLK 12 (4.20a-d)
Dislocated terrestrial ice-contact fan	Oversteepened units (dip 50-70°), high-angle reverse and normal faults, small-scale folds	Proglacial lateral compression	Prolonged readvance after formation	SLK 3 (4.20a-d)
Overridden terrestrial ice-contact fans	Long-wavelength, low-amplitude folding, boudins, symmetric drape around boulders, tectonic laminae, streaked-out lenses of sediment	Proglacial compression, subglacial simple shear, non-penetrative glaciotectonism	Prolonged readvance possibly long after formation, overriding of several moraines in one event	SLK 17 (4.20a-f)
Deformed and partly overridden terrestrial ice-contact fan	Hybrid of the above, deformation styles spatially separated	Proglacial lateral compression and subglacial simple shear	Partly overridden, partly pushed	SLK 4 (4.20a-e)
Deformed outwash sediments	Low- and high-angle reverse and normal faults, small-scale folds	Proglacial lateral compression	Prolonged readvance	SLK 5

[1] Section ID code referred to in the text (steps of evolution shown in Fig. 4.20)

The wealth of information provided by 52 logged exposures (Appendix 1) allows inferences about relative frequencies (and thus importance) of the individual process-combinations responsible for the formation of individual moraines to be made. Applying the criteria provided in Table 4.1 to classify all of these sections results in the frequency distribution shown in Fig. 4.21. This distribution highlights the following key elements: (a) undeformed terrestrial ice-contact fans that were produced at a temporarily stationary ice margin are rare (making up only 5.8%) compared to those containing clear evidence of deformation of varying intensities that make up 55.8 % of all moraines studied. (b) Only 3.8% (2 moraines) do not consist of supraglacial debris flows and intercalated fluvial horizons; these represent outwash fans that also indicate a two-staged process of (1) fan formation and (2) deformation and moraine formation during a readvance. However, (c) the fact that in the majority of small surface exposures units showing deformation structures (e.g. Sh, Sm, Sl, Fl) could not be found illustrates the problem of using surface shallow exposures alone to infer the mode of formation.

Although the nature of ice-marginal sedimentation, i.e. the presence of units compatible with terrestrial ice-contact fan formation, can be established with confidence, subtle and important information such as deformation structures and a possible lateral change in deformation style can only be recognised in larger exposures that make up two thirds (67.3%) of the moraines studied. This implies that studies relying on shallow surface exposures alone cannot be regarded as providing reliable information about genetic processes in ice-marginal (moraine-forming) environments.

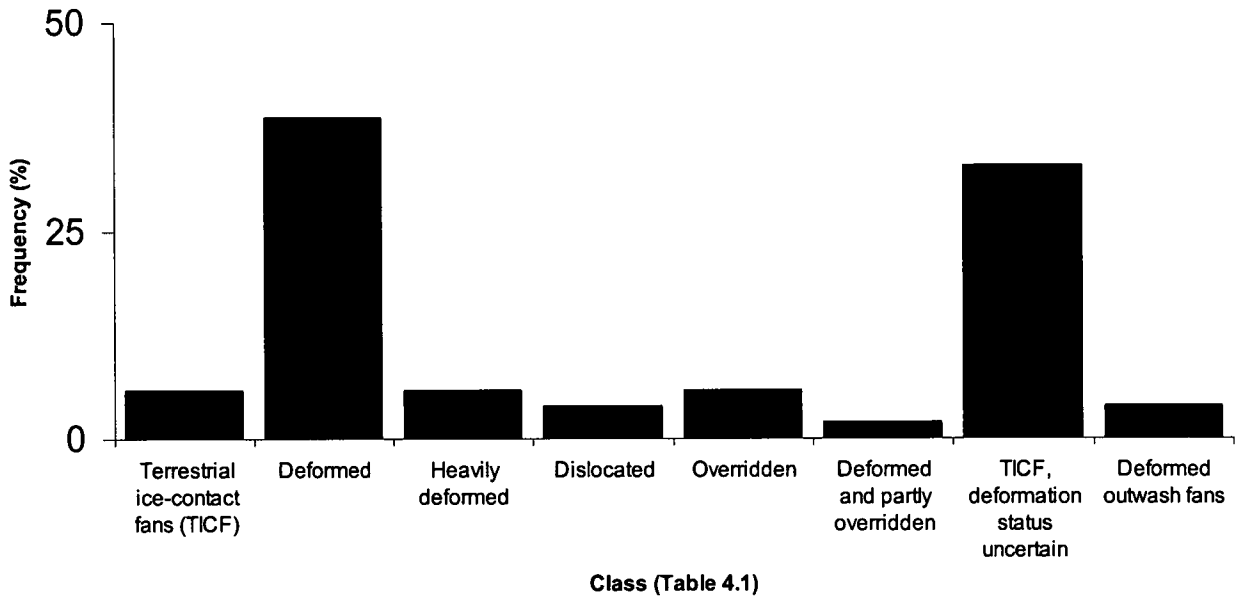


Fig. 4.21 Frequency distribution of the types of moraines found in the study area. The classification follows that in Table 4.1.

4.3.2 Transport paths and significance for glacier dynamics

4.3.2.1 Introduction

Debris in glaciers can be entrained by a variety of mechanisms from a number of sources, e.g. by rockfall from surrounding valley walls or by subglacial plucking and entrainment; both processes can incorporate bedrock and/or unconsolidated sediments. Once entrained, such material can then be transported either "actively" along the glacier bed or "passively" englacially or supraglacially (Boulton, 1978; Ballantyne, 1982; Benn and Evans, 1998; Benn and Owen, 2002; Hambrey and Ehrmann, 2004). Past research has established a strong correlation between different transport paths and clast shapes, and it has been demonstrated that the measurement of clast properties enables an effective reconstruction of the palaeo-transport paths (Ballantyne, 1982; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Evans, 1999; Lukas *et al.*, 2005a). Material may ultimately be deposited in a number of different ways (e.g. Benn, 1992a; Benn and Evans, 1998), two of which are of importance

in the Scottish context: (a) Release in a subglacial setting may result in the accumulation of glacigenic or glaciofluvial sediments (e.g. till, conduit fill) whereas (b) ice-marginal sedimentation may lead to the building of a landform (e.g. moraine, outwash fan, kame). Determining the source of glacigenic material found in sediments and landforms can yield valuable information about the past glacier system.

This section presents clast shape measurements obtained from ice-marginal moraines in order to reconstruct transport paths and to supplement the palaeo-environmental evidence provided by sedimentology and geomorphology. The methods used in this approach are presented in Chapter 2.1.2. Pilot studies showed that smaller-scale surface characteristics of importance such as striae were not preserved on Moine lithologies, and, consequently, such features were not recorded for individual clasts in measurements; however, clear features such as stoss-and-lee-forms ("bullet-shaped clasts") and facets in the clasts were recorded regularly during section logging.

4.3.2.2 Sampling strategy

As illustrated in Chapter 2.1.2, control samples are required to guide the interpretation of clast samples of unknown origin, so that transport paths can effectively be reconstructed. Control samples of fluvial sediments were obtained from river bars as these were thought to represent the closest analogue to clasts transported and deposited in glaciofluvial systems. This is mainly because large diurnal discharge variations cause particularly high flow velocities leading to larger grain sizes being transported as bedload compared to rivers in many non-glaciated catchments (e.g. Gurnell and Clark, 1987; Benn and Evans, 1998; Knighton, 1998). Modern scree was sampled at steep slope sections above the Younger Dryas glacier limits as a surrogate for supraglacial sources following the approach of Benn (1990, 1992a), Bennett *et al.* (1997) and Spedding and Evans (2002). Subglacial control samples are more difficult to obtain with certainty. Benn (1990, 1992a) obtained fragments broken from the leeside of roches moutonnées and clasts from exposures of subglacial till as subglacial control samples. The former approach was found to be not applicable in the NW Highlands due to the absence of joints near roches moutonnée lee faces and extensive peat cover; subglacial control clasts were thus taken from four different exposures of subglacial till.

Clasts were sampled from 20 lithofacies units representative of those encountered in ice-marginal moraines in the NW Highlands (cf. Chapter 4.3.1; Appendix 1). In addition, clasts perched on top of roches moutonnée surfaces were sampled. Details of the location of the samples and brief facies descriptions are listed in Table 4.2.

Table 4.2 Location and characteristics of samples from moraines taken for clast shape and roundness measurements; characteristics employ the lithofacies code introduced in Chapter 2.1.1.

Sample ID	Grid reference (NC)	Lithofacies characteristics	Figure number
SLK 1-1	39333 34445	Sandy-silty Dcs	4.23a
SLK 1-2	39333 34445	Very fine-sandy Dcs	4.23b
SLK 2-1	40616 35646	Fine to medium-sandy Dcs	4.23c
SLK 3-1	43850 38270	Fine-sandy Dcs	4.23d
SLK 3-2	43850 38270	Very fine-sandy to silty Dcs	4.23e
SLK 3-3	43850 38270	Fine-sandy Dcm	4.23f
SLK 4-1	44220 38710	Sandy Dcs	4.23g
SLK 4-2	44220 38710	Fine-sandy Dcs	4.23h
SLK 4-3	44220 38710	Silty to very-fine sandy Dms	4.23i
SLK 4-4	44220 38710	Fine-sandy Dcs	4.23j
SLK 5-1	44317 39171	Fine-sandy Dcs	4.23k
SLK 5-2	44317 39171	Sandy GRh	4.23l
SLK 12-1	35820 34650	Fine-sandy to silty Dcs	4.23m
SLK 12-2	35820 34650	Sandy-silty Dms	4.23n
SLK 12-3	35820 34650	Fine to medium-sandy Dcs	4.23o
SLK 13-1	35750 34770	Sandy GRm	4.23p
SLK 13-2	35750 34770	Silty to fine-sandy Dcs	4.23q
SLK 14-1	38383 41438	Sandy Dcs	4.23r
SLK 17-1	39180 24950	Fine-sandy Dcs	4.23s
SLK 50-1	40821 44673	Silty to fine-sandy Dms	4.23t

4.3.2.3 Results

Control samples

Fig. 4.22 shows the shape and roundness characteristics of subglacial (Fig. 4.22a-d), fluvial (Fig. 4.22e-f) and scree control samples (Fig. 4.22g-i). Subglacial control samples all have RA-indices of zero with a dominant subangular (SA) component. Only one sample (Fig. 4.22a) shows a peak in the subrounded (SR) class whereas rounded (R) clasts are generally absent. Clast shapes are dominantly blocky, i.e. spherical or equiaxial, as indicated by C_{40}-indices of ≤ 20 (Fig. 4.22a-d). The absence of very angular (VA) and angular (A) clasts in these control samples is interpreted as evidence for efficient edge-rounding during glacial transport and corresponds well to earlier results obtained elsewhere (Ballantyne, 1982; Benn, 1990, 1992a; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Benn and Evans, 1998; Evans, 1999, and references therein). Likewise, the predominantly blocky shape of clasts was found in aforementioned studies to be characteristic of subglacially transported material.

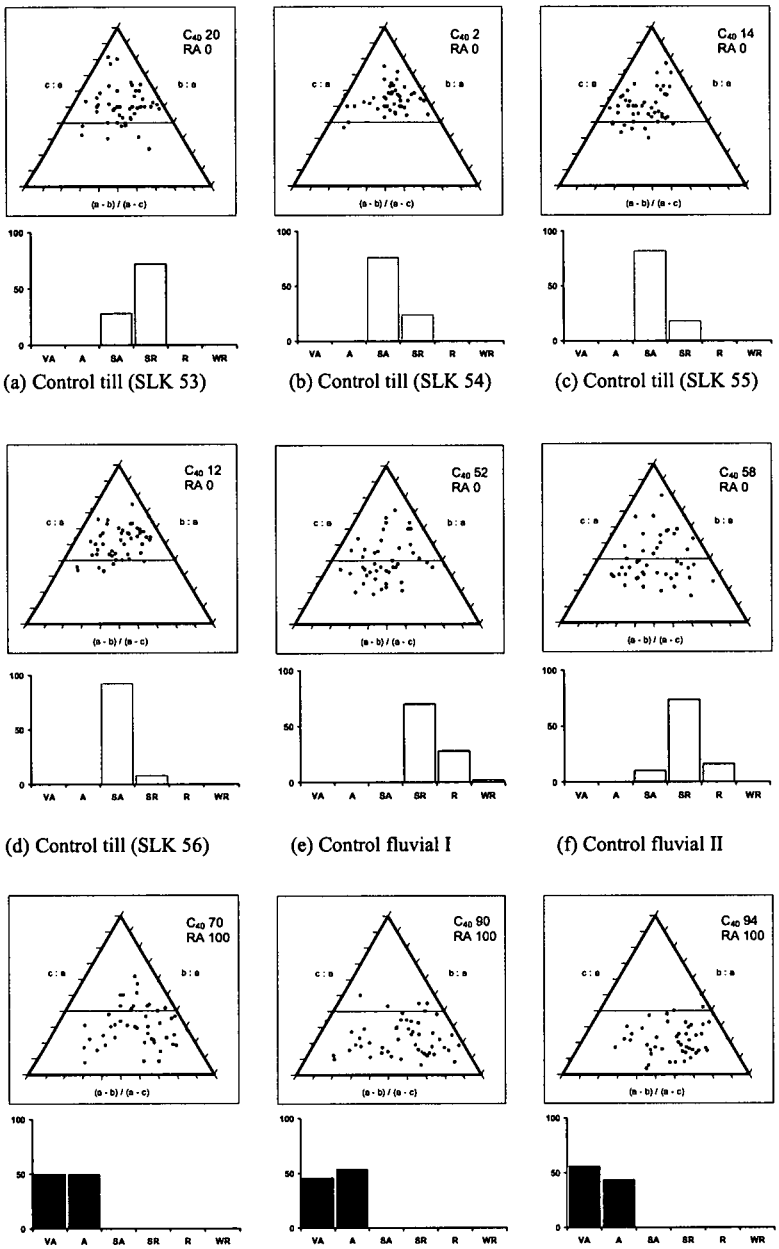
Contrasting with these subangular to subrounded and high-sphericity clasts, those from fluvial sources show dominant proportions of subrounded and rounded clasts with only few subangular ones present in one of the samples (Fig. 4.22f); these characteristics are manifest in extremely low RA indices (0 % for both samples) and interpreted as a result of grinding and rolling of clasts in bedload

transport during flood events (Knighton, 1998). The shapes of fluvially transported clasts show an almost equal mixture of prolate and oblate on the one hand and blocky clasts on the other (Fig. 4.22e-f) with C_{40}-indices between 50 and 60.

Conversely, the supraglacial control samples, represented by scree, consistently show an absence of edge-rounded (subangular, SA) clasts indicated by their very high RA-indices (100% for each of the samples; Fig. 4.22g-i). Likewise, clasts from these locations show C_{40}-indices of ≥ 70, indicating that clast populations are mixtures of prolate (rod-shaped or elongated) and oblate (disc-shaped or platy) shapes. These shapes are very similar to supraglacially-derived clasts observed elsewhere, and their form is inferred to reflect breakage along former joints enlarged by frost-weathering (Benn, 1990, 1992a; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Evans, 1999; Spedding and Evans, 2002; Lukas *et al.*, 2005a).

Moraine samples

20 lithofacies units from ten ice-marginal moraines were analysed for clast shape and roundness (Fig. 4.23a-t). These samples can be divided into clasts recovered from clast- and matrix-supported, stratified diamict units, interpreted as supraglacial debris flow units, and those from sorted sediments interpreted as fluvial horizons (cf. Chapter 4.3.1). Most samples were taken from diamict units as these make up the bulk of sediments exposed in ice-marginal "hummocky moraines" in the study area (Chapter 4.3.1).

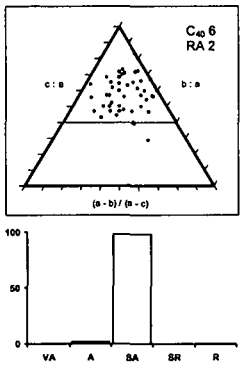


(a) Control till (SLK 53) (b) Control till (SLK 54) (c) Control till (SLK 55)

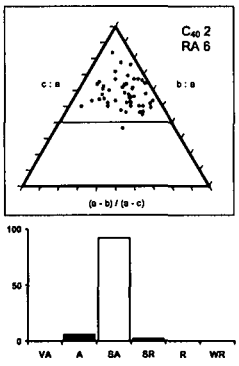
(d) Control till (SLK 56) (e) Control fluvial I (f) Control fluvial II

(g) Control scree I (h) Control scree II (i) Control scree III

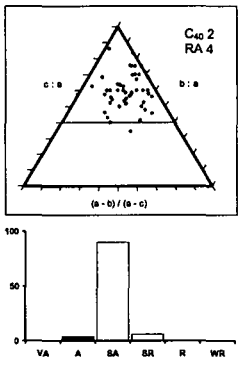
Fig. 4.22 Shape diagrams and frequency distribution of roundness classes for control samples from the NW Highlands, Scotland. A full explanation is given in the text.



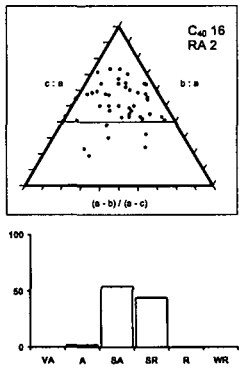
(a) SLK1-1 (Dcs)



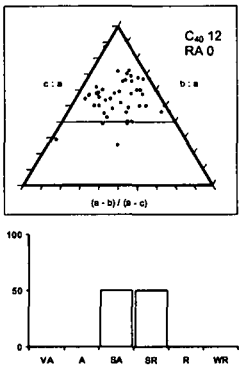
(b) SLK1-2 (Dcs)



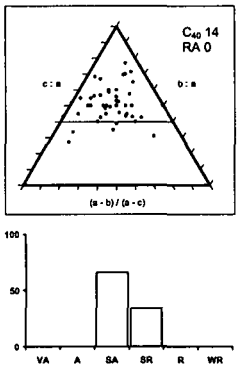
(c) SLK2-1 (Dcs)



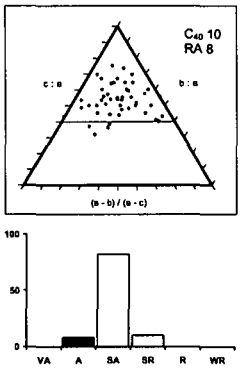
(d) SLK3-1 (Dcs)



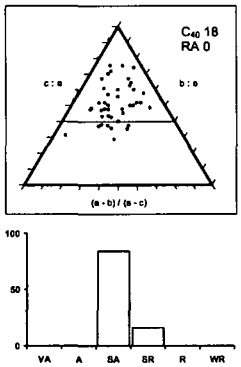
(e) SLK3-2 (Dcs)



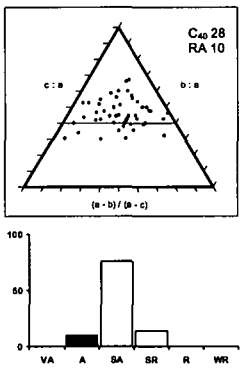
(f) SLK3-3 (Dcm)



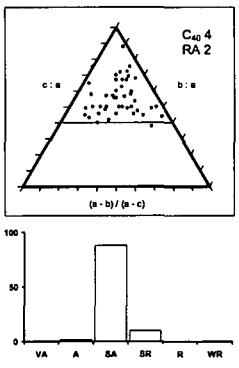
(g) SLK4-1 (Dcs)



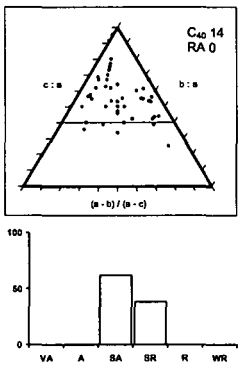
(h) SLK4-2 (Dcs)



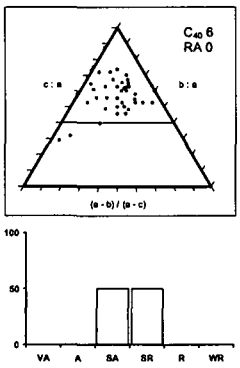
(i) SLK4-3 (Dms)



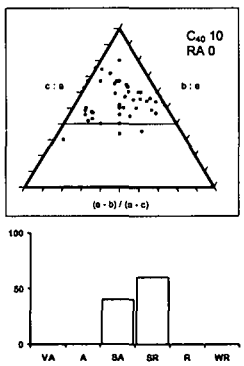
(j) SLK4-4 (Dcs)



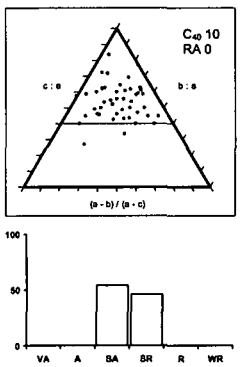
(k) SLK5-1 (Dcs)



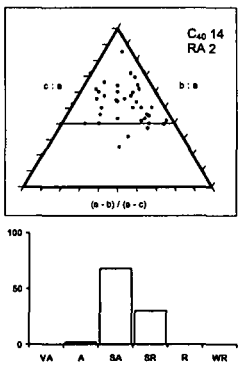
(l) SLK5-2 (GRh)



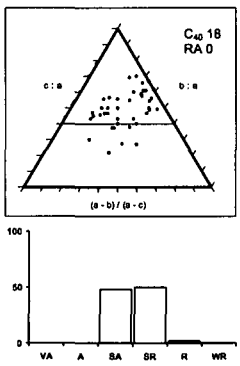
(m) SLK12-1 (Dcs)



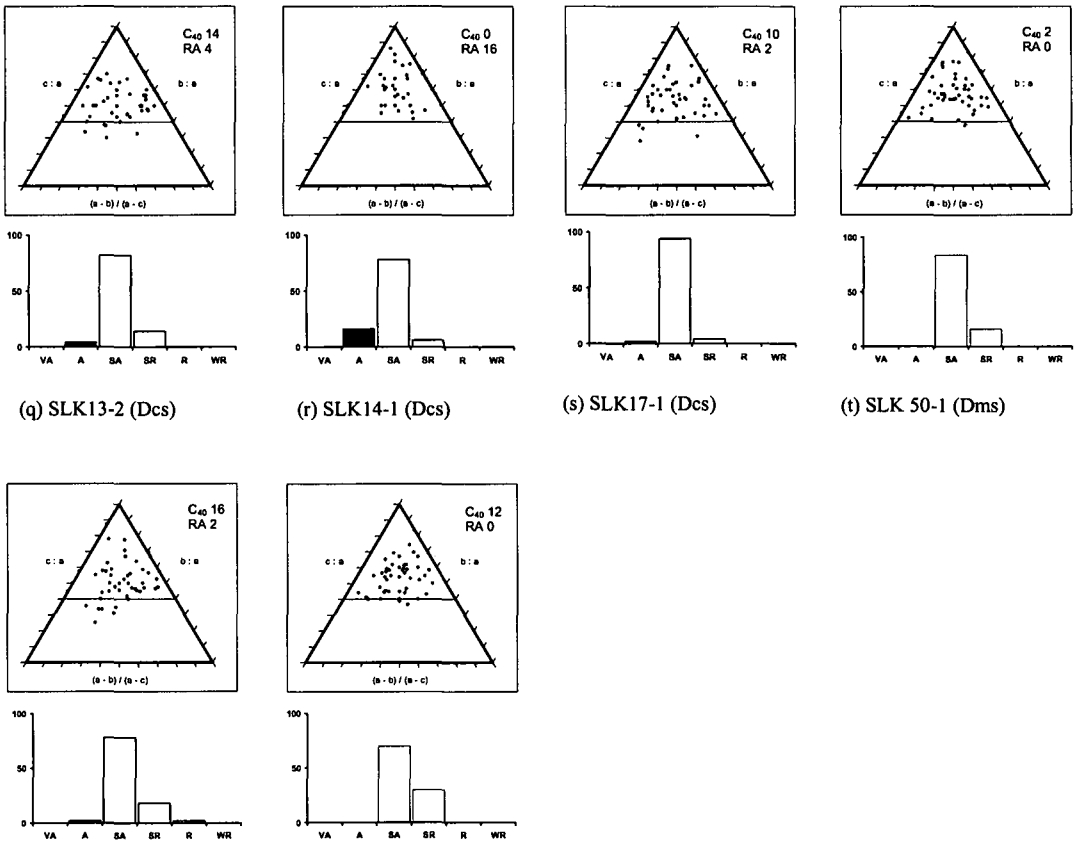
(n) SLK12-2 (Dms)



(o) SLK12-3 (Dcs)



(p) SLK13-1 (GRm)



(q) SLK13-2 (Dcs) (r) SLK14-1 (Dcs) (s) SLK17-1 (Dcs) (t) SLK 50-1 (Dms)

(u) Clasts roches moutonnées I (v) Clasts roches moutonées II

Fig. 4.23 Clast shape diagrams and frequency distribution of roundness classes. (a)-(t) are from samples obtained from ice-marginal moraines (cf. Table 4.2), (u) and (v) are clasts picked from the surface of roches moutonnées. For explanation see text.

The shape of these clasts can be described as equiaxial or blocky throughout since C_{40}-indices are consistently low to very low, between 0 and 28 (Fig. 4.23a-t). At the same time, the majority of the samples only contain up to two angular clasts which is reflected in very low RA-indices between 0 and 16 (Fig. 4.23a-t). Numerous clasts show lunar-shaped facets on their surfaces and display a bullet-shape with one side smooth and the other rugged.

Surficial samples

Two samples were taken from the surface of roches moutonnées to characterise their mode of transport. Such data might enable a reconstruction of the dominant clast category contained within the glacier as perched clasts would generally have been deposited by meltout from material contained within the ice or on its surface. The two samples (Fig. 4.23u-v) show RA-indices of 2 and 0, and a very low proportion of prolate and oblate clasts, evident from C_{40}-indices of 16 and 12, respectively (Fig. 4.23u-v).

4.3.2.4 Reconstruction of glacial transport paths and clast origin

The results presented above are most effectively compared and interpreted using a co-variance plot of roundness (RA) plotted against shape (C_{40}) following Benn and Ballantyne (1994). Fig. 4.24 shows three distinct clusters of control samples, the supraglacial clasts with high RA indices and low sphericity values, the fluvial clasts with a low RA index and intermediate sphericity and the subglacial control samples with low RA values and a very high sphericity. Most importantly, these clusters are distinctly different as no overlap exists. The mixture of shapes and the dominantly subrounded shape of fluvially transported clasts is interpreted be a result of mixing of subglacially and supraglacially sourced material that was rounded in bedload transport; these mixtures and the better rounding than the subglacial clasts set them apart.

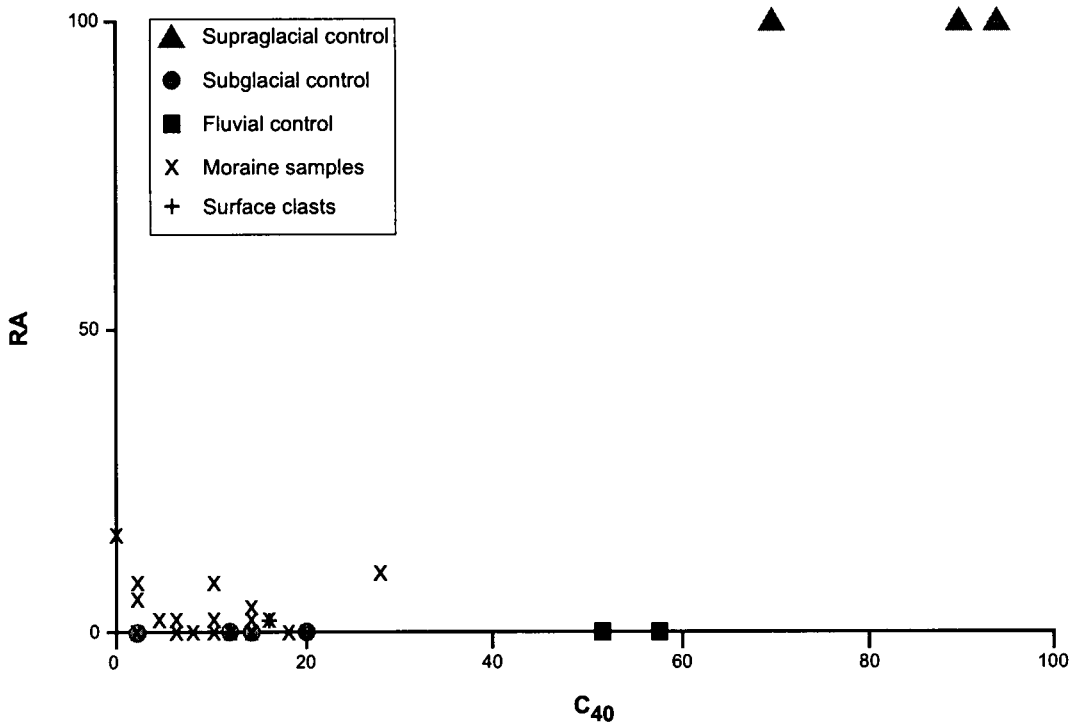


Fig. 4.24 Co-variance plot of roundness (RA) and shape (C_{40}) indices following the approach introduced by Benn and Ballantyne (1994).

The clasts obtained from ice-marginal moraines show some variation, but are consistently characterised by very low to low RA values and a high sphericity (Fig. 4.24). All these data plot very close to, and overlap with, the subglacial control samples strongly suggesting a dominantly subglacial mode of transport for the clasts found in ice-marginal moraines (Fig. 4.24). Furthermore, the frequent occurrence of facets on the surface of clasts taken from moraines and their bullet-shape is characteristic of subglacially transported sediments (e.g. Krüger, 1979, 1984; Benn, 1994, 1995, 2004; Benn and Evans, 1998). The two samples taken from the surface of roches moutonnées plot in the

same area as the subglacial control samples and those taken from moraines (Fig. 4.24), also indicating that they were subglacially-transported and probably released by meltout from debris-rich basal ice.

Together, the data strongly suggest that the clasts were (a) dominantly subglacially transported prior to deposition and that (b) this was the case for all the diamicts and sorted sediments making up recessional moraines throughout the study area (Fig. 4.24, Table 4.2). It is very likely that this dominance of a subglacial transport path reflects a subglacial origin, i.e. the clasts were entrained by plucking of bedrock. Evidence in support of this argument comes from a number of observations: (a) roches moutonnées are abundant in the study area whereas (b) only few locations of steep slopes overlooking the glacier surface and capable of producing supraglacial debris were found (cf. Appendix 2). Indeed, the majority of such slopes would have become exposed only during the later stages of deglaciation when a large number of moraines would have been formed already. Another indication for a dominant subglacial debris source and transport path and a very limited supraglacial component in the Younger Dryas glacial transport system in the NW Highlands comes from (c) the two samples obtained from the surface of roches moutonnées that were also subglacially transported. In addition, (d) blocky shapes similar to the control and moraine samples reported here have been obtained from lee faces of roches moutonnées elsewhere (Benn, 1990, 1992a, and references therein) and are generally associated with subglacially-derived and -transported material (e.g. Ballantyne, 1982; Benn and Ballantyne, 1993, 1994; Bennett *et al.*, 1997; Evans, 1999; Spedding and Evans, 2002).

Boulton and Eyles (1979) and Eyles (1983b) suggested that debris in glaciated valley landsystems was predominantly derived from supraglacial positions and that supraglacial transport was dominant with a less important subglacial component. The findings of the recent study contradict this idea and indicate that the assumption of a single glaciated valley landsystem that accounts for every setting is unrealistic (cf. Benn and Evans, 1998; Spedding and Evans, 2002; Benn *et al.*, 2003). It would appear that in the Scottish case such a landsystem dominated by supraglacial transport does not reflect the currently available data.

Another possible sediment source is the reworking of older material as speculated by Ballantyne (2002b). This possibility is not directly testable, although it can be noted that till sheets or blankets of glacigenic sediments both inside and outside the Younger Dryas maximum extent are not very thick, and signs of postdepositional reworking are very limited throughout the NW Highlands compared to other sites in the Scottish Highlands where extensive reworking of debris has been reported on hillslopes inside and outside the Younger Dryas maximum (e.g. Curry, 1999, 2000; Curry and Ballantyne, 1999; Ballantyne, 2002b). Thus, the rates of paraglacial adjustment for the Scottish Highlands which Ballantyne (2002b) compares to Norway may not be applicable to the far NW Highlands where bedrock control is stronger compared to other areas. Another indirect way of testing a large paraglacial debris source is by assuming a constant subglacial debris source. If there was a

dominant paraglacial component, then the outermost moraines should be the largest with successive moraines decreasing in size as a result of progressive debris removal from "paraglacial stores" in the course of a glacial advance (Ballantyne, 2002b: 1994ff.). This type of pattern does not occur in the study area, and moraine size appears to vary randomly (Chapter 4.2.1). This indicates that other factors, such as the delivery of freshly-produced material to the glacier margins and the duration of stillstands/readvances that formed a moraine during overall retreat, played a significant role.

All the evidence presented here appears to imply a dynamic environment in which plucking and the constant production of fresh material appears to have dominated the glacial landsystem.

4.4 Interpretation and significance of "hummocky moraine"

4.4.1 Genetic processes

Mapping of the location and planform patterns of areas of "hummocky moraine" in the far NW Highlands (Appendix 2, Chapter 4.2.1) reveals that all areas of "hummocky moraine" display very similar morphological characteristics: (a) they contain a large number of asymmetrical moraine ridges and mounds, (b) these trend obliquely across the slopes in a downvalley direction and are aligned in individual chains, (c) the connection of crestlines yields continuous, arcuate lines throughout, (d) crestline bifurcations are frequent and (e) these moraines are closely associated with meltwater channels. Although these characteristics have previously been adduced as evidence of incremental retreat (e.g. Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003), an interpretation based purely on surface morphology and planform patterns retains a large amount of uncertainty.

Sedimentological logging of 52 exposures in such moraines (Appendix 1) has revealed that the majority of moraines contain alternating debris flows and glaciofluvial units that were deposited in an ice-marginal position during temporary stillstands (Chapter 4.3.1). Deformation structures indicate that many moraines were affected postdepositionally by lateral compression, which is interpreted as evidence for readvances into the moraines at a later stage. Therefore, the sedimentology demonstrates that the moraines did indeed form ice-marginally and were not derived from englacially thrusted slabs. This implies that one moraine arc, reconstructed by connecting the crestlines, represents one former ice-marginal position thus confirming the geomorphological interpretation of "hummocky moraines" as recessional moraines (e.g. Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003). Furthermore, the sedimentology strongly suggests that the retreat mode was oscillatory and that readvances were the norm rather than an exception.

The fragmented nature of the moraines that gives them their hummocky appearance can be explained by a number of factors. First, glacial meltwater that emerges at the glacier portal either

prevented moraines from being formed there or postdepositionally breached formerly deposited moraines and eroded parts of them (cf. Bickerton and Matthews, 1993). Second, lateral meltwater channels that sometimes cross several moraine arcs indicate that formerly continuous moraines might have been dissected during retreat by laterally emerging meltwater. Third, on hillsides snowmelt can have contributed to dissection and gully initiation (cf. Lukas, 2003, 2004b). However, a fourth possibility, namely that moraine arcs were not continuous at the time of deposition but were deposited in discrete positions along the ice margins has to be taken into account as well.

4.4.2 Implications for palaeo-glacier dynamics in the NW Highlands

The results presented above demonstrate an ice-marginal mode of formation of "hummocky moraines" during oscillatory retreat in the study area. The dense spacing and large number of moraines implies that glaciers were close to equilibrium throughout retreat during the second half of the Younger Dryas, thus sedimentologically supporting Ballantyne's (2002a) suggestion based on morphological observations on the Isle of Mull.

As the moraines represent truly ice-marginal landforms, connecting their crestlines following the approach of earlier workers yields the position of palaeo-ice margins (Fig. 4.25). Since they record successive ice-marginal positions during retreat, these lines can be used to reconstruct the pattern and direction of ice retreat across the study area (Benn, 1990; Bennett, 1990; Benn *et al.*, 1992; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003). This pattern reveals that the individual glaciers making up the transection glacier complex retreated towards the watershed. Furthermore, and perhaps more importantly, most glaciers irrespective of valley aspect remained close to equilibrium throughout retreat over most parts along the valley axes (Fig. 4.25). Only the shorter glaciers, such as those sourced in the corries east of Foinaven and the corrie glacier complex on Arkle, show a different pattern with only a few retreat stages being recognised inside from the maximum position. The areas near the source areas of these steeper glaciers are almost exclusively devoid of moraines (Fig. 4.25). This suggests that basin morphology and glacier size had an effect on the response of glaciers to Younger Dryas climate warming. Similar findings have been reported by Benn (1990) and Benn *et al.* (1992) from the Isle of Skye, supporting these observations. Collectively, it would appear, therefore, that ice cap size had an influence on glacier response times, possibly because larger ice masses can exert some control on local climatic conditions and respond less readily (Sutherland, 1984b; Benn *et al.*, 1992; Nesje *et al.*, 1995; Winkler, 1996b; Benn and Evans, 1998).

In addition to the almost continuous nature of active, oscillatory retreat, there is no evidence of chaotic moraine assemblages, strongly suggesting that stagnation was entirely absent. This noticeable absence of evidence agrees well with the findings of Benn (1990, 1992a) who found that "chaotic" moraines were only locally important. On the Isle of Skye, they were restricted to areas below former ice-sheds in the final stages of deglaciation where accumulation areas were lost and thus

only occurred under specific topographic conditions. Similar observations have also been made in the Central Grampian Highlands (Lukas, 2003). The continuous occurrence of "hummocky" recessional moraines from Younger Dryas maximum to the source areas in most corries within the present study area differs from the observations of Benn *et al.* (1992). These authors (p. 142) observed that "recessional moraines are absent from the upper parts of glaciated basins throughout the Scottish Highlands". They interpreted this evidence in terms of a two-staged retreat phase with rapid, uninterrupted retreat and localised stagnation following an initial phase of oscillatory retreat. The findings from the far NW Highlands imply that conditions differed from those in that most glaciers remained closely coupled with climate retreating into their source areas in an oscillatory fashion. On the whole, this continuous retreat pattern is more similar to that observed by Bennett and Boulton (1993a, b) for the northern part of the Western Highland Ice cap than the restricted, initial phase of oscillatory retreat found on Skye. This disparity indicates that local climatic conditions might have varied across Scotland during the Younger Dryas.

Field observations in the present study area have shown that some moraines directly merge at their bases while others are frequently less than 20 m apart. Similar observations from detailed mapping of moraine spacing in selected areas have been reported by Benn (1990) from the Isle of Skye. It is hypothesised here that the spacing of moraine arcs (palaeo-ice fronts) can be utilised to extract information about the relative timing of the response of individual outlet glaciers to climate warming during the second half of the Younger Dryas. This is based (a) on the fact that these moraines have been demonstrated above to have formed ice-marginally and (b) from the notion that the spacing of moraine sequences at modern glaciers can often be correlated with climatic records (e.g. Krüger, 1995; Bradwell, 2004). The spacing between palaeo-ice fronts was thus recorded to test in how far the timing of retreat differed for individual outlet glaciers. This was done by marking with a vertical bar the point where each reconstructed arc crossed the valley axis. Tracking successive palaeo-ice fronts along valley axes, which correspond to glacier flowlines during retreat, yields "barcode patterns" that allow direct comparison of retreat patterns in different valleys. Fig. 4.26 shows such barcodes for selected valleys across the study area. A selection has become necessary for visualisation purposes since the possible retreat routes in the study area total 32.

Fig. 4.25 Location of palaeo-ice fronts reconstructed from "hummocky moraine" sequences for glaciers in the far NW Highlands. Ice flow directions are omitted for clarity. Letters denote locations of the transects displayed as barcodes in Fig. 4.26. Inset map shows the exact transect lines.

Fig. 4.26 Barcode patterns depicting the retreat patterns of selected Younger Dryas glaciers in the far NW Highlands. The barcodes are aligned so that the Younger Dryas maximum is located on the left and the source areas on the right. Valley aspect is indicated above each transect. The scale for all transects is the same. Location of transect lines is shown in Fig. 4.25.

It is assumed here that the glaciers reached their maximum sometime around the middle of the Younger Dryas, in agreement with recent findings from numerical modelling (Hubbard, 1999; Golledge and Hubbard, 2005) and palaeoclimatic studies (Brooks and Birks, 2000; Ballantyne, 2002a; Benn and Ballantyne, 2005). These imply an onset of incremental glacier retreat in the second half of the Younger Dryas, around 12100 cal a BP (see Hubbard, 1999; Brooks and Birks, 2000; Golledge and Hubbard, 2005). In order to allow direct comparisons of individual retreat routes, it is assumed that the glaciers reached their maximum synchronously.

Examination of the individual glacier barcodes suggests that the response to climate change was different since the spacing differs markedly even for two adjacent valleys with the same aspect (Fig. 4.26b, c). The observed spacings and inferred asynchroneity of retreat confirm the notion that external control factors such as basin morphology, aspect, snow-bearing winds and glacier size have a large influence on glacier response to climate change (e.g. Sutherland, 1984b; Benn, 1990; Bickerton and Matthews, 1993; Winkler, 1996b; Benn and Evans, 1998; Benn and Lehmkuhl, 2000). Only in an ideal situation where such factors were negligible could moraine spacings be expected to be identical for any pair of retreat routes.

In addition to the spacing patterns, the number of moraine ridges along each transect was counted to make informed statements about the average distance that the ice margin retreated between the formation of two consecutive moraines. Assuming that retreat took place in the second half of the Younger Dryas and that glaciers did not persist into the early Holocene (cf. Ballantyne, 1991, 2002a; Benn *et al.*, 1992; Benn, 1997a; Gray, 1997; Clapperton, 1997), this leaves ca. 600 years for the glaciers to retreat from the Younger Dryas maximum back into the source areas. In order to account for survival of active ice into the early Holocene, two additional hypothetical time spans were used, namely 800 and 1000 years. Dividing this time span by the number of moraines along individual flowlines yields an approximation to the time passed between the formation of two successive chains of moraines (Table 4.3). From this data an approximate retreat velocity can then be calculated (Table 4.4).

Table 4.3 Approximate number of moraines formed along selected flowlines with different aspects and the calculated mean time span between the formation of two successive moraine ridges.

Transect [1]	Length of transect (m)	Number of moraines	Average spacing (m)	t=600a [2]	t=800a [3]	t=1000a [4]
A-A'	9400	197	48.0	3.0	4.0	5.1
B-B'	18000	188	95.7	3.2	4.3	5.3
C-C'	17200	210	82.0	2.9	3.8	4.8
D-D'	10500	109	96.3	5.5	7.3	9.2
E-E'	6250	85	73.5	7.1	9.4	11.8
F-F'	4200	53	79.2	11.3	15.1	18.9
G-G'	8300	107	77.6	5.6	7.5	9.3
Mean	10550	135.6	78.9	5.5	7.3	9.2

[1] Transect location from Fig. 4.26
[2] Number of years that passed between the formation of two consecutive moraines (assuming t=600a)
[3] Number of years that passed between the formation of two consecutive moraines (assuming t=800a)
[4] Number of years that passed between the formation of two consecutive moraines (assuming t=1000a)

Table 4.4 Average retreat velocities (m a^{-1}) for the three hypothetical periods over which retreat might have taken place.

Transect	t=600a	t=800a	t=1000a
A-A'	15.8	12.0	9.0
B-B'	30.0	22.5	18.0
C-C'	28.6	21.5	17.2
D-D'	17.5	13.1	10.5
E-E'	10.4	7.8	6.3
F-F'	7.0	5.3	4.2
G-G'	13.8	10.4	8.3
Mean	17.6	13.2	10.5

The figures calculated above have to be regarded as a minimum since the barcode patterns are not complete along the loch axes, mainly because the location of palaeo-ice fronts could not be established with confidence. For the same reason, the barcodes do not contain crestline bifurcations that would yield additional information on the average frequency of moraine formation. Accounting for the readvances that can be inferred for at least two thirds of the moraines means that each counted moraine also includes the period of deformation. Thus, the true marginal response time must have been higher than the number shown in Table 4.3. As the glacier response times cannot be calculated without taking into account glacier velocity and mass balance (Bahr *et al.*, 1998), it can only be stated at present that the glaciers had a high mass turnover and thus very likely short response times. These speculations are strengthened by the evidence of high palaeo-precipitation values (Chapters 3.4.2, 3.4.3). It cannot be ruled out that some highly densely spaced recessional moraine arcs over small

areas reflect localised frontal response on an annual basis, but according to the evidence available at present (Tables 4.3, 4.4) this would seem unlikely for the whole period of the Younger Dryas. Moraines were formed on average every ca. 3-9 years at the larger glaciers and every ca. 7-19 years at the smaller glaciers (Table 4.3). The retreat velocity ranges from ca. 9 to 30 m a^{-1} for the larger glaciers to ca. 4 to 13 m a^{-1} for the smaller and shorter ones (Table 4.4).

To conclude, the present data strongly suggest that the glaciers during the Younger Dryas in the far NW Highlands were very active ones that generally responded dynamically to climate changes on subdecadal timescales.

4.4.3 Implications for the applicability of the englacial thrusting model

The data presented above carry a number of implications for the applicability of the englacial thrusting model, and these will be addressed below.

Hambrey *et al.* (1997) and Bennett *et al.* (1998) attempt to explain the occurrence of proximal rectilinear slopes with angles near 30° at one site in Glen Torridon by invoking englacial thrusting leaving a large number of debris septa stacked on top of each other after deglaciation is complete. As shown above (Fig. 4.4), most moraines in the study area have rectilinear slopes with similar angles to those observed by Bennett *et al.* (1998), Graham and Midgley (2000a) and Graham (2002). However, sedimentological evidence unequivocally demonstrates that these proximal rectilinear slopes correspond to the former ice-contact faces of small terrestrial ice-contact fans where material is at the angle of repose. Likewise, individual hummocks do not simply display "single facies or facies associations" as proposed by Hambrey *et al.* (1997: 624) and Bennett *et al.* (1998: 19) but a variety of sedimentary facies and facies associations affected by complex sedimentary structures.

Clast shape data have only sparsely and selectively been used by proponents of the englacial thrusting hypothesis (Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002). In all these cases, however, isolated clast shape measurements from shallow surface exposures that indicate subglacial transport have been used to infer the subglacial origin of a massive, matrix-supported diamicton, although sedimentary structures necessary to support such an interpretation cannot always be observed at such a small scale (Fig. 4.21). Nevertheless, such evidence has been extrapolated to whole suites of moraines which have in turn been inferred to consist of subglacially derived sediment that was elevated into an englacial position from where it melted out to produce stacked moraines. Although clast shape measurements are a powerful tool to discriminate different *transport histories* (cf. Chapters 2.1.2, 4.3.2), these data in isolation cannot be used to infer the *mode of deposition of entire units*. Only when clast shape data is placed into the context of large exposures, where the mode of deposition of individual units can be established, may important insights into the complex relationships between the transport paths and depositional processes be gained. By employing such a combined approach, this study has shown that material can be subglacially derived

and yet be deposited from a supraglacial position. This further supports the validity of the earlier note of caution on the sole use of small exposures as such an isolated sampling strategy does not allow for alternative hypotheses to be thoroughly tested.

The claim that thrusting is facilitated or aided by flow compression as the glacier advances up reverse bedrock slopes (Bennett, 1996; Hambrey *et al.*, 1997; Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002) is unsubstantiated by field evidence. Although never explicitly expressed in print, this claim appears to refer to steep reverse bedrock slopes only, because gentle reverse slopes such as those that occur along the valley flanks are regarded to be insufficient to facilitate thrusting by advocates of the thrusting model (N.G. Midgley, pers. comm., 2005). However, in the study area at least one location can be found where glacier margins advanced against steep slopes, but where evidence of thrusting is completely absent. Moraines in the lower parts of Bealach nam Meirleach indicate flow up a steep reverse bedrock slope (Fig. 4.13a), yet sections in these moraines show the same facies associations as those in SLK 1 (Appendix 1). It would therefore seem unlikely that reverse bedrock slopes *per se*, in a similar vein as the criterion of proximal rectilinear slopes in "moraine-mound complexes", can be used as an indication of possible locations where englacial thrusting has taken place. Clearly, such a hypothesis must be sedimentologically tested before it can be accepted as a general criterion that indicates palaeo-thrusting.

To conclude, transfer of the thrusting model to Scottish Younger Dryas glacier systems is based on a number of assumptions that do not fit the available evidence in large parts of the Scottish Highlands and Islands. The proposal of englacial thrusting as a formative process of Scottish "hummocky moraine" hinges on the morphological criterion of the presence of proximal rectilinear slopes, reverse bedrock slopes and sedimentological data extrapolated from shallow surface exposures. There is no sedimentological proof for englacial thrusting being responsible for any of the moraines summarised under "hummocky moraine" when investigated in detail. Consequently, the concept of englacial thrusting as a formative process of Scottish "hummocky moraine" has to be abandoned.

4.4.4 Implications for hitherto-proposed modern analogues

Two conflicting and mutually-exclusive modern analogues have been suggested to apply to the Younger Dryas in upland areas of Great Britain (Chapter 1.1.3.2). Based on the geomorphological and sedimentological data presented above and that reported in the literature (e.g. Benn, 1990, 1992a; Wilson and Evans, 2000) the proposal that Svalbard may form a good analogue for Upland Britain during the Younger Dryas (Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998; Graham and Midgley, 2000a) has to be seriously questioned. Likewise, however, Eyles' (1983a) suggestion that active Icelandic glaciers form a good analogue to those in the Scottish Highlands cannot be readily accepted as he did not provide any sedimentological data from Scotland. Thus, the present data situation is insufficient to identify a modern analogue for Younger Dryas glaciers in the NW Scottish Highlands

(and possibly elsewhere), and a re-evaluation of the evidence based on fieldwork and the published literature is necessary. The results of this work will be presented in Chapter 5.

CHAPTER 5 MODERN ANALOGUES

5.1 Introduction

The use of modern analogues has a long tradition in earth sciences as it allows informed inferences to be made about former environmental conditions based on observations of modern environments and processes. Such an approach helps to determine the genetic processes responsible for the deposition of sedimentary facies and formation of landforms with a high degree of confidence (e.g. Benn, 1990, 1992a; Reading, 1996; Evans *et al.*, 1999; Evans, 2003b, c; Evans and Benn, 2004). Knowledge about modern analogues can either be gained through studying the subject-specific literature or through comparative studies during fieldwork. In the present thesis, a field-based approach was chosen to ensure that the palaeo-environmental reconstructions inferred for Scotland were closely linked to observations and experience in the real world in order to avoid unrealistic output.

As discussed above, two conflicting models of moraine formation with opposing implications for the palaeo-glacial landsystem have been proposed for the Younger Dryas in Scotland. In order to ensure a realistic and effective comparison with the palaeo-environmental conditions reconstructed for the Younger Dryas in NW Scotland and to test the applicability of the models, the modern analogue study areas had to fulfil a number of criteria:

(a) they should represent endmembers with characteristic, contrasting glaciological and sedimentological conditions, e.g. a temperate and a cold-based glacial environment, to allow an informed statement to be made as to where the Scottish Younger Dryas landsystem would have its place between these two landsystems.

(b) Both landsystems should allow insights into the origin and formation of characteristic sediment-landform associations to be gained so that this experience could be related back to the NW Highlands.

(c) The contrasting suggestions of a modern temperate (Eyles, 1983a; Benn, 1990, 1992a; Bennett, 1990; Bennett and Boulton, 1993a; Wilson and Evans, 2000) and polythermal analogue for the Scottish Younger Dryas (Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002) necessitate that close similarities with the two proposed Scottish Younger Dryas landsystems exist so that a sensible comparison can be made without making too many (unreasonable or unjustified) inferences.

(d) Both study areas should have been studied previously so that the efforts can be concentrated on the formation of characteristic sediment-landform associations, and

(e) both areas should be easily accessible for field study.

Below, field data from two modern glacier forelands will be discussed, and similarities with the Scottish Younger Dryas glacier characteristics will be highlighted.

5.1.1 Nordenskiöldland, Central Spitsbergen

Three small, cold-based cirque glaciers in the more arid parts of Central Spitsbergen were chosen as the first modern analogue study in a continuous permafrost environment. This area shares many characteristics with the sites used to develop the englacial thrusting model of Scottish "hummocky moraine" formation (cf. Glasser and Hambrey, 2003). Glacier behaviour at both sites is influenced by the meltout of buried ice cores in a continuous permafrost environment, so that any potential postdepositional modification of landforms and sediments would occur under the same boundary conditions. As such, this allows an assessment into an aspect not adequately discussed in the englacial thrusting hypothesis, i.e. the problems associated with the meltout of glacier ice between englacial debris septa. As the glaciers chosen for the present study are less dynamic, cold-based and operate under more arid conditions, they are anticipated to be somewhat different in character from the maritime polythermal glaciers that have been investigated by proponents of the englacial thrusting model (cf. Hambrey and Glasser, 2002; Glasser and Hambrey, 2003). Last but not least, the three glaciers chosen for study are easily accessible and local support was already established, an important criterion as logistics in the high arctic can present a problem.

5.1.2 Krundalen, Jostedalsbreen, SW Norway

The second area chosen as a contrasting modern analogue case study is Krundalen in maritime SW Norway where three temperate outlet glaciers of the Jostedalsbre ice cap occupy the upper parts of the valley. SW Norway is latitudinally sufficiently close to Scotland, and the mountains have a similar character in terms of altitude and style of valley and ice cap glaciation. The area is also maritime in character, receives a large amount of precipitation and is thus comparable to the Scottish west coast today and possibly during the Younger Dryas (Chapter 3.4). In common with other valleys surrounding Jostedalsbreen, Krundalen contains clear sets of recessional moraines deposited since the Little Ice Age maximum (Bickerton and Matthews, 1993; Winkler, 1996a) that can be used to compare the geomorphological "signature" of a temperate, maritime valley glacier to the Younger Dryas evidence in Scotland. The area is well studied, but no research has been conducted explicitly on the formation and significance of ice-marginal, recessional moraines, and sediment-landform associations in this area in general. The only study of importance in this respect, although very different in its approach, is that of Winkler and Nesje (1999) on Brigdalsbreen on the SE side of Jostedalsbreen.

5.2 High-arctic analogues – examples from central Spitsbergen

5.2.1 Introduction

Extensive complexes that have been termed "ice-cored moraines" characterise the margins and mark the neoglacial maximum positions of central Spitsbergen glaciers. These "ice-cored moraines" consist of a zone of marginal supraglacial debris between 0.1 and 4 m thick, which retards the melting of underlying glacier ice (cf. Etzelmüller, 2000; Lyså and Lønne, 2001; Sletten *et al.*, 2001; Sørbel *et al.*, 2001). Because these features consist of supraglacial material covering the frontal part of the glacier, as opposed to isolated pockets of buried ice that occur within larger bodies of sediment, they are referred to as debris-covered ice-margins or zones below. In the context of this study, the effects of the continuous permafrost on glacier behaviour, landform formation and preservation potential are of particular interest.

The work presented here aims to (a) determine the processes leading to the formation of the supraglacial debris cover and geomorphological features developed within it, (b) improve the understanding of sedimentary processes associated with the decay of stagnant glacier ice in a high-arctic environment and (c) develop a conceptual model of landform evolution.

5.2.2 Study area and glacier characteristics

The Svalbard archipelago is underlain by continuous permafrost up to 500 m thick (Landvik *et al.*, 1988; Humlum *et al.*, 2003), and the mean annual air temperature at Longyearbyen airport is –6°C (Hagen *et al.*, 1993). With about 25% glacier cover, the study area of Nordenskiöld Land in central Spitsbergen is one of the least glaciated areas in Svalbard due to its relative aridity. The margins of three glaciers were investigated (Fig. 5.1): Larsbreen, Longyearbreen and an unnamed glacier on the northern side of Nordenskiöldtoppen, hereafter termed Nordenskiöldtoppenbreen, all form part of the Longyeardalen catchment. Mean annual precipitation at the equilibrium line altitudes of these glaciers is between 500 and 700 mm water equivalent (Humlum, 2002).

Radio-echo soundings and glaciological investigations at Longyearbreen and Larsbreen demonstrate that these are largely cold-based and that subglacial topography is V- rather than U-shaped, indicating that subglacial erosion has been minimal (Tonning, 1996; Etzelmüller *et al.*, 2000). In contrast, Nordenskiöldtoppenbreen occupies a broad, flat, cirque-like depression that leads into a plateau (Platåberget, ca. 450 m asl). Nordenskiöldtoppenbreen is smaller and at a higher altitude, so it is assumed to also be predominantly cold-based due to the penetration of permafrost underneath thin glacier margins (cf. Björnsson *et al.*, 1996). There is no known surging history for these three glaciers (Liestøl, 1969, 1993; Hagen *et al.*, 1993; Etzelmüller *et al.*, 2000).

Glaciers on Svalbard reached their Little Ice Age (LIA) maximum around 1900 AD (Svendsen and Mangerud, 1997), and the surface of Longyearbreen has lowered by up to 50 m since ca. 1936

(Justad, 1997) with similar figures having been obtained for nearby Rieperbreen (Lyså and Lønne, 2001) and other glaciers in the area (Ziaja, 2001). Signs of frontal retreat cannot be found at any of these three glaciers, which is characteristic of many non-surging central Spitsbergen glaciers (Etzelmüller, 2000; Etzelmüller *et al.*, 2000; Ziaja, 2001). Hence, the marginal positions of these glaciers are out of equilibrium with climate by > 100 a in this part of Svalbard, as they have been stagnating and downwasting since reaching their LIA maximum extent around 1900 AD.

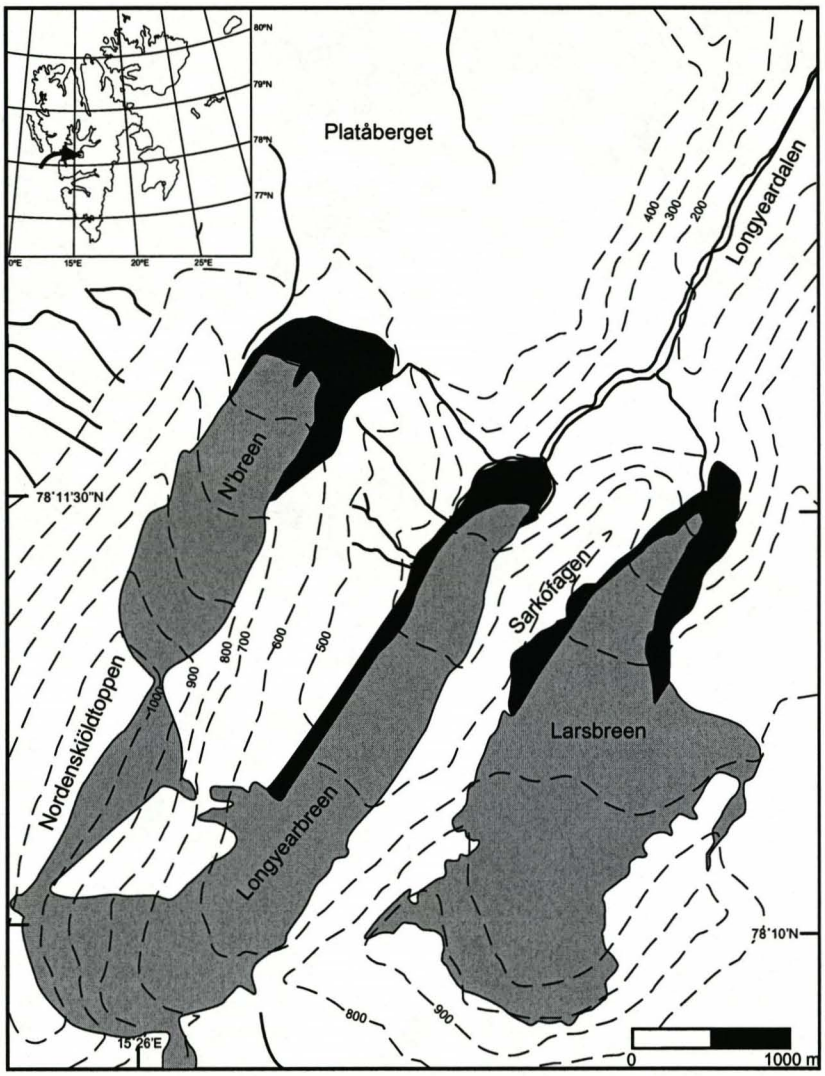


Fig. 5.1 Overview map of the study area in Central Spitsbergen showing the three glaciers studied. Black envelopes mark the location of the ice-marginal debris-covers (cf. Appendix 4).

At all three glaciers the LIA maximum is marked by accumulations of supraglacial debris along the presently buried ice front. The width of these debris-covered ice-margins is up to 600 m on Nordenskiöldtoppenbreen, 400 m on Longyearbreen and ca. 200 m on Larsbreen (Fig. 5.1), covering up to a third of the glacier surface. Debris thickness varies across individual zones (Table 5.1). Larsbreen and Longyearbreen share a similar surface form; concave in the accumulation zone and

convex at lower altitudes, while Nordenskiöldtoppenbreen has a more linear slope. Larsbreen and Nordenskiöldtoppenbreen flatten out where the debris-cover begins.

Table 5.1 Summary of characteristics of the three glaciers studied.

Glacier	Nordenskiöldtoppenbreen	Longyearbreen	Larsbreen
Size (km^2)	1.7	3.0	3.2
Area of frontal belt (km^2)	0.55	0.55	0.50
Area covered by debris (%)	32.4	18.3	15.6
Average debris thickness (m)	0.38	1.84	1.33
Standard deviation (m)	0.34	1.32	0.72
Number of measurements	32	10	12
Minimum (maximum) thickness of debris (m)	0.05 (1.2)	0.2 (4.0)	0.2 (2.5)

Tertiary sandstones, siltstones, shales and localised coal seams of the Palaeocene and Eocene Van Mijenfjorden Group, which dip gently to the WSW, underlie the study area (Hjelle, 1993; Dallmann *et al.*, 2001). One of the most prominent strata is the Grumantbyen Formation, which consists of sandstones and forms prominent cliffs and plateau surfaces such as those surrounding Longyeardalen and Platåberget (Dallmann *et al.*, 2001). Fossils such as bivalves, calcareous foraminifera, worm tracks and impressions of leaves and other plant remains are common in the Battfjellet Formation which partly overlooks the slopes in the accumulation areas of all three glaciers (Major *et al.*, 2000; Dallmann *et al.*, 2001) providing ideal tracers for the source area of glacial debris.

5.2.3 Debris-covered ice-margins

5.2.3.1 Physical characteristics

All three margins are covered by a compact, structureless (massive) clast to matrix-supported diamicton (Fig. 5.3a) (see above and Table 5.1). Its matrix is usually silt-rich (Fig. 5.2), and the a-axes of embedded clasts vary from 0.2 to 4 m with all local lithologies present, although larger clasts are predominantly sandstone. On Nordenskiöldtoppenbreen and Larsbreen, blocks of siltstone and shale of up to 7 m in diameter stand out at the surface surrounded by weathered aprons of scree (Figs. 5.3d, 5.4, 5.5).

All three glaciers show ice-cored lateral moraines of varying dimensions (e.g. Fig. 5.3b). In addition, concentric controlled moraines mimicking the buried ice margin can be traced throughout the debris-covered margin of Nordenskiöldtoppenbreen (Figs. 5.3a, 5.5). On both sides of Larsbreen, lateral moraines grade into an amorphous, debris-covered zone (cf. Fig. 5.1). On Longyearbreen, the whole western side of is flanked by a prominent lateral moraine (Fig. 5.1), although along the eastern margin, the ice-cored lateral moraine is less well-developed. In all cases, test pits along transects perpendicular to these moraines have shown that the thickness of debris does not exceed 0.15 m.

Moreover, the debris surface closely mimics the surface of the underlying ice, so that controlled moraine ridges on the surface are the result of ridges in the glacier ice.

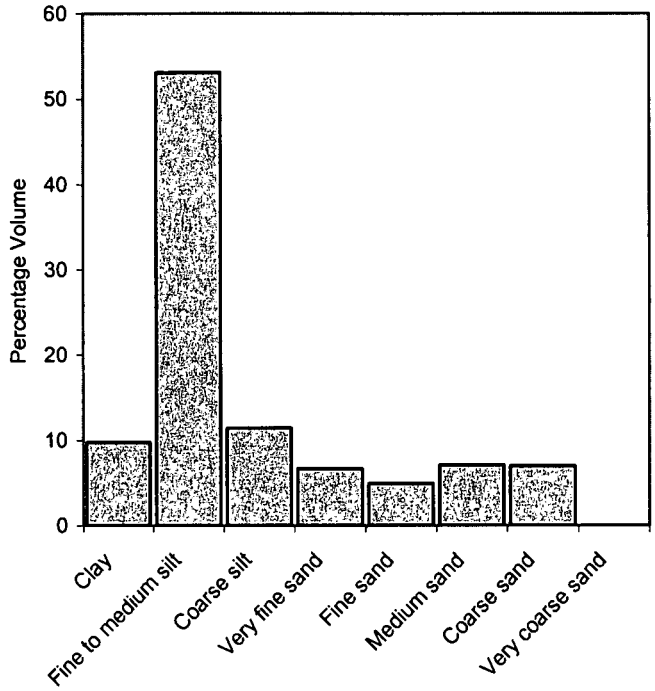


Fig. 5.2 Particle size fractions of matrix material in the debris-covered zone of Larsbreen.

The north-eastern margin of Larsbreen is bounded by three ridges, separated by steep-sided gullies (Fig. 5.4). Clast shape indicates that these ridges (Figs. 5.6b-d), which are devoid of fine matrix material, were derived from avalanche debris originating on the eastern valley side (Figs. 5.3i, 5.6a, f). Both ridges and avalanche cones consist of very angular to angular, prolate clasts of sandstone and siltstone with very few subangular clasts. These ridges have been interpreted as a talus-derived rock glacier that was pushed up in front of the glacier during its neoglacial advance (Humlum, 2005). Avalanche material also overlaps the sides of Larsbreen and Longyearbreen indicating a potential source area (Fig. 5.6e, g), although some mixing with rounded material is evident (see below). Steep sandstone cliffs (Fig. 5.3j) overlook the three glaciers in their source areas, and accumulations of openwork, very angular to angular rockfall material are frequently found at the foot of such cliffs on the glacier surface.

Localised sediment accumulations are found on all three glaciers. A discrete debris ridge unequivocally associated with an englacial wedge was found on the surface of Larsbreen (Figs. 5.3g, 5.4). This ridge, which is lithologically anomalous to the rest of the debris layer, consists of laminated and cross-bedded sands. Due to the preservation of bedding structures that would undoubtedly be destroyed during prolonged glacial transport this ridge is interpreted as a crevasse-fill

formed from supraglacial stream deposits. Moss fragments contained within the sands gave an uncalibrated ^{14}C age of 150+/-39 yr (calibrated 1δ age: AD1682-1947; AAR-7999). The source area for both sands and moss is inferred to be nivation niches just above the LIA trimline on the slopes north of Trollsteinen ca. 3 km south-east of the sampling site; the ^{14}C-age is consistent with a formation during, or since, the Little Ice Age (Lukas *et al.*, 2005a).

Six mounds with heights of up to 1.2 m, widths of up to 4 m and lengths of up to 5.2 m occur at the eastern side of Larsbreen; they form a line along which they are regularly spaced. Artificial exposures revealed well-sorted, rounded to well-rounded sand and gravel (Figs. 5.6j, k) beneath a thin (< 0.1 m) veneer of angular, supraglacially-derived clasts. Towards the margins of these mounds, stratification is disturbed, and the deposits show normal faults, interpreted as collapse features. The sand and gravel rests on diamicton. These isolated deposits are interpreted as pool fills of former supra- or englacial meltwater channels that were topographically inverted during downmelting. Clast shape analysis of neighbouring deposits shows that mixing of supraglacial and glaciofluvial material has occurred near the cones (Fig. 5.6h, i).

Upglacier of the main debris-covered zone on Longyearbreen, a ridge of matrix-supported diamicton trends perpendicular to ice flow (Fig. 5.3h). Similar features can be found in the accumulation area of Nordenskiöldtoppenbreen where they trend both parallel and perpendicular to ice flow. In all cases the clasts are angular, the debris cover is < 0.2 m thick, and excavation shows no evident connection to englacial debris bands. This material might represent rockfall onto the ice surface which was then transferred englacially until it melted out at the surface (cf. Kirkbride, 1995; Sletten *et al.*, 2001).

5.2.3.2 Clast provenance

Based on the V-shaped subglacial topography of Larsbreen and Longyearbreen and absence of substantial temperate ice patches, Tonning (1996) and Etzelmüller *et al.* (2000) suggested that basal erosion is minimal today. This implies that the primary debris source is supraglacial. Etzelmüller *et al.* (2000) note that clasts in the supraglacial debris belts of polythermal glaciers appear to be dominantly angular, which is also suggestive of subaerial debris sourcing. These assumptions are likely to apply at Nordenskiöldtoppenbreen as well, although the overall topography is less confined than for the former two glaciers. Clast shape analyses were employed to determine the origin and modification of clasts and to identify sedimentary processes.

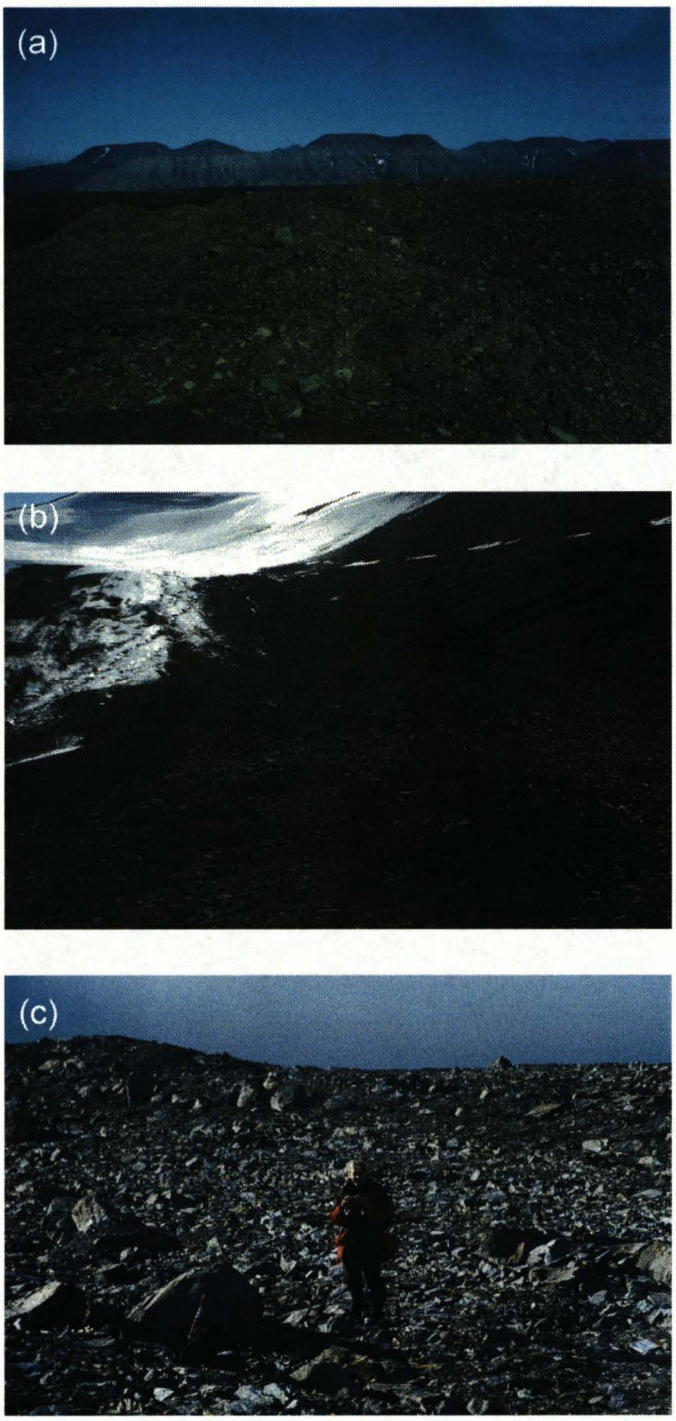


Fig. 5.3 (a) Surficial "ridges" on the surface of the debris-covered zone of Nordenskiöldtoppenbreen (view to the east). (b) Western lateral moraine on Nordenskiöldtoppenbreen (view to the south). (c) Close-up of subangular, blocky and partly striated clasts on the surface of Nordenskiöldtoppenbreen (person for scale).

Fig. 5.3 (continued) (d) Detail of a large shale bedrock fragment with surrounding scree slopes on the western side of Larsbreen. Rifle at the bottom left is 1.10 m long. (e) Close-up of openwork angular, prolate and oblate clasts in an area connected to a meltwater channel on Nordenskiöldtoppenbreen. Openwork structure is interpreted as evidence of winnowing of fine material. (f) Englacial debris bands cropping out on the surface Larsbreen suggesting horizontal stratification (view to the south-west).

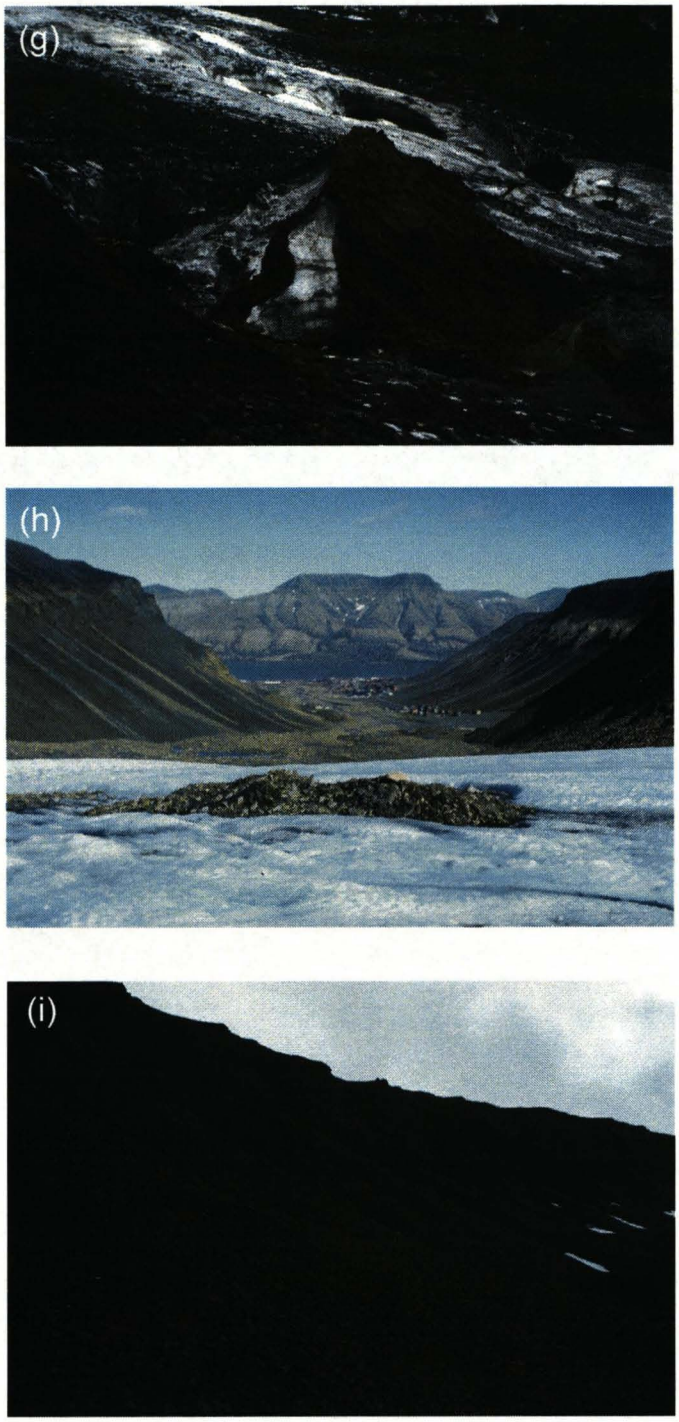


Fig. 5.3 (continued) (g) Cross-bedded sand filling a buried crevasse trace on Larsbreen. (h) Discrete debris accumulation on the surface of Longyearbreen perpendicular to ice flow. The thickness of this debris accumulation is < 0.2 m (view to the north-east). (i) avalanche debris cones on the eastern side of Larsbreen linking the free faces (left) with the glacier surface (outside right margin).

Fig. 5.3 (continued) (j) Overhanging rock towers overlooking the surface of Nordenskiöldtoppenbreen in its source area (view to the north-east).

Some clasts can be traced to source areas due to their containing distinctive fossil assemblages; on Nordenskiöldtoppenbreen, for example, a distinct unit of the Battfjellet Formation crops out in a narrow zone in the glacier source area above the ice, and several blocky and rod-shaped clasts of this unit were found in the debris-covered zone allowing a reconstruction of the respective transport paths. Embedded clasts of up to 1 m are dominantly rod-shaped and very angular (VA) to angular (A) (Fig. 5.7a, b, d), indicating little alteration during transport and, thus, supraglacial or englacial transport paths (cf. Boulton, 1978; Benn and Ballantyne, 1993, 1994). However, up to a third of the larger clasts in the same samples are blocky (having a high c:a ratio) and/or subangular (Figs. 5.7a, b, c). Together with striae on such clasts, this indicates that they have been transported subglacially. Frost shattering of the finer-grained lithologies is evident in the production of very angular prolate and oblate shards whereas sandstones tend to produce blockier shapes. Coal is rare in the debris-covered zones due to its low strength, and, where present, tends to be blocky.

Clasts in all three ice-marginal debris-covered zones (Figs. 5.6, 5.7, 5.9) have high C_{40} indices indicating predominantly oblate and prolate clasts with few blocky clasts. Larger proportions of subrounded, or even rounded, clasts only occur under special circumstances as described for Larsbreen above. On Longyearbreen, the admixture of subangular and subrounded material (Figs. 5.8, 5.9c-d) appears to indicate more efficient transport which might be linked to this being the longest glacier. Admixture of a small portion of fluvially-rounded clasts in sample LS5 (Fig. 5.9e) might relate to localised fluvial reworking of subglacial material.

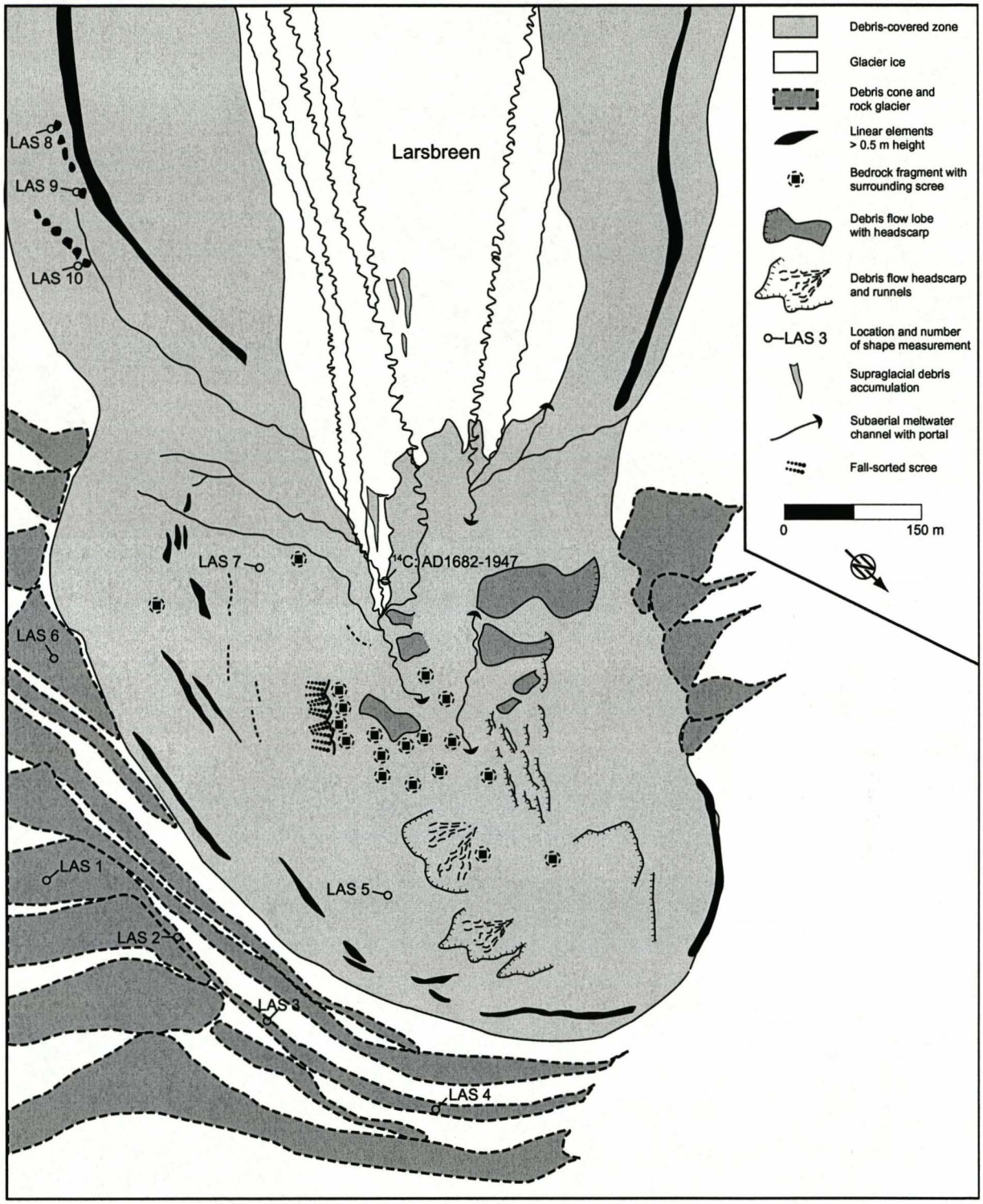


Fig. 5.4 Geomorphological map of Larsbreen showing the major elements that were identifiable in 2002 and 2003. Compare Fig. 5.13.

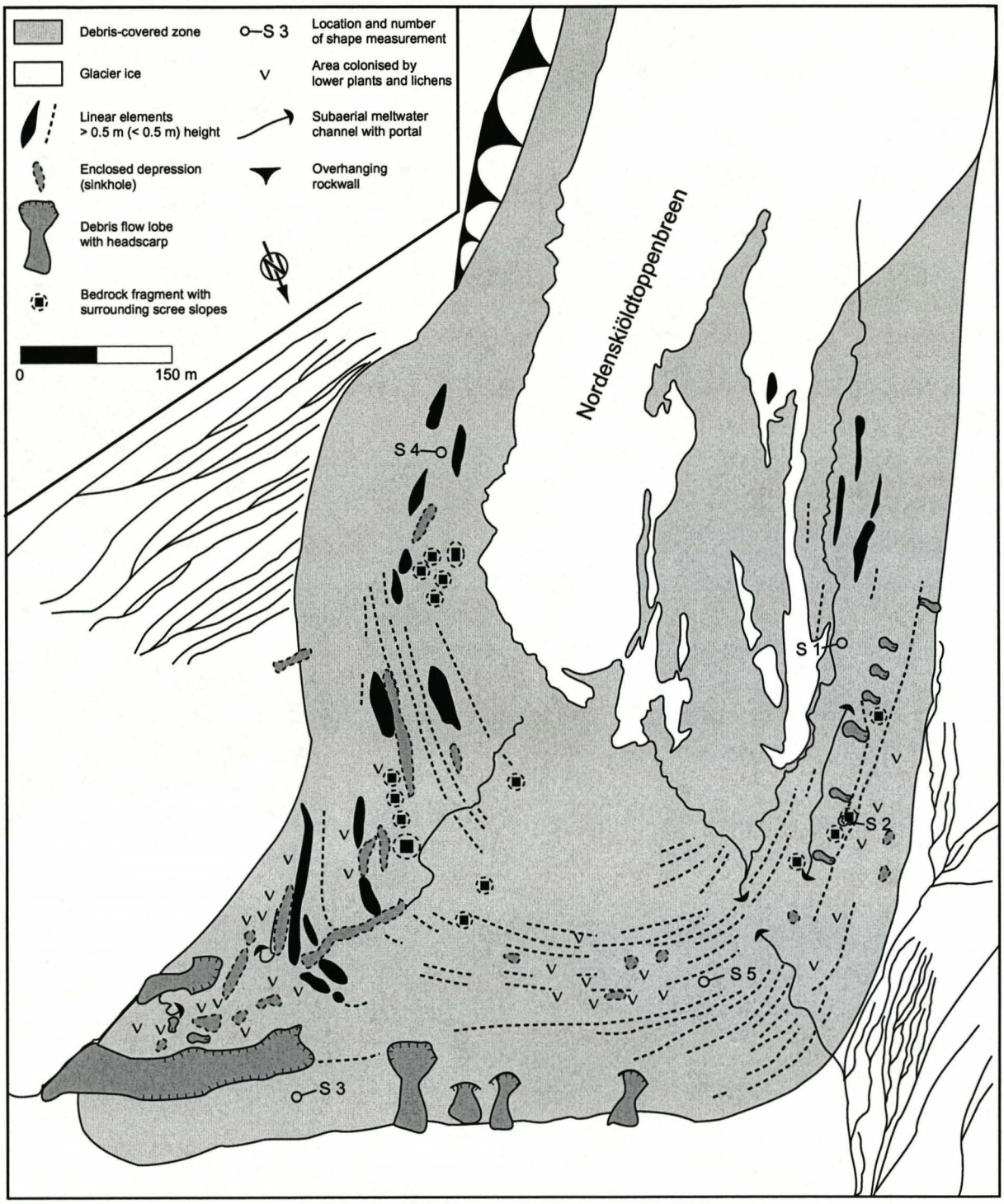
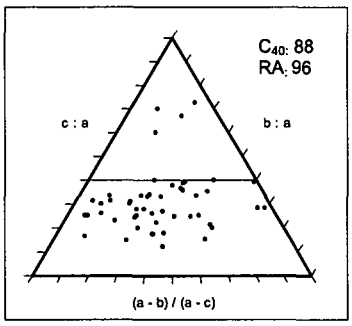
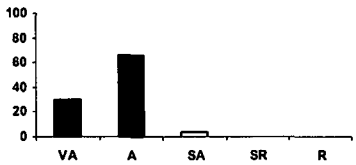
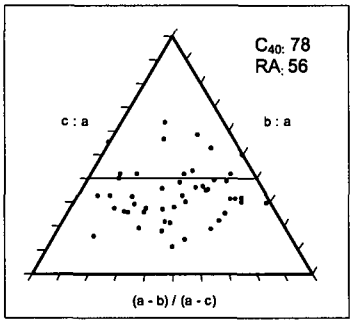


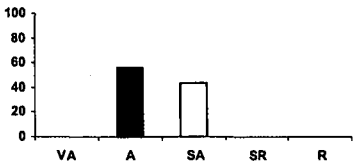
Fig. 5.5 Geomorphological map of Nordenskiöldtoppenbreen.



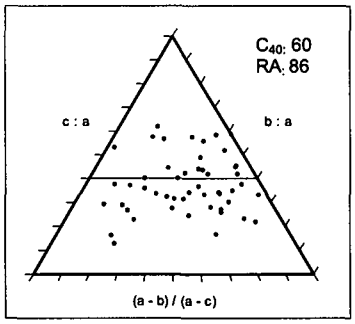


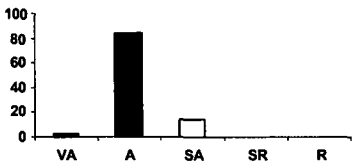
(a) LAS1



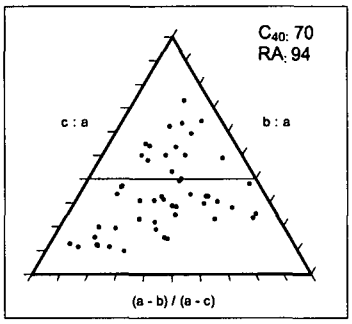


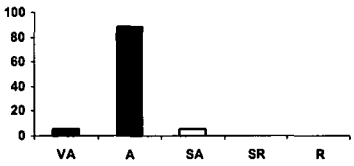
(b) LAS2



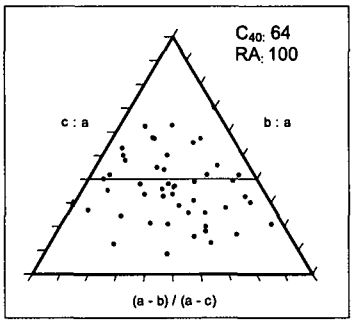


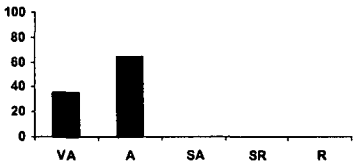
(c) LAS3



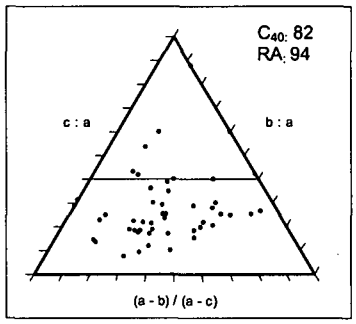


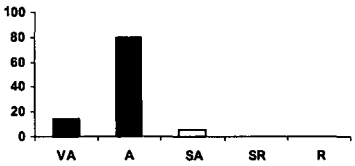
(d) LAS4



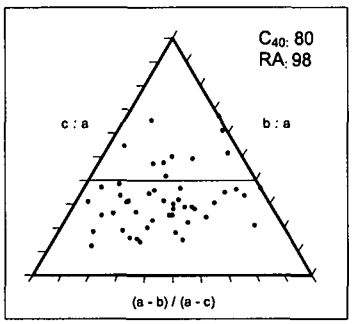


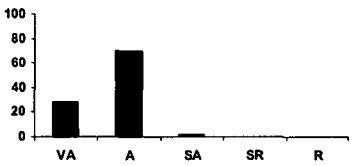
(e) LAS5



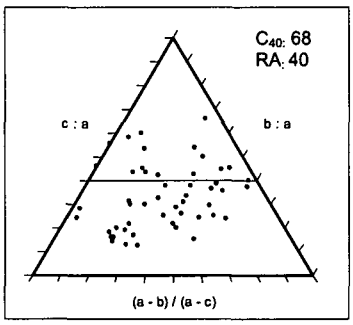


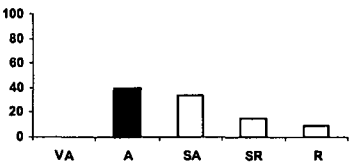
(f) LAS6



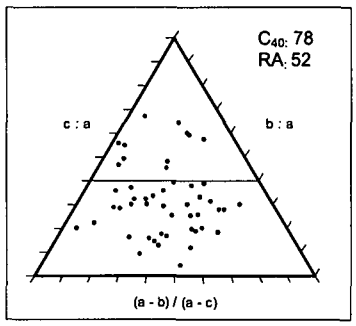


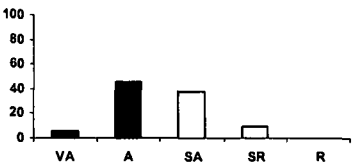
(g) LAS7





(h) LAS8





(i) LAS9

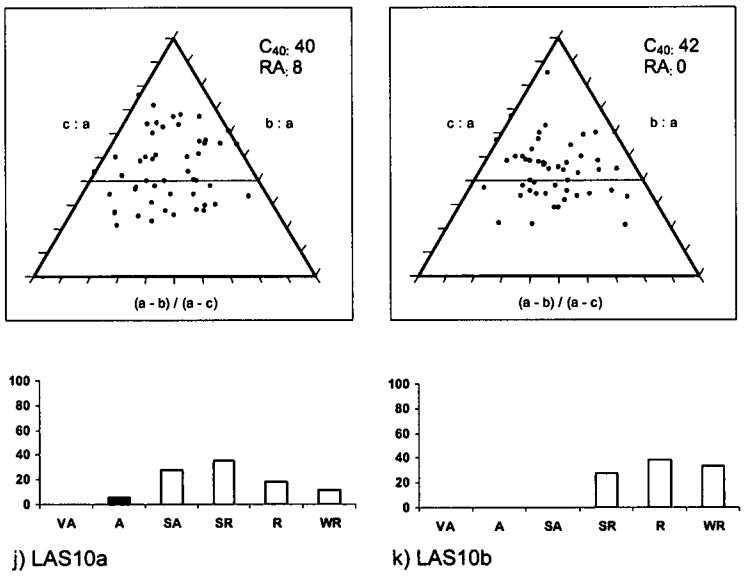


Fig. 5.6 Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Larsbreen (cf. Fig. 5.4).

The difference between purely subaerially sourced and transported material and the mixed composition of the ice-marginal debris-covered zones can best be seen at the margin of Larsbreen. Here, control samples taken from avalanche cones (Figs. 5.6a, f) have a high C_{40} index and a negligible subangular component. The latter is attributed to edge-rounding as a result of clast impacting during avalanches. Clast shape largely falls in the continuum between slabs and blocks. This compares well with samples taken from the rock glacier ridges in front of the debris-covered zone (Fig. 5.6b-d), although the C_{40} index is somewhat lower, probably due to weathering. A good example of the influence of weathering is given by sample LAS2 (Fig. 5.6b) which contains an unusually large proportion of subangular clasts. This was found to be due to the presence of voids underneath larger clasts at the surface which had been infilled with smaller clasts that had undergone weathering and rounding. In contrast, samples taken from within the debris-covered zone show a much wider spread of the C_{40} and RA indices (Figs. 5.6e, g-i), which is attributed to mixing of supraglacial and englacial with subglacial material rather than purely subaerial weathering. The presence of striae and facets on most of the subangular clasts strengthens this interpretation.

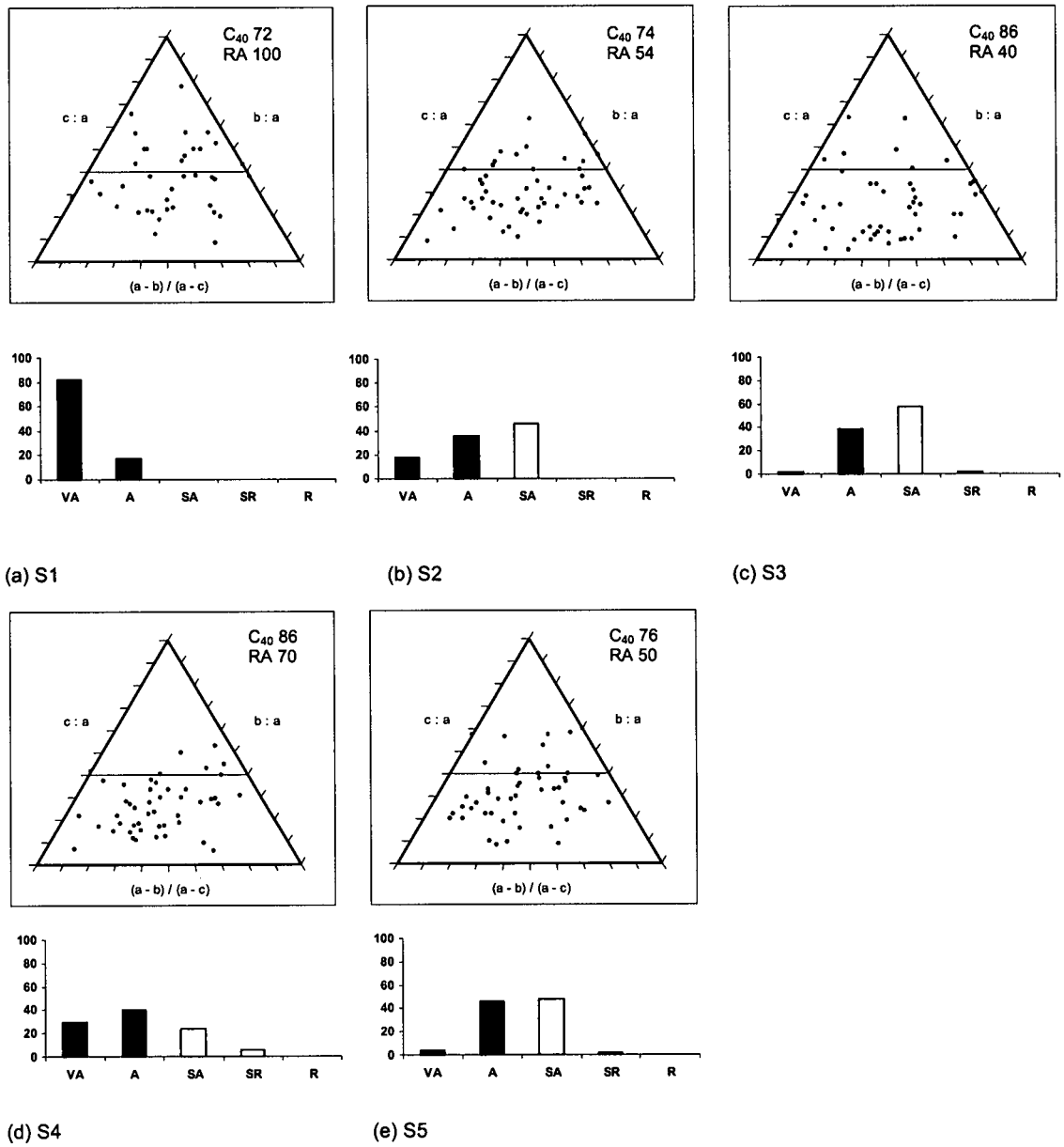


Fig. 5.7 Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Nordenskiöldtoppenbreen (cf. Fig. 5.5).

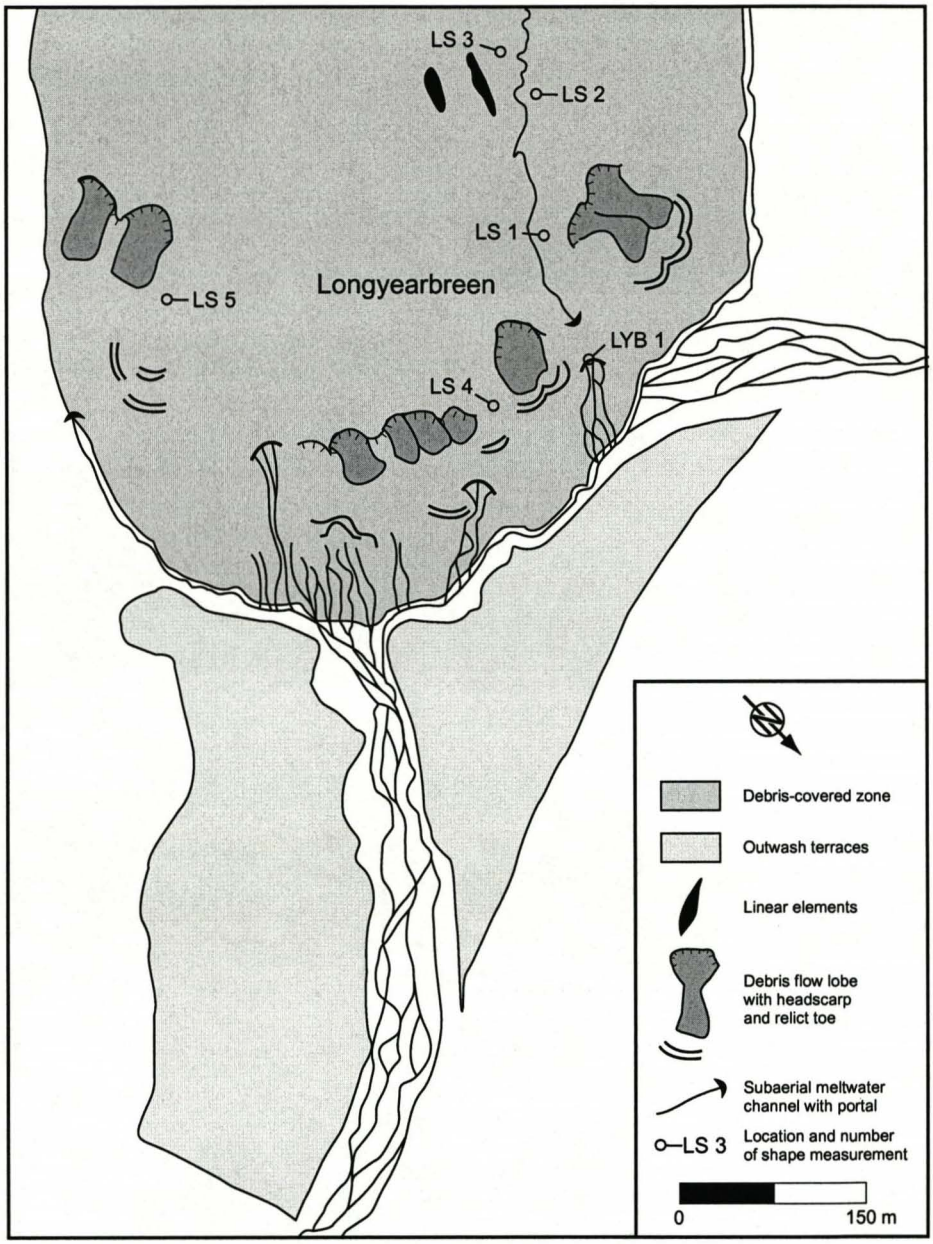


Fig. 5.8 Geomorphological map of the frontal part of the ice-marginal debris-covered zone of Longyearbreen.

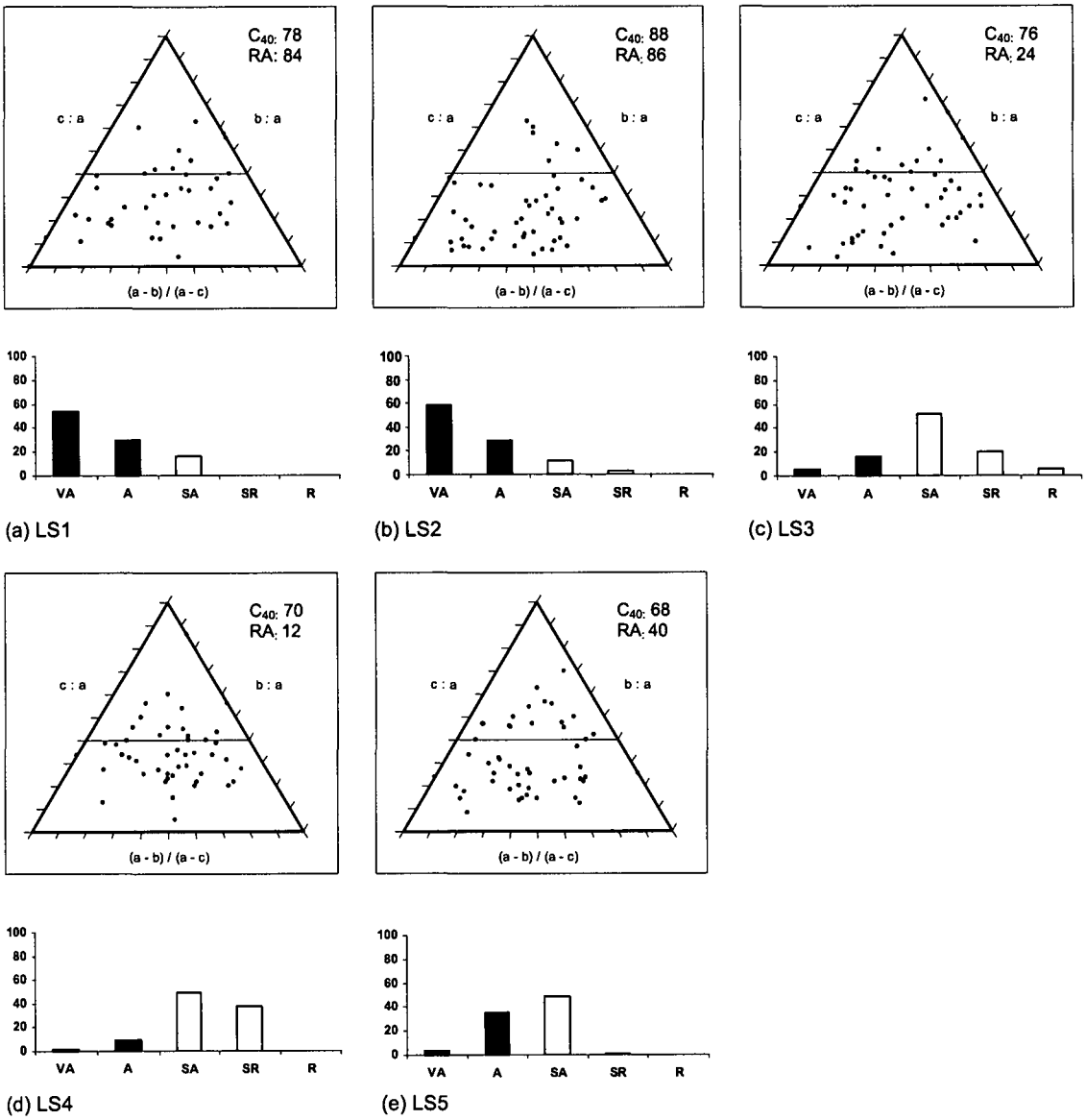


Fig. 5.9 Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Longyearbreen (cf. Fig. 5.8)

5.2.3.3 Mode of Formation

Comparison of clast shapes that dominate the debris-covered zones (Figs. 5.6, 5.7, 5.9) with control samples (Figs. 5.6a, f) indicates supra- or englacial transport. The fact that many clasts on Nordenskiöldtoppenbreen and Longyearbreen can be traced to distinct strata in the headwall indicates that rockfall is the predominant source of material. This material will then be buried by snow and incorporated into the glacier, most likely being transported englacially (cf. Kirkbride, 1995; Spedding and Evans, 2002). Along the frontal margin of Larsbreen, cliffs are separated from the glacier surface by debris cones over a horizontal distance of up to 100 m. Thus, in this case it is not the immediate effect of larger blocks fallen onto the glacier surface, but the weathered and smaller avalanche material that largely contributes to the formation of the finer-grained debris-covered zone. In all three cases,

however, supraglacial and englacial transport of supraglacially-derived (rockfall and avalanche) deposits appears to dominate in the present glacial transport systems.

Subglacial transport has also played a role in the formation of all three debris-covered zones as evident from the presence of striated, edge-rounded and blocky clasts. Also, although there is a large component of very angular and angular oblate and prolate clasts indicative of supraglacial and englacial transport in the debris-covered zones, a large matrix component is present. Supraglacial transport does not generate large amounts of fine material due to the absence of crushing and grinding (Boulton, 1976; Benn and Evans, 1998); additional and potentially significant nival (see above) and aeolian dust sources are suspected (Lukas *et al.*, 2005a). Hence, subglacial transport, as evident from the presence of striated boulders, contributed to the formation of the debris-covered zones; subaerial weathering, nivation and aeolian processes alone can probably not account for the large amount of matrix material. Therefore, all three glaciers must have had more extensive temperate areas and were much more active than today during the Little Ice Age due to an increased thickness (cf. Sletten *et al.*, 2001; Lyså and Lønne, 2001). Such a formerly polythermal regime would explain the widespread occurrence of subglacially-transported clasts.

In summary a dominant source of material was from supraglacial positions can be reconstructed. However, the elevation of subglacially-transported debris, presumably by meltout along flowlines near the snout, also contributed to the formation of the supraglacial debris covers of all three glaciers and indicates that the glaciers were polythermal during the LIA and are possibly cold-based now as a result of their comparatively small size (cf. Glasser and Hambrey, 2001; Lyså and Lønne, 2001).

Where relatively thin (≤ 0.3 m), isolated debris patches melt out, cones tend to form, as ice beneath the thickest part of the debris patch is protected from ablation. As the surrounding ice surface melts down, the slope angles of the cone increase until debris becomes unstable resulting in radial redistribution (cf. Fig. 5.3g, h), explaining the relatively even debris cover parallel to the underlying ice surface along "medial moraines" (see above). Evidence of thrusting, as inferred at the margins of some Svalbard glaciers (e.g. Hambrey *et al.*, 1997, 1999; Bennett *et al.*, 1998), was not found, and this generally agrees with findings from glaciers nearby (Sletten *et al.*, 2001; Lyså and Lønne, 2001).

Melting out of the bedload of former drainage channels (cf. Pelto, 2000; Spedding, 2000), individual rockfalls and crevasse-fills form localised concentrations of debris which are likely to be reworked by meltwater and gravitational sliding to form a more continuous debris cover. Emerging debris bands or crevasse fills containing fine debris would probably become liquefied to form extensive debris covers (cf. "flow tills", Boulton, 1968). In addition, compressional longitudinal flow towards the margins tends to cause medial moraines and surface debris to spread laterally across the glacier terminus (Anderson, 2000).

5.2.4 Degradation of ice-marginal zones

Wet, cohesive debris flows are ubiquitous on sloping surfaces in the marginal zones (Figs. 5.4, 5.5, 5.8, 5.10a, b, e, f). Active debris flows commonly incise relict flow lobes, in places giving the surface a staircase-like appearance (Figs. 5.4, 5.5, 5.8). No relationship between debris flow activity and slope angle was found with both active and inactive flows on Larsbreen existing on slopes ranging from 9-43°, suggesting that the trigger for debris flows is likely to be a result of variation in moisture conditions rather than a critical slope angle. Mean slope angle of active debris flows was 21° while that of inactive debris flow was 20°, suggesting that drying out rather than slope angle may be responsible for debris stabilisation. Active flow scarps are located around the crestlines of ridges, such as along the lateral moraines (Figs. 5.4, 5.5). The largest debris flows are bounded by headscarps up to 2 m high (Figs. 5.10a, b) that usually show signs of retrogressive failure in the form of opening tension cracks (Figs. 5.10a, c). Observations of retreat rates and processes show that, in summer conditions, up to 0.5 m width of debris around head scarps of up to 2 m in thickness and several tens of metres in length can be removed in a week (Lukas *et al.*, 2005a). Failure occurs in discrete events with surface cracks expanding behind the scarp in the days preceding the toppling failure of the scarp face (Fig. 5.10a). Failed material is then liquefied by meltwater and flows downslope, forming flow lobes and runnels (Fig. 5.10b).

The largest flows occur on the distal terminal slopes of Longyearbreen and Nordenskiöldtoppenbreen and flanking the central supraglacial drainage channel on Larsbreen (Figs. 5.4, 5.5, 5.8). Smaller debris flows form along the margins of meltwater channels (Fig. 5.10e, f). Failures propagate upslope, creating further instability, and the remobilised material is rapidly evacuated from the debris-covered zone via the fluvial system (Fig. 5.10e, f; cf. Barsch *et al.*, 1994; Alley *et al.*, 1997; Etzelmüller, 2000). Where debris flows occur, they expose the underlying buried glacier ice allowing rapid melting. Variation in the distribution of debris thickness results in differential ablation beneath the debris cover and the creation of local relief at the ice surface (Lukas *et al.*, 2005a). Meltwater at the ice-sediment interface facilitates translational failure of the whole layer as a cohesive unit, with incipient failures indicated by tension cracks (Figs. 5.10c, d). Removal of material in a narrow zone subparallel to the channel by thin flows results in further instability which in turn is compensated for by additional debris flows upslope. The effectiveness of the ensuing chain reaction can best be seen at Larsbreen where comparison of photo-mosaics suggests that the glacier surface has lowered by up to 5 m around the central meltwater stream between August 2002 and August 2003 (cf. Fig. 5.13).

Fig. 5.10 (continued on page 136) (a) Close-up of debris flow scarp on eastern side of Larsbreen showing numerous tension cracks and dried material of previous slump. (b) Overview of north-eastern funnel-shaped debris flow complex showing several inset flow lobes, runnels and skinflows. (c) tension cracks in thin (0.15 m) debris cover on Nordenskiöldtoppenbreen. The ice surface in the foreground has been cleaned to show the conditions at the sediment-ice interface where fine material has been winnowed by topmelting of the ice to form an openwork fine gravel lag on which the debris cover rests. (d) Wider (0.2 m) tension cracks on ca. 1.0 m-thick debris cover on western lateral moraine of Nordenskiöldtoppenbreen. (e) Exposed ice along the main meltwater channel on Larsbreen. Debris flows along this channel propagated upslope and led to increased melting of the formerly buried ice surface during the observation period in 2003. (f) Smaller-scale erosion of a former englacial meltwater channel flanked by debris flows and unstable material upslope in the north-eastern part of Nordenskiöldtoppenbreen (view to west). (g) Alternating units of clast-supported, stratified diamicton interpreted as stacked debris flows in the steep "frontal wedge" of Longyearbreen. Compare Fig. 5.13.

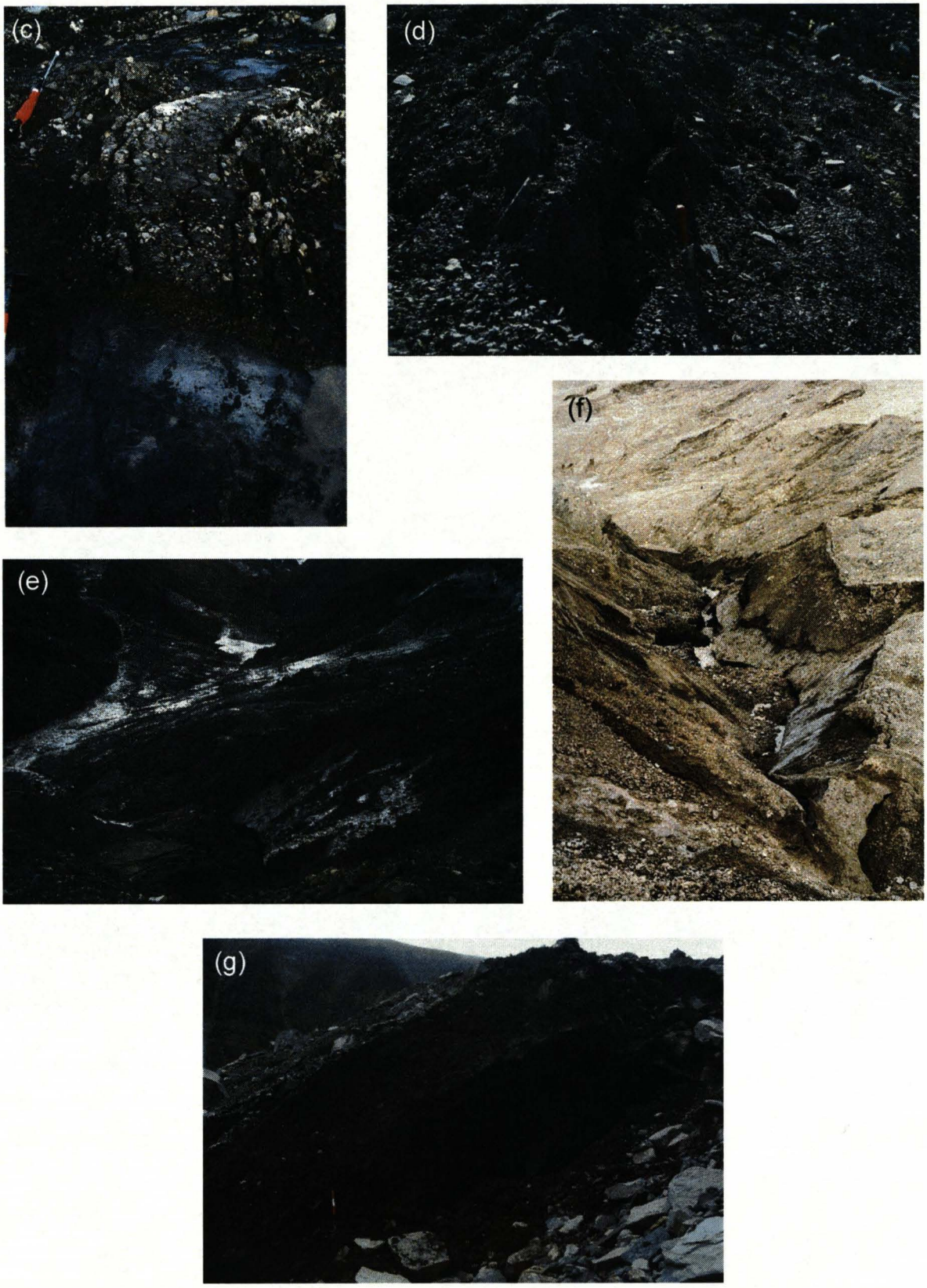


Fig. 5.10 (continued)

Meltwater saturation of debris can also result in large skinflows (Fig. 5.10b). Increased mobility of thinner debris, which is more likely to be saturated by meltwater, creates a positive feedback for mass movement processes, in which areas of more rapid melt maintain local slope angles, leading to rapid widening of the channel and downslope movement destabilising the debris cover upslope, thus perpetuating the erosive undercutting of the surrounding debris cover. Therefore, enhanced melting leads to further debris flows, and the system enters a self-reinforcing cycle.

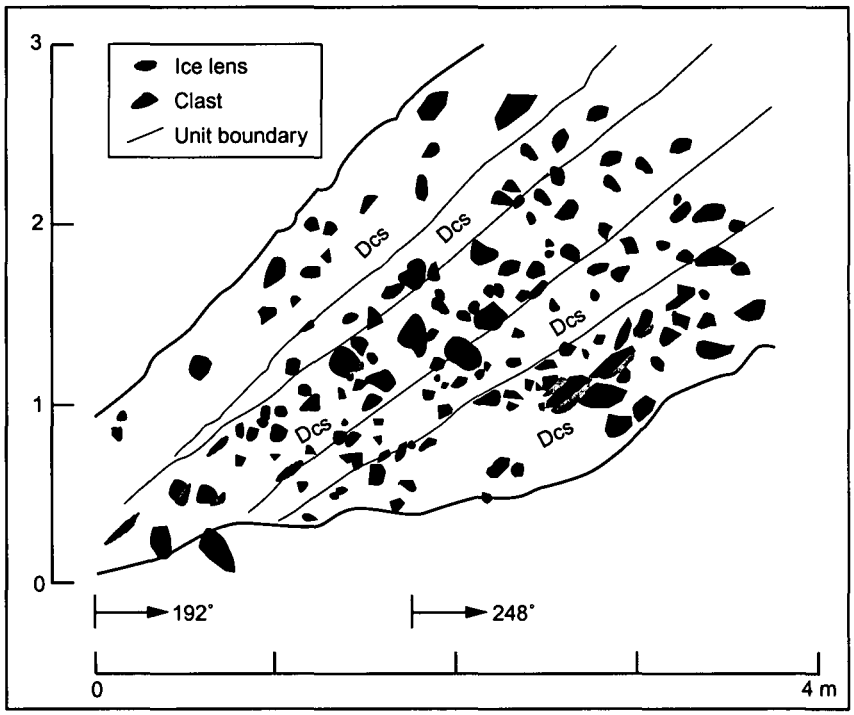


Fig. 5.11 Sedimentary log of section LYB 1 exposing alternating units of clast-supported, stratified diamicton (Dcs) interpreted as stacked debris flows.

The observed increase in debris thickness towards the glacier margin can be accounted for by a combination of supra-, en- and subglacial debris concentrations being greatest at the margins and ice flow further concentrating debris towards the terminus (Kirkbride, 2000). This study shows that, in the case of central Spitsbergen glaciers, these effects are exacerbated by ubiquitous debris flows which shift debris centrifugally thereby constantly increasing the 'frontal wedge' by stacking (Fig. 5.10g). Indeed, this may be the dominant process in this environment. Only one section exposing the sedimentary end product of this process was found (LYB 1 in Fig. 5.8). Fig. 5.11 shows very compact, crudely stratified, clast-supported diamict units that steeply dip towards the NNE by ca. 20° and are interpreted as stacked debris flows. The matrix throughout the exposure is composed of silty to fine sand with numerous aligned clasts in the fine to coarse gravel fractions and boulders with a-axes of up to 2 m; the alignment of clasts causes the apparent stratification. Numerous ice lenses were observed

in cracks parallel to the dip of the units and around larger boulders. Undercutting by an emerging englacial stream led to collapse of a large part of the exposure two days after it had been logged.

Slope steepening and undercutting due to incision of supraglacial meltwater is a frequent process that leads to surface lowering, inducing instabilities and leading to debris flows. The association of the largest flows around the frontal distal slopes of Nordenskiöldtoppenbreen with supraglacial and englacial channels is best explained this way as indeed some of the largest debris flows appear to have deposited very little sediment at the foot of the flow. Meltwater activity also tends to winnow out fine material from the diamicton, leaving localised areas of coarse, openwork deposits (Fig. 5.3e).

Subhorizontal marginal areas support a thin cover of lichens, mosses and small vesicular plants (Figs. 5.4, 5.5, 5.8). Enters (2000) has described such small plant colonies on Rieperbreen in neighbouring Endalen and was able to link the occurrence of vegetation to relatively stable areas that were not disrupted by debris flow activity. This evidence of marginal stability emphasises the importance of supraglacial meltwater in destabilising the debris cover.

5.2.5 Synthesis of degradational processes

The observations and data presented above suggest complex links between individual processes. Fig. 5.12 shows a conceptual model that synthesises the observed evidence into a degradational process-response system. Conditions of the central Spitsbergen glacier system are initially characterised by a steep frontal slope that presents the first instability along which material is mobilised and transferred away from the glacier front. Collapse of englacial meltwater channels forms steep gradients within a formerly continuous debris cover. Debris flows into these channels result in removal of material and probably enhance fluvial downcutting into the ice. Removal of material in a zone close to the channels induces further debris flows upslope. In all steps, debris flows result in the thinning or complete removal of supraglacial debris, hence enhancing melting. Enhanced melting results in further debris flows. Supraglacial material – once mobilised by debris flows – thus enters a self-maintaining and self-reinforcing cycle of degradation (Fig. 5.12).

The strong link between the degradation of ice-marginal debris-covered zones and the occurrence of roof collapse triggered by fluvial undercutting is particularly evident on the three glaciers studied. The main drainage routes on Larsbreen and Nordenskiöldtoppenbreen cross the ice-marginal debris-covered zones as opposed to those of Longyearbreen that run largely parallel to the debris-covered zones in a marginal position (Etzelmüller *et al.*, 2000). The data therefore support the notion that (a) most surface-debris reworking is initiated by erosion and destabilisation by surface meltwater, and (b) once fluvial erosion has commenced, rapid degradation of the debris-covered zones ensues (King and Volk, 1994; Etzelmüller, 2000). These observations confirm theoretical reasoning that, where there is efficient coupling between glacial and glaciofluvial systems, most of the sediment

will be removed (Alley *et al.*, 1997; Benn *et al.*, 2003). This effect can best be seen at Larsbreen where large-scale debris removal along the central channel initiated enhanced melting and degradation (Figs. 5.13).

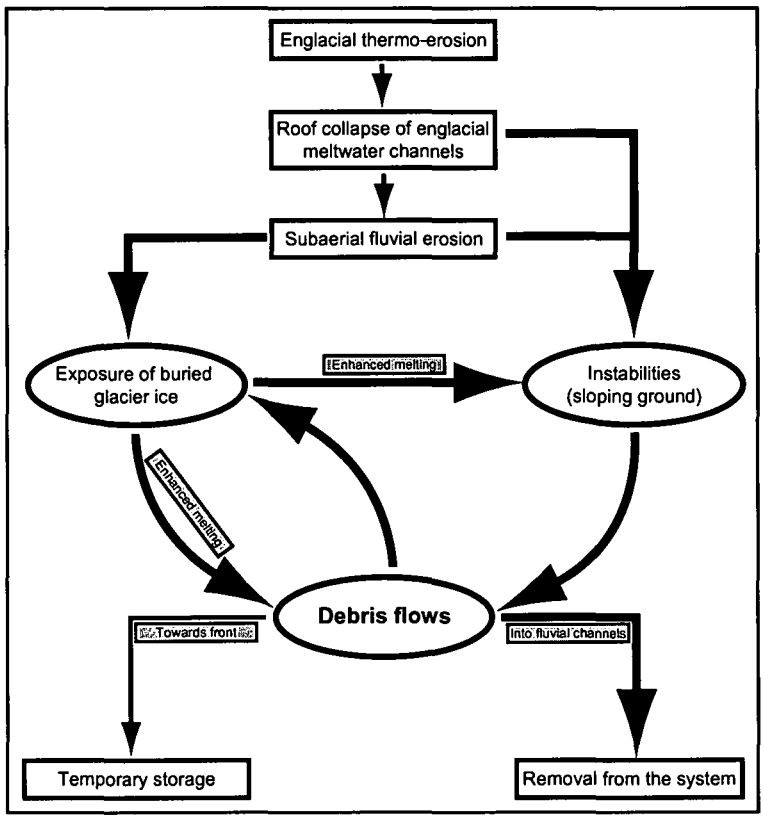


Fig. 5.12 Conceptual model of the process-response system of the degradation of ice-marginal debris-covered zones of cold-based to polythermal Central Spitsbergen glaciers. Temporarily-stored material will be subject to reworking due to gradual degradation of the underlying ice.

5.2.6 Implications for preservation potential and landform genesis

Debris flows redistribute material where the supraglacial debris cover is thinner than the active layer. On some glaciers, mass movements redistribute debris in talus fans to topographic lows (Iwata *et al.*, 1980), within which ablation is then retarded. Debris redistribution can cause topographic inversion, where hollows become highs and vice versa (Clayton, 1964; Kjær and Krüger, 2001; Hands, 2004). Such inversion may happen several times during a period of glacial retreat and is an important process by which debris is distributed more uniformly across the glacier (Clayton, 1964; Drewry, 1972; Watson, 1980; Anderson, 2000). However, in the case of the glaciers studied here, most of this material is evacuated along meltwater channels. Consequently, the sedimentary end product of this style of marginal deposition is unlikely to retain forms inherited from the supraglacial system.

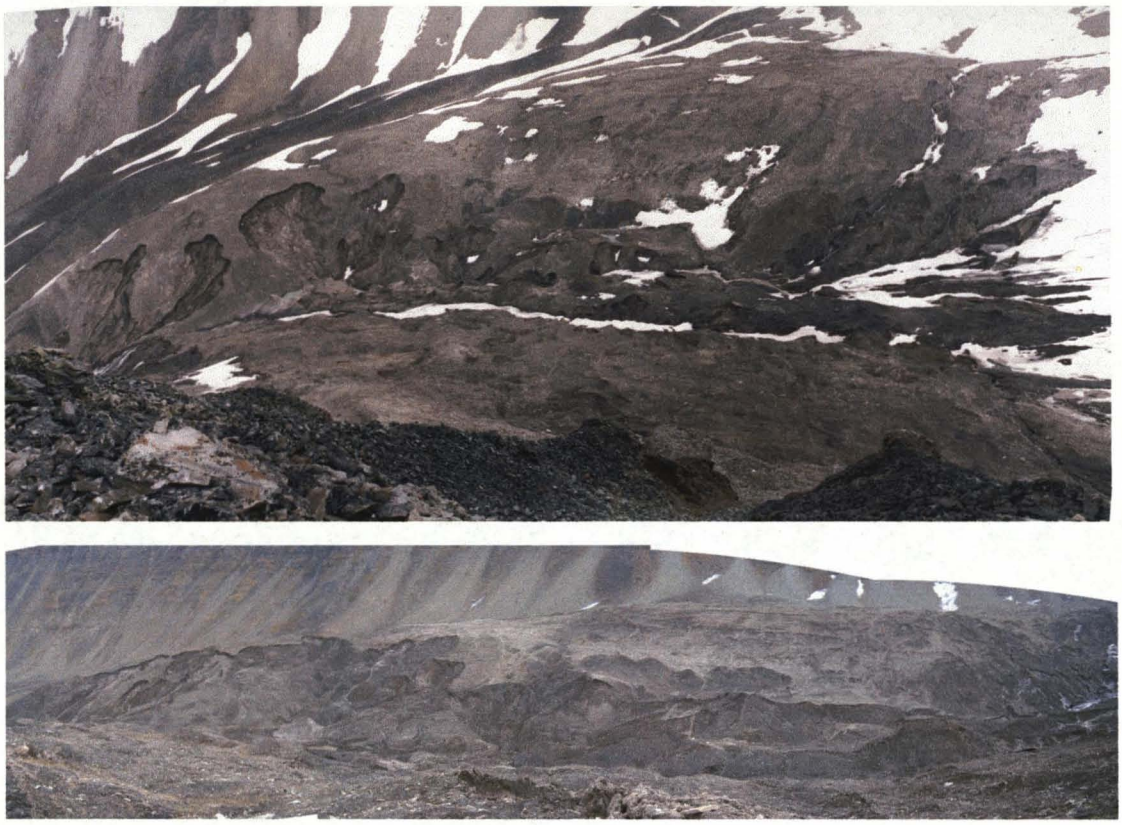


Fig. 5.13 Panorama picture of the eastern side of Larsbreen. The top photograph shows the situation in August 2002 while the bottom that in August 2003. Note the enlargement of the failure complex on the left that has grown together in 2003 and the relatively unaltered part farther away from the central meltwater channel on the right.

The observations made here have implications for long-term landscape evolution. The larger proportion of mobilised material is removed from the system to be stored in outwash plains or the sea, and only a small proportion might survive in what has been termed the "frontal wedge" (Fig. 5.12). Even this thickening "frontal wedge" will degrade with time due to solifluction under continuous permafrost conditions and little constructional evidence of a formerly glaciated terrain will be preserved on timescales ranging from centuries to millennia. These observations add to a growing body of evidence from Svalbard that the preservation potential of glacigenic sediments in a continuous permafrost environment is very limited due to a very efficient coupling of the slopes with the fluvial system via glaciers (e.g. Barsch *et al.*, 1994; Blümel *et al.*, 1994; Etzelmüller, 2000; Etzelmüller *et al.*, 2000; Lyså and Lønne, 2001; Lønne, 2005). This also explains the problems of constructive landform preservation encountered in attempts to reconstruct the history of high-arctic glaciers on Svalbard (e.g. Blümel *et al.*, 1994; Landvik *et al.*, 1998; Lyså and Lønne, 2001; Sletten *et al.*, 2001; Sørbel *et al.*, 2001; Eitel *et al.*, 2002; Lønne, 2005). These findings contrast with clear glacial geomorphological evidence from high-arctic North America where classic glacial landforms have been preserved over several millennia (e.g. Dyke and Evans, 2003). The steeper topography in Svalbard and the smaller

size of the glaciers compared to the large lowland ice sheet lobes in North America might explain the different response of these two systems to reworking.

In addition, the short-lived nature of the LIA advance in central Spitsbergen might have affected the preservation potential of these moraines. The longer a terminus position is sustained, the more debris can be delivered to it by meltout along englacial flowlines. Consequently a longer glacial cycle could lead to a thicker marginal debris cover. This may inhibit the onset of the decay cycle (Fig. 5.12), and in some cases this might aid the survival of ice-cored moraines. A thicker debris cover masking the glacier margin could mean that even a decomposed marginal ice-cored moraine may be more likely to leave an observable marginal landform following de-icing. However, two counterarguments can be made. Firstly, a thicker debris layer that accumulated over a longer time span inhibits melt more effectively than a thinner layer. This reduces the amount of ablation that can occur and, assuming constant climatic conditions, would necessitate further glacier advance in order to maintain a mass balance equilibrium. During an advance the sediment accumulated at the glacier margins would be reworked and redeposited by meltwater and glacial processes. Thus, if an equilibrium situation is re-established, the process of debris accumulation that may lead to ice-cored moraine formation must begin again. Secondly, although a thicker debris cover may inhibit the onset of the positive feedback cycle (Fig. 5.12), the present field observations suggest that thick debris accumulations are as vulnerable to debris remobilisation as thinner ones once the cycle is initiated. This second point also applies to the addition of material by gradual meltout of debris contained in buried ice bodies, which will cause the debris cover to become thicker over time. The preservation potential of remnant features could be increased this way. However, as the decay feedback cycle described in Fig. 5.12 is a positive one, regardless of the thickness of debris, a thicker accumulation of debris on an older feature still does not guarantee its survival as any meltwater or slope failure event could trigger the onset of decay and the inevitable wasting of the buried ice.

5.2.7 Implications for the applicability of the englacial thrusting model

As discussed above, the glaciers studied here contain a number of striated, facetted and thus subglacially transported clasts. Therefore, in agreement with studies elsewhere, temperate basal ice must have been more extensive during the Little Ice Age (Sletten *et al.*, 2001; Lyså and Lønne, 2001). This implies that the glaciers studied here were polythermal at the time of the Little Ice Age and thus not too different from those studied by the proponents of the englacial thrusting model. Processes of degradation during the meltout of buried ice bodies can thus be transferred to the latter sites.

As shown above, where dead ice bodies are present within or underneath sediment bodies, postdepositional reworking will affect the preservation potential of sediments and landforms upon meltout of the buried ice. This issue has never been adequately discussed by the proponents of the englacial thrusting model. Although Hambrey *et al.* (1997: 627) state that "the only evidence of dead

ice is in lateral moraines *and in moraine mound complexes now disconnected from the glacier*" (present author's emphasis) they do not address the question of alteration of the moraine mound complexes during dead ice meltout. Likewise, Bennett *et al.* (1998: caption to Fig. 2) state that "the moraine ridge [...] still contains some buried ice", but they later (p. 23) argue that moraine mounds which were formed from material at the base of the thrusts "incorporate very little, if any, buried ice." Similar arguments have been made by Midgley (2001: 133) and Graham (2002: 119f.).

It is interesting to note that the presence of dead ice is generally acknowledged but then ignored in all the thrusting contributions. In their diagrams, alteration by dead ice meltout is not considered adequately in a theoretical way. To the contrary, only the initial status of the glacier at its neoglacial maximum and *after* retreat is considered, but what happens in between these two extremes is not discussed (Fig. 4.1d-e).

According to the results presented above and general physical principles, as the ice surface lowers, material would melt out successively forming a veneer of supraglacial debris that would partly cover the ice surface and partly slide and flow down the ice surface to accumulate in front of the ice (Fig. 5.14). As thin debris covers retard ablation compared to clean ice, debris ridges and cones can form on the surface that merely mimic structures within the underlying ice (Fig. 5.14b-c; Lukas *et al.*, 2005a, and references therein). A close-up view of an inferred thrust moraine in Hambrey *et al.* (1997: Fig. 4b) where a thin veneer of sediment is draped over a ridge in the ice surface similar to the situation described above for Nordenskiöldtoppenbreen, suggests that this process is currently ongoing at the inferred "thrust moraines". Upon completion of meltout of buried ice, however, constructional landforms, certainly ones with proximal rectilinear slopes, are unlikely to survive unaltered as predicted by the englacial thrusting model (Hambrey *et al.*, 1997; Bennett *et al.*, 1998). It would seem logical to regard the apparently stable thrust moraines reported by the advocates of the thrusting model (Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998; Graham and Midgley, 2000a; Midgley, 2001; Graham, 2002) as a momentarily stable snapshot situation (e.g. Fig. 5.14e) that would change when the protected buried ice was to be exposed. In this case, the processes reported above would lead to widespread obliteration or complete destruction of the constructional "thrust moraines". Instead, a relatively uniform spread of supraglacial material is most likely the resulting end product (Fig. 5.15f).

Thus, on the basis of theory and process observations, the englacial thrusting model appears highly unrealistic and open to question.

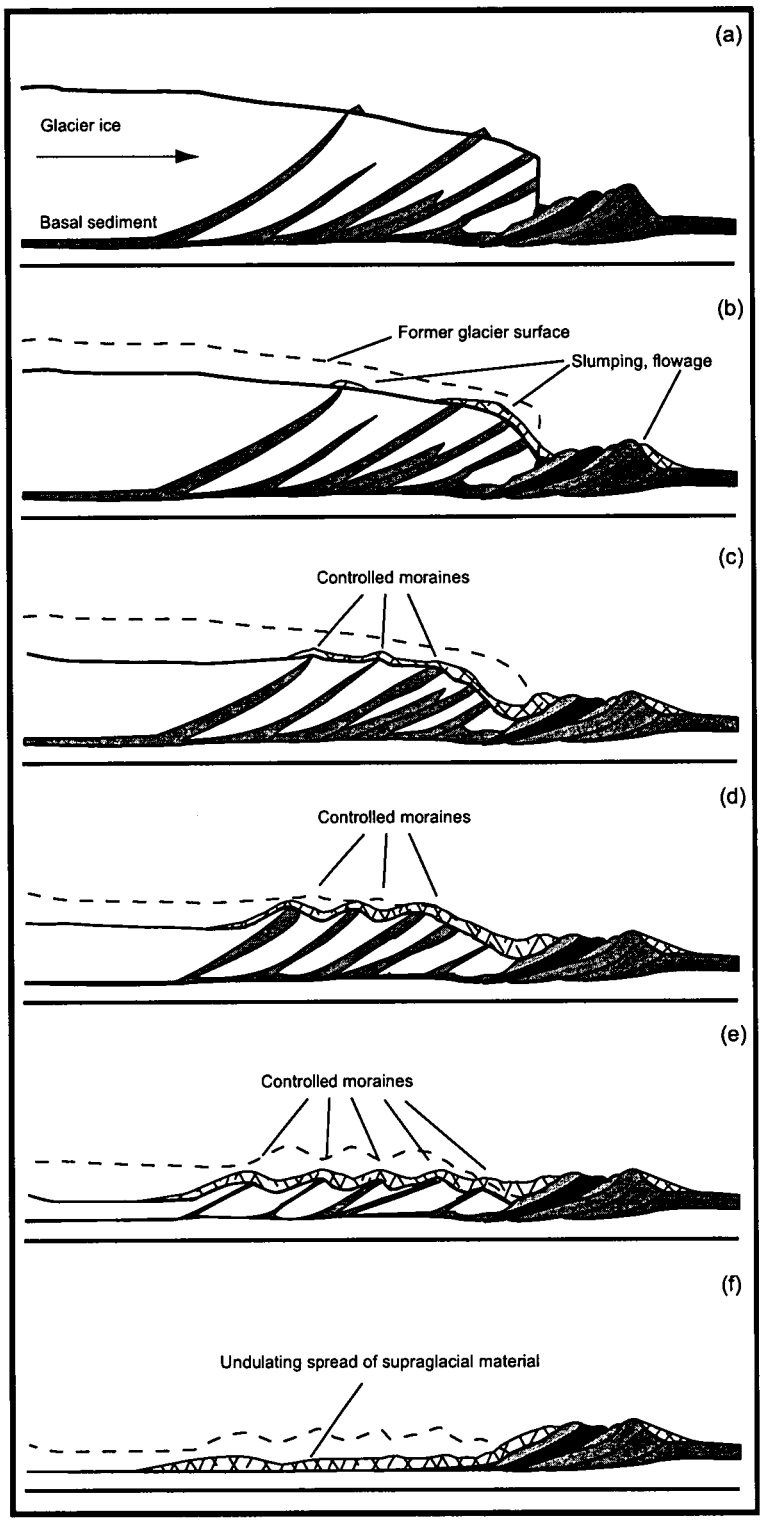


Fig. 5.14 Schematic representation of de-icing progression and its effects on material contained within englacial thrusts (developed from the original diagram of Bennett *et al.*, 1998). For explanation see text.

Apart from these implications that directly result from this study, further criticism can be directed at the englacial thrusting model based on observations at Kongsvegen, Svalbard, used by Hambrey *et al.* (1997) and Bennett *et al.* (1998) to develop the thrusting model. Woodward *et al.* (2002, 2003) re-examined the evidence for thrusting presented by Bennett *et al.* (1998) and Glasser *et*

al. (1998). They found that the criteria used by these authors to identify englacial thrusts were equivocal and that the importance of thrusting in debris entrainment and deposition had probably been overestimated. Englacial debris bands were re-interpreted by Woodward *et al.* (2002, 2003) as former crevasse fills rotated during previous surges. Taking into account sediment-landform associations commonly found at the margins of surging glaciers, Evans and Rea (1999) also concluded that the structures interpreted as thrusts by, for example, Bennett *et al.* (1998) and Glasser *et al.* (1998) more likely originate from the rotation of crevasse-squeeze sediments. These observations cast further doubt on the validity of the englacial thrusting model.

5.2.8 Implications for Svalbard glaciers as analogues for the Scottish Younger Dryas

As discussed above, the most likely geomorphological signature of small high-arctic valley glaciers is meltwater channels and outwash fans radiating out from areas formerly covered by glaciers (cf. Lyså and Lønne, 2001; Sletten *et al.*, 2001; Lønne, 2005), which also corresponds to findings from high-arctic Canada (e.g. Dyke and Evans, 2003, and references therein). Supraglacial landforms such as ice-cored moraines will most likely be modified, obliterated or completely destroyed. Therefore, the cold-based to polythermal landsystem of Svalbard is not reconcilable with the landform record of the far NW Scottish Highlands. The presence of clear recessional moraine sequences, implying short glacier response times and oscillatory retreat, and the absence of evidence for dead ice meltout and stagnation strongly contradict the idea that polythermal Svalbard glaciers form a modern analogue for the Younger Dryas in western Scotland (e.g. Hambrey *et al.*, 1997, 2001; Bennett *et al.*, 1998). Likewise, the absence of widespread evidence for continuous permafrost in western Scotland and the high precipitation values reconstructed across western and central Scotland (Ballantyne, 2002a; Benn and Ballantyne, 2005; Chapters 3.4.2, 3.4.3) do not agree with the idea of Svalbard forming a good modern analogue. However, palaeo-winter temperatures in Scotland during the Younger Dryas (Isarin and Bohncke, 1999; Ballantyne, 2002a; Benn and Ballantyne, 2005) are similar to those observed in Svalbard today (cf. Hambrey *et al.*, 1997) therefore indicating some similarities in boundary conditions. Because Svalbard does not represent a modern analogue, the possibility of a temperate modern analogue will be explored below.

5.3 Temperate analogues – Jostedalsbreen, SW Norway

5.3.1 Study area

Krundalen, an east-west orientated valley, contains three outlet glaciers of the Jostedalsbre plateau ice cap (Bergsetbreen, Tuftebreen and a smaller unnamed glacier) (Fig. 5.15). The regime of all outlet glaciers of Jostedalsbreen is temperate, i.e. the ice is at the pressure melting point throughout (cf. Finsterwalder, 1951; Østrem *et al.*, 1988; Winkler *et al.*, 1997; Winkler and Nesje, 1999). With an

MAAT of 3.8°C (Winkler, 1996a) and a mean annual precipitation that varies between 700 and 2000 mm a^{-1} for individual glacier forelands around Jostedalsbreen (Erikstad and Sollid, 1986), the climate can be classified as maritime and relatively mild. Permafrost is absent from the area, and only shallow ground frost occurs in the winter. Most outlet glaciers reach down to ca. 350 m a.s.l. where they terminate below the treeline (Winkler, 1996a). Outlet glaciers of Jostedalsbreen have been shown to have a high mass turnover, with typical response times between ca. 2-5 years (Nesje *et al.*, 1995; Winkler, 1996b; Winkler *et al.*, 1997), although larger glaciers show response times of up to 35 years (e.g. Winkler, 1996b). As a result of these relatively short response times, outlet glaciers of Jostedalsbreen advanced in the early 1990s triggered by increased winter precipitation during the last years and decades showing that glaciers can be regarded as being very responsive to short-term climate fluctuations and thus close to equilibrium with regional climate (Nesje *et al.*, 1995; Winkler, 1996a, b; Winkler *et al.*, 1997).

Krundalen is underlain by augen-gneisses that are part of the larger Precambrian Jostedal Complex (Sigmond *et al.*, 1984). These rocks are very resistant to erosion and provide limited amounts of rockfall material as a result of which the glacier surfaces are relatively clean (e.g. Winkler, 1996a). The valley sides in Krundalen have responded to glacier retreat following the neoglacial maximum by paraglacial debris cone formation in a number of places (Fig. 5.16; Ballantyne, 1995), and abundant avalanche cones and tracks in forested areas demonstrate high contemporary geomorphological activity. Below, geomorphological and sedimentological evidence from LIA and recent moraines in the forelands of Bergsetbreen and Tuftebreen will be described.

5.3.2 Little Ice Age moraines

5.3.2.1 Geomorphology

Moraines formed by Bergsetbreen and Tuftebreen during and following the Little Ice Age (LIA) maximum (ca. 1779 and 1789, respectively; Bickerton and Matthews, 1993) are clearly identifiable in the upper reaches of Krundalen (Fig. 5.16). These moraines comprise mostly ridges and a few mounds that reach maximum heights of ca. 3 m, widths of 50 m and lengths of up to 500 m. They are generally densely spaced and have arcuate planforms that are open in an upvalley direction. Moraines are generally absent from the central part of the valley floor due to proglacial fluvial activity. Individual moraine ridges are commonly fragmented by small meltwater channels. Many moraines exhibit crestline bifurcations, and crestlines have an undulating appearance in places. Subangular to subrounded, faceted boulders of local augen gneiss are widespread along moraine crestlines and occasionally occur as distinct boulder lines on flat areas of outwash between two moraine fragments, effectively connecting them.

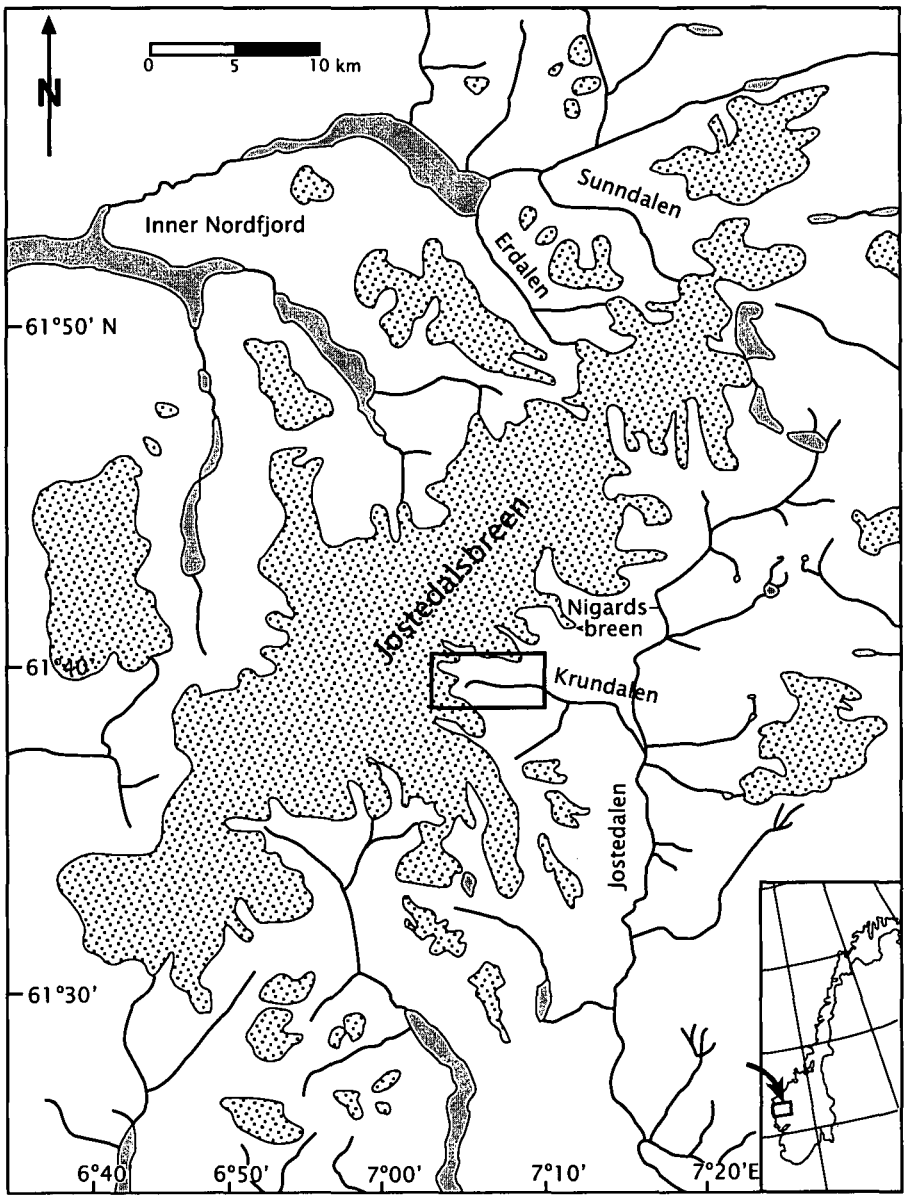


Fig. 5.15 Overview map of Norway showing the location of the study area. The frame marks the location of Fig. 5.16.

The planform of the discrete moraine fragments and the frequent alignment of boulders reflects the shape of the ice margin at the time of deposition. Crestline bifurcations indicate that the glaciers retreated in an oscillatory fashion allowing the moraines to be interpreted as recessional moraines. Such an interpretation is consistent with lichenometric dating studies that have shown that these moraines formed during glacier retreat from the maximum LIA position (Bickerton and Matthews, 1993). Comparison of dated moraine sequences from different valleys around Jostedalsbreen indicates that many moraines were formed synchronously, at intervals of between 2-5 years (Bickerton and Matthews, 1993; Winkler *et al.*, 1997).

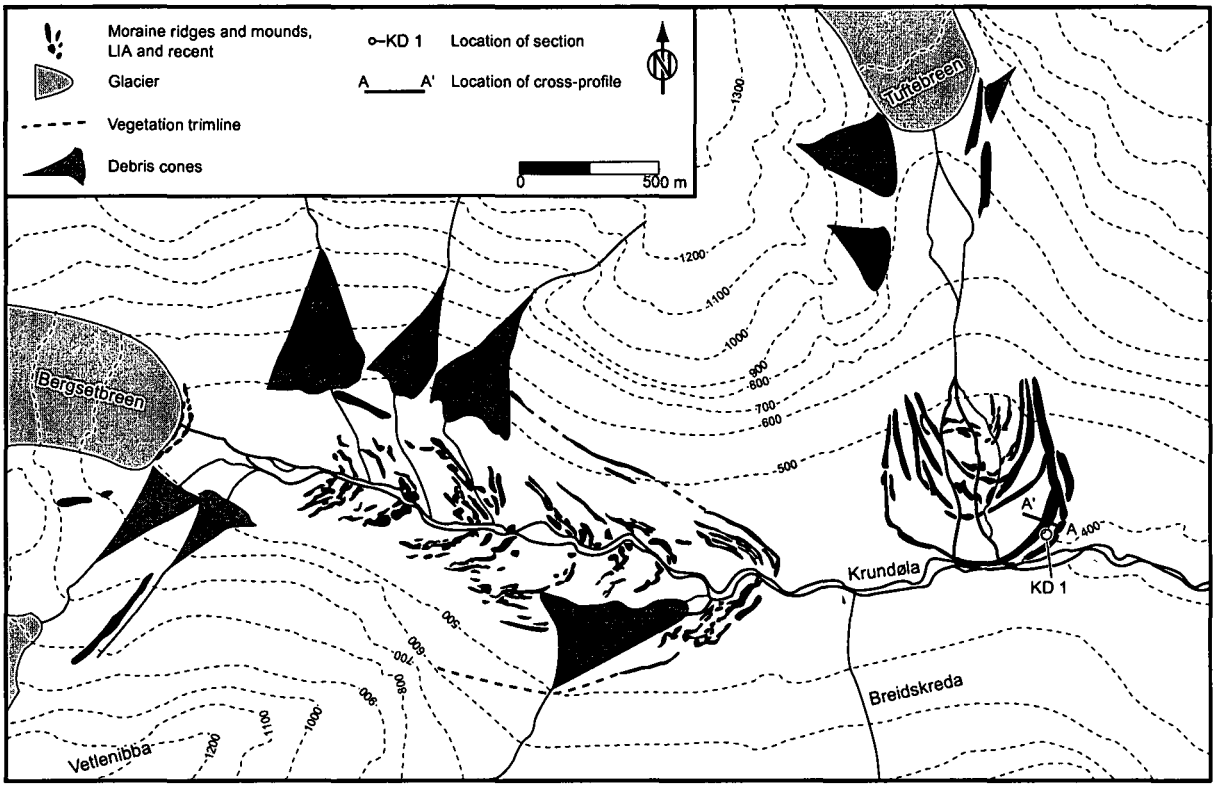


Fig. 5.16 Geomorphological map of study area in Krundalen, SW Norway showing the location of moraines, site KD 1 and the cross-profile shown in Fig. 5.17a.

5.3.2.2 Sedimentology

Most moraines are composed of mixtures of diamictic and sorted sediment facies as apparent from shallow surface exposures. However, some moraines are largely composed of stacked openwork boulders with maximum a-axis lengths of 3 m. One section from an inter-moraine area close to the Little Ice Age maximum of Tuftebreen will be described below. At this site, the outermost LIA moraine ridge is separated from the next by a gently sloping, planar surface. An exposure of ca. 0.9 m depth and 2 m width was excavated parallel to the crestlines of these two moraines with additional pits of ca. 0.4 m width perpendicular to the crest, i.e. parallel to the former ice flow direction (KD 1; Figs. 5.16, 5.17).

The frontal view parallel to the crestline reveals alternating subhorizontal layers of four lithofacies units: (A) alternating layers of openwork fine-coarse gravel with a massive appearance and massive silty fine sand with scattered medium and coarse sand and fine gravel particles (Sm/GRm) in the uppermost 0.5 m of the main pit; (B) horizontally-bedded fine sand to medium gravel (Sh), often forming channel- and half-channel-structures or lenses; (C) well-sorted very fine to fine sand exhibiting wavy bedding (Sl), often intercalated with (B); and (D) silt to very fine sand, displaying a wavy bedding particularly emphasised by darker bands of silt (Fl/Sl). All five units are characterised by sharp upper and lower contacts. A large subangular boulder of augen gneiss (0.8 m a-axis) occurs in the middle part of the section (Fig. 5.17b).

The exposures in the pit walls perpendicular to the crestline (i.e. parallel to glacier flow) show the same facies, but display numerous deformation structures. While the units in the left hand (southern) side of the exposure only show moderate signs of deformation, the right hand (northern) side displays a system of reverse faults in Unit (D) (Fig. 5.17c, d). Part of a lens of (B) extends upwards into the lower void. Displacement along individual faults ranges from 2 to 6 cm. Detailed structural measurements of fault plane orientation were not possible due to the high gravel content of the sections which frequently caused collapse of these delicate structures.

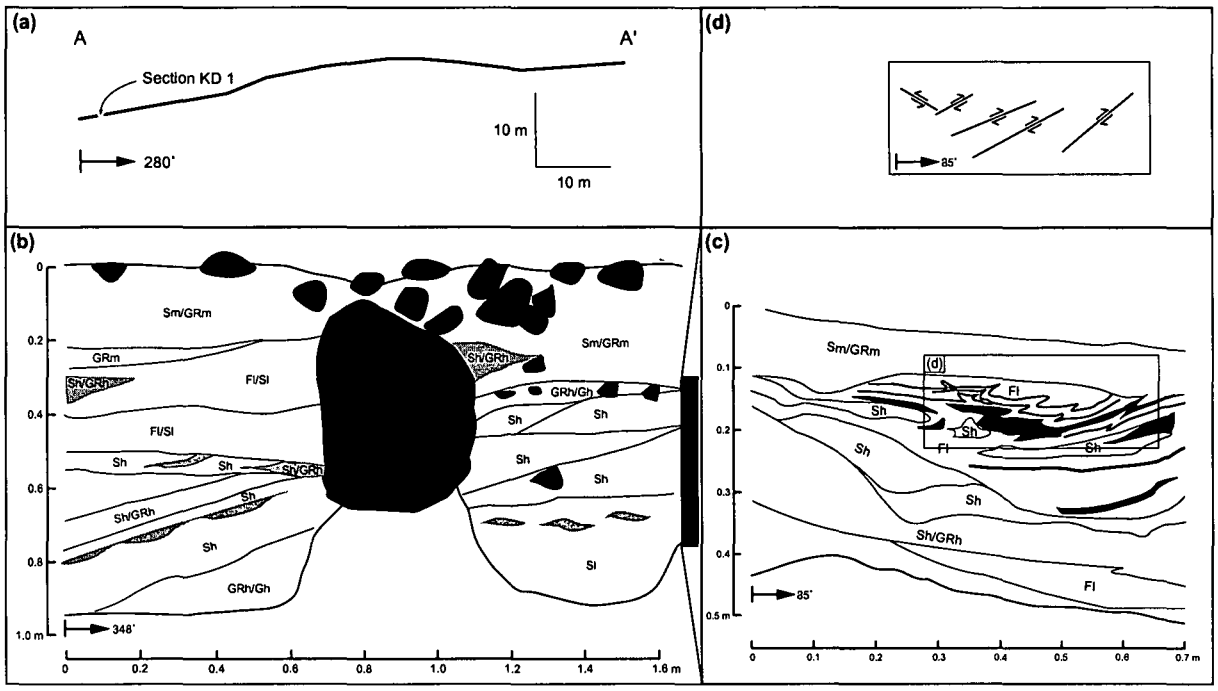


Fig. 5.17 (a) Cross-profile across moraine ridge; (b) sedimentary log of main section KD 1; (c) close-up of lithofacies units and deformation structures in right-hand wall perpendicular to (b). Thick black lines and envelopes depict darker silt horizons that visualise the deformation structures. Box shows the location of (d); (d) nature and orientation of fault lines as depicted in (c).

Together, the lithofacies are interpreted as proximal outwash (Fig. 5.17b). Differences in grain-size fractions of individual units reflect varying melt rates, meltwater input and transport capacity on diurnal and seasonal cycles.

Deformation structures in the finer-grained sediment facies demonstrate that the style of deformation is undoubtedly ice-marginal and of a laterally-compressional nature. This can be deduced from the presence of several *en echelon* reverse faults (Fig. 5.17c, d; cf. Benn and Evans, 1993; Benn and Clapperton, 2000; van der Wateren *et al.*, 2000; McCarroll and Rijsdijk, 2003; Lukas and Merritt, 2004). The orientation of the faults conforms with a push direction from the WNW, i.e. parallel to the exposed section wall in the right pit (Fig. 5.17c, d). Drawn together, the evidence indicates a glacier readvance into a recently-formed small ice-contact outwash fan causing the formation of both brittle and ductile deformation structures. It is likely that the readvance took place during, or shortly prior to,

formation of the subsequent moraine located to the NW, possibly in the following winter and spring. This evidence is in good agreement with independent geomorphological, glaciological and climatological data from glaciers around the Jostedalsbre ice cap that highlight the short marginal response times of individual outlet glaciers (Winkler *et al.*, 1997) and the frequent occurrence of minor winter readvances during overall retreat in years of cooler summers and/or increased snow supply in winter (cf. Bickerton and Matthews, 1993). Evidence for meltout of dead ice pockets or more substantial ice cores has not been found. Indeed, such presence can be ruled out due to the climatic regime and the preservation of delicate deformation structures that would otherwise have been modified or destroyed (cf. Kjær and Krüger, 2001).

5.3.3 Ice-marginal moraines at the front of Bergsetbreen

The immediate ice margin of Bergsetbreen was visited in early June 2003 and shows a series of discrete, sharp-crested, boulder-rich moraines that are between 3 and 5 m high, up to 5 m wide, 15 m long resting on scoured gneiss bedrock and a thin veneer of outwash sediments (Fig. 5.18). Moraines are absent from the central part of the valley, where glaciofluvial reworking processes predominate. Traces of outwash deposits can also be found at gaps in the moraines, indicating the former position of meltwater streams. The moraine ridges recently formed at the margin of Bergsetbreen show a marked asymmetry with rectilinear slopes on either side. Slope angles vary between 30-37° for proximal and 22-28° for distal slopes.

Although no exposures were available in these moraines, an alternation of finer sorted sediments and boulder-rich layers can be identified (Fig. 5.18). However, in contrast to the gently sloping proximal outwash fan in the LIA moraines in front of Tuftebreen, the distal side dips away from the glacier at about 22° while the proximal side is much steeper, at ca. 35°, interpreted as a clear indication for material at its angle of repose. The close proximity of the moraines to the glacier, their size and relationship to currently forming outwash fans suggests that they began to form during summer when sediments were deposited on the bedrock during summer retreat. These were then pushed up during the following winter to form the discrete arc of moraines observed in June 2003. Such an interpretation is supported by the steep gradient of the glacier snout in 2003 indicating an active, advancing front.

Fig. 5.18 Photograph showing the ice-marginal moraines formed during winter readvances by a combination of pushing and dumping at the present ice margin of Bergsetbreen, which is just off the right margin of the photograph. Group of students in foreground for scale.

5.3.4 Synthesis of evidence from Norway

Geomorphological and sedimentological evidence demonstrates that the temperate maritime glacial landsystem of SW Norway is characterised by glaciers that respond in a highly active way to short-term climate forcing. Recessional moraine sequences during the LIA were formed during net retreat interrupted by smaller readvances, possibly triggered by a combination of a seasonal signature superimposed on longer-term (decadal) mass balance changes. Such readvances appear to have caused sediment deformation in small ice-contact outwash fans (LIA) that were formed prior to the readvance and the formation of small push and dump moraines (recent). This evidence is consistent with detailed investigations of the coupling between climate and glaciers that demonstrate short marginal response times (between 2 and 35 years) for glaciers of the Jostedalsbre ice cap (Bickerton and Matthews, 1993; Nesje *et al.*, 1995; Winkler, 1996a, b; Winkler *et al.*, 1997) and thus a highly dynamic marginal environment. Due to the limited to absent incorporation of dead ice during moraine formation, the latter are not altered postdepositionally due to meltout and thus have a relatively high preservation potential.

5.3.5 Implications for Norwegian temperate valley glaciers as modern analogues

Maritime Norwegian temperate valley glaciers for which two examples have been presented above share a number of characteristics of the Scottish Younger Dryas glaciers in the far NW Highlands. Most importantly recessional moraine sequences reflect the shape of the former ice margin. Lichenometric dating of LIA moraines and modern mass balance observations demonstrate that the

glaciers have very short response times, which is also similar to the inferences made for Scotland on the basis of moraine spacing and frequency (Chapter 4.4.2). Such short response times also agree with the high modern precipitation values in SW Norway which are in a similar range to those during the Younger Dryas in NW Scotland (Chapter 3.4.2). The dominant mode of transport appears to be largely subglacial as inferred from clast shape observations and is thus similar to that in the Scottish Younger Dryas palaeo-environment. However, the size of individual moraine ridges and the modes of deposition do not agree with those in Scotland. Younger Dryas moraines in the far NW Highlands are up to 5 times the size of Norwegian LIA moraines, and the Norwegian moraines do not consist of terrestrial ice-contact fans derived from the deposition of supraglacial material. To the contrary, the latter glaciers are largely clean due to the resistant nature of the bedrock. Despite these contrasts, both environments contain unequivocal evidence of two-phased moraine formation during oscillatory retreat. While the temperate Norwegian glaciers cannot explain every aspect of the Younger Dryas glacial landsystem in the far NW Scottish Highlands, it is a much closer match compared to the cold-based to polythermal glaciers on Svalbard discussed above. Since neither landsystem can adequately explain the characteristics of the Scottish palaeo-environment in the far NW Highlands, additional landsystems will be evaluated below.

5.4 High-relief, debris-covered glaciers as modern analogues

In addition to the above evaluation of temperate and cold-based valley glacier landsystems as modern analogues, this section will explore to what extent debris-covered glaciers can serve as modern analogues to Younger Dryas glaciers in the NW Highlands. This is necessitated by the fact that the moraine sedimentology implies deposition by subaerial gravitational and glaciofluvial processes from a supraglacial position, analogies with which exist at debris-covered glaciers in high-relief settings (e.g. Boulton and Eyles, 1979; Small, 1983, 1987; Owen and Derbyshire, 1989; Benn and Owen, 2002; Benn *et al.*, 2003). In such environments, the distal parts of latero-frontal moraines typically expose alternating debris flow and sorted sediment units that dip away from the ice margin, whereas proximal parts are prone to collapse and reworking following ice withdrawal and removal of support. Terrestrial ice-contact fans and ramps formed in this way closely resemble parts of the glacial landsystem during the Younger Dryas in the study area of NW Scotland, albeit on a much larger scale. However, there are a number of important differences between the two settings. Firstly, debris-covered glaciers receive much of their material by rockfall and as debris transferred in snow avalanches (Shroder *et al.*, 2000; Benn and Owen, 2002; Benn *et al.*, 2003). Secondly, the debris cover typically inhibits marginal responses to climate fluctuations, so that many debris-covered glaciers respond to mass balance changes by thickening and thinning rather than marginal advance or retreat (e.g. Benn and Evans, 1998; Kirkbride and Warren, 1999; Benn and Owen, 2002; Benn *et al.*, 2003). The latter is

reflected in much larger fans with single or multiple crestlines that indicate long-term marginal stability (e.g. Boulton and Eyles, 1979; Small, 1987).

These factors differ markedly from the Scottish Younger Dryas landsystem in the far NW Highlands since the sediment is transported subglacially and then elevated to the surface from where it is deposited; a supraglacial sediment source can be ruled out (Chapter 4.3.2.4). Additionally, Younger Dryas glaciers in NW Scotland were highly responsive to climate fluctuations as evident from the large number of closely spaced recessional moraines and sedimentological evidence for readvances (Chapters 4.3.1.7, 4.4.2).

Thus, although the depositional processes at modern debris-covered glacier margins are very similar to those during the Younger Dryas in Scotland, the dynamics of these glaciers are very different. Therefore, apart from the sedimentary end product of terrestrial ice-contact fans or ramps, modern debris-covered glaciers cannot explain most features of the Scottish Younger Dryas glacial landsystem in the far NW Highlands.

Since an attempted comparison of three potentially suitable modern analogues has so far been unsuccessful, individual characteristics and their implications for modern analogues will be reviewed below.

5.5 Comparison with Scottish evidence and discussion

Chapter 4 introduced the characteristics of the Scottish Younger Dryas palaeo-environment in the far NW Highlands that can be reconstructed with confidence. Based on these data, boundary conditions that need to be met in order to explain the Younger Dryas landsystem in the far NW Highlands will be reviewed here. The aims of this review are: (a) to explore whether individual processes of moraine formation inferred from the field evidence are diagnostic of particular climatic or glaciological settings and (b) to ask whether the observed combination of inferred processes occurs in any one modern environment.

A comprehensive model of the formation and glaciological significance of "hummocky moraine" must be capable of explaining all of the evidence, including:

- the large volume of debris represented by suites of "hummocky moraine",
- how this debris was transported and elevated to the former glacier surfaces,
- the range of inferred depositional and deformational processes,
- the distinctive morphology of "hummocky moraine" and
- the highly dynamic, oscillatory behaviour of Younger Dryas outlet glaciers in the NW Highlands.

Furthermore, such a model must also be compatible with palaeoclimatic reconstructions based on palaeo-precipitation estimates and independent data obtained from faunal assemblages and other types of proxy data.

5.5.1 Debris provenance, transport and elevation

The work presented here has shown that the material making up closely spaced "hummocky" recessional moraines in the far NW Highlands is dominantly subglacially derived and transported (Chapter 4.3.2.4). Both these aspects appear to be very typical of the Younger Dryas glacial landsystem along the Scottish west coast since they correspond well with earlier findings on debris provenance, moraine morphology and spacing (Hodgson, 1982, 1986; Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Bennett and Boulton, 1993a; Wilson and Evans, 2000; Lukas, 2003). Earlier suggestions by Eyles (1983a) that the debris in "hummocky moraines" might be derived from supraglacial sources have not been confirmed so far, and his arguments remain unsupported by detailed provenance studies. Likewise, as discussed above (Chapter 4.4.2), the reasoning by Ballantyne (2002a, b) that the large volume of debris in "hummocky moraine" could reflect the overriding and entrainment of paraglacial slope deposits produced during and after ice sheet deglaciation is unsubstantiated by moraine morphology and clast provenance data.

Together with the ubiquitous occurrence of roches moutonnées, striated and ice-moulded bedrock and subglacial flutes in the accumulation areas the evidence strongly suggests a temperate glacier regime (Chapter 4.4.2). This inference is based on the fact that basal sliding is required to create subglacial erosional and depositional landforms and to transport large quantities of material subglacially. The location of fluted moraines in the glacier source areas and the occurrence of overridden and drumlinised moraines near the maximum extent along Loch Shin indicate that temperate ice extended close to the margins of Younger Dryas glaciers. Similar fluted moraines and drumlinised ice-marginal moraines have been observed on the forelands of modern temperate glaciers (e.g. Boulton, 1978, 1987; Krüger and Thomsen, 1984; Benn, 1992a, 1994, 1995; Benn and Evans, 1998; Evans and Twigg, 2002; Glasser and Bennett, 2004).

It is evident from the clast morphological data and section logs that subglacially-transported sediment was elevated at least to the height of moraine crests (commonly > 10 m) prior to deposition at the glacier margins. A number of mechanisms that elevate debris near the margin are recognised in the literature (e.g. Alley *et al.*, 1997). These include regelation, freeze-on by hydraulic supercooling, freeze-on by conductive cooling and marginal folding and thrusting.

Debris entrainment by regelation can occur in the lee of bedrock bumps, or at the interface between ice and unfrozen subglacial sediment. Alley *et al.* (1997) state that this process is unlikely to be significant under temperate glaciers due to inefficient heat conduction through larger bedrock obstacles such as those that commonly occur in the valley bottoms in the far NW Highlands. Freeze-

on by hydraulic supercooling occurs when subglacial water is elevated out of basal overdeepenings, depressurises and freezes. While such overdeepenings make up parts of the valleys in the far NW Highlands, areas of "hummocky moraine" occur irrespective of their location, and hydraulic supercooling cannot account for the large amounts of debris being elevated to the glacier surface in the majority of cases. In addition, the clast shape data indicate that glaciofluvial transport was not significant, and this should be the case if frozen supercooled water was the principal agent of debris entrainment (cf. Spedding and Evans, 2002).

Conductive freeze-on is encouraged by transitions from warm-based to cold-based ice near the margins of polythermal glaciers and can lead to the formation of thick sequences of basal debris near glacier margins (e.g. Weertman, 1961; Knight, 1997). These can be subjected to thickening by folding near the glacier margin. As discussed earlier (Chapters 1.1.2, 1.1.3.2) debris elevation by thrusting and folding has been reported to be a very widespread process at the margins of modern polythermal glaciers in permafrost environments, where the transition from warm- to cold-based ice near the margins leads to flow compression and englacial thrust formation (Glasser and Hambrey, 2003). In addition, it has been suggested that temperate glaciers advancing against reverse bedrock slopes might also experience flow compression near the margins, resulting in the development of englacial thrusts (e.g. Hambrey *et al.*, 1997; Bennett *et al.*, 1998). The latter case, however, cannot explain the occurrence of most areas of "hummocky moraine" since steep reverse bedrock slopes are only encountered in relatively few locations in the present study area. Although it has been shown that englacial thrusting is not responsible for the formation of "hummocky moraines" (Chapters 4.4.3, 5.2.7), it cannot be ruled out as a process that elevated subglacially-transported material to the glacier surface from which it would then have been deposited by gravitational and glaciofluvial processes.

To conclude, the geomorphological and sedimentological evidence in the NW Scottish Highlands indicates that Younger Dryas glaciers were predominantly temperate. However, to continuously elevate subglacially-transported debris to the glacier surface a mechanism is required that can occur irrespective of special situations such as the presence of overdeepenings or steep reverse bedrock slopes (external controls). Thus, a glaciological factor (internal control) is most likely responsible for an efficient elevation of material to the glacier surface close to the margins as this would "migrate" with the glacier, allowing moraines to be deposited regardless of topographic boundary conditions. Therefore it is argued here that only a very narrow zone near the margins of Younger Dryas glaciers in the NW Highlands was cold-based, facilitating folding and thrusting and thus elevation of basal debris to the glacier surface. The reasoning for a narrow zone of cold-based ice is further strengthened by (a) the widespread evidence for a temperate thermal regime and (b) the lack of evidence for dead ice meltout. The latter process would necessarily have to occur if the transition between warm- and cold -based ice was farther upglacier since cut-off from supply upglacier and local stagnation of a wide frozen margin would be encouraged (cf. Glasser and Hambrey, 2003).

5.5.2 Depositional processes

As described above (Chapter 5.4), the sediments of which "hummocky moraines" in NW Scotland are composed have their closest modern analogues in high-relief glaciated valleys where sediment is deposited by gravitational and fluvial processes around debris covered glacier margins (e.g. Boulton and Eyles, 1979; Small, 1983, 1987; Owen and Derbyshire, 1989; Benn and Owen, 2002; Benn *et al.*, 2003). However, as discussed above, the size of ice-marginal fans in both settings is different, and the response of modern debris-covered glaciers to climate fluctuations cannot account for the oscillatory retreat patterns and rapid marginal fluctuations recognised in Scottish Younger Dryas glaciers in the NW Highlands.

The evidence for significant ice marginal oscillations in "hummocky moraine", combined with very limited evidence for ice stagnation, relatively high retreat velocities and short response times of Younger Dryas glaciers in NW Scotland (Chapter 4.4.2) indicates that the glacier margins did not support extensive debris covers. If this had been the case glacier response to short-term climate forcing would have been significantly dampened and recessional moraines would not have been formed. Furthermore, an important difference between modern debris-covered glaciers and Younger Dryas glaciers in the study area is the dominantly supraglacial as opposed to subglacial sediment delivery to the glacier surface in the former (Benn and Owen, 2002). Together with the evidence for dominant subglacial transport and absence of dead ice meltout reviewed above, this suggests elevation of basal debris to the surface only very close to the terminus. Deposition from a supraglacial position by gravitational and glaciofluvial processes must have occurred fairly rapidly in order to prevent extensive debris covers from forming.

5.5.3 Moraine morphology

The hummocky appearance of moraines in Scotland is very distinctive and requires explanation. First, glacial meltwater emerging from glacier portals either prevented moraines from being formed there or postdepositionally breached formerly deposited moraines and eroded parts of them (e.g. Bickerton and Matthews, 1993; Lukas, 2003, 2004b; Chapter 5.3). Second, lateral meltwater channels crossing moraine arcs indicate that formerly continuous moraines might have been dissected during retreat by meltwater. Third, on hillsides snowmelt could have contributed to dissection and gully initiation (cf. Lukas, 2003). However, a fourth possibility, namely that moraine arcs were not continuous at the time of deposition but were deposited in discrete positions along the ice margins has to be taken into account as well. This would require discrete point sources of supraglacial material such as medial moraines or the meltout of localised debris concentrations (e.g. Eyles, 1979; Krüger and Aber, 1999; D.I. Benn, pers. comm., 2005). In reality, a combination of these factors is likely to be responsible for Scottish "hummocky moraines" being as fragmented and pointed as they are, but the large number of

meltwater channels surrounding, and cutting through, moraines suggests that glacial meltwater played a more important role than discrete foci of deposition.

5.5.4 Ice margin dynamics

The planform patterns and dense spacing of "hummocky moraines" in the NW Highlands closely resembles those formed by modern temperate glaciers as shown in Fig. 5.16 for Tuftebreen and Bergsetbreen (Chapter 5.3.2.1). The latter moraines form discontinuous arcs that have been locally dissected by outwash and are separated by abandoned glaciofluvial fan and terrace surfaces indicating breaching and reworking of formerly continuous ridges. Moraines are between tens to hundreds of metres apart and commonly exhibit crestline bifurcations. In common with recessional moraines in other temperate environments, those in Krundalen mark the limit of minor winter advances and more substantial readvances in response to multi-annual climatic fluctuations (e.g. Boulton, 1986; Bickerton and Matthews, 1993; Krüger, 1994, 1995; Winkler, 1996a; Winkler and Nesje, 1999; Evans and Twigg, 2002). The short glacier response times recognised in this modern temperate landsystems is thus very closely compatible with the data for the NW Highlands that also indicate highly dynamic ice marginal responses and frequent oscillations (Chapter 4.4.2).

5.5.5 Climatic setting and conclusion

Based on ELA estimates and mean July temperatures during the Younger Dryas based on chironomids (Brooks and Birks, 2000), the palaeoclimatic conditions during the Younger Dryas in the far NW Highlands could be reconstructed (Chapter 3.4). According to these results, annual precipitation totals of 2358±337 mm a^{-1} at sea-level were ca. 26% higher than at present. These results agree well with the evidence for highly active glaciers with short response times that retreated in an oscillatory fashion. Numerical modelling has also shown that the winter temperatures were <–25°C (Isarin and Renssen, 1999), representing a severely arctic climate in which continuous permafrost could be expected (e.g. Karte, 1979; French, 1996). However, the available evidence suggests that permafrost was restricted to higher ground and was not widespread near sea-level during the Younger Dryas along the Scottish west coast (Ballantyne and Harris, 1994). Furthermore, glaciers in a permafrost environment do not respond as dynamically to climate change as those in maritime, non-permafrost environments (Chapters 5.2, 5.3).

According to the presently available evidence of palaeo-temperature and precipitation values and the geomorphological and sedimentological evidence from the NW Highlands (and elsewhere) it can be argued that relatively warm summers and high amounts of snowfall would have suppressed the development of deep continuous permafrost by insulating the ground surface (French, 1996). At the same time, high precipitation would have caused high mass turnover in glaciers, thereby encouraging fast flow and basal melting (Benn and Evans, 1998), also explaining the short response times and

frequent ice-marginal oscillations recognised in Younger Dryas glaciers in the far NW Highlands (Chapter 4.4.2). High precipitation totals would also have resulted in relatively narrow zones around the margin being frozen rather than widespread cold-based conditions at the bed as observed for smaller and less active glaciers in Svalbard. The relatively fast ice flow would also have encouraged rapid advection of debris to the margins, from where it would have been evacuated relatively quickly so that a continuous, thick debris cover would have been prevented from forming and insulating the margins.

To conclude, it would appear that no one modern analogue is able to explain the characteristics recognised in "hummocky moraine" in the far NW Highlands, but also elsewhere along the west coast. Presently available evidence suggests that the high palaeo-precipitation totals provide the "missing link" that connects characteristics of modern temperate, polythermal and debris-covered glaciers in the glaciated valley landsystem (Benn *et al.*, 2003) to a unique combination that lacks a single modern analogue.

CHAPTER 6 SUMMARY AND IMPLICATIONS

6.1 Introduction

This chapter summarises the main findings reported in the present thesis and discusses issues arising from these. This will be done in five separate sections. First, the implications of the work presented here for the understanding of the extent of glaciers and the palaeoclimatic conditions during the Younger Dryas chronozone in the far NW Highlands of Scotland will be summarised. Second, the genetic processes identified from detailed geomorphological and sedimentological analyses of "hummocky moraine" and the implications of these results for Younger Dryas glacier dynamics in NW Scotland and hitherto-proposed conflicting models of moraine formation will be evaluated. Third, the results obtained from two modern environments will be summarised and discussed in the light of potential modern analogues for Younger Dryas glaciers in Scotland. Fourth, more general methodological implications of the above results will be discussed and fifth, areas where future research is needed to answer questions that result from the work presented here will be identified.

6.2 Younger Dryas glaciation in the far NW Highlands and palaeoclimatic significance

6.2.1 Distribution of Younger Dryas glaciers in NW Scotland

Detailed mapping of the glacial geology and geomorphology both in the field and from aerial photographs has led to the reconstruction of a coherent ice mass in the far NW Highlands. A sequence of radiocarbon dates of organic sediments recovered from an open lake basin just outside an end moraine, the lowermost of which is 11,230 to 10,680 cal a BP, allows correlation of this limit with the Younger Dryas (ca. 12.7-11.5 cal ka BP). This evidence is compatible with previous work in NW Scotland from the surrounding areas (Pennington, 1977), and ice limits were correlated using a morphostratigraphical approach that utilises clear landsystem contrasts. A coeval transection glacier complex ("ice cap") occupying an area of 211 km^2 in the central mountains of Sutherland and a smaller corrie glacier complex of 2 km^2 could thus be reconstructed based on a combination of numerical dating and detailed geomorphological mapping. The southern and western limits of the transection glacier complex lay outside the present study area, however, but will be added once mapping by the British Geological Survey is complete.

6.2.2 Palaeoclimatic implications

Reconstructions of the equilibrium-line altitudes (ELAs) using Area-Altitude Balance Ratio (AABR), Accumulation Area Ratio (AAR) and area-weighted mean altitude approaches has resulted in a data set which can be compared with previous studies across Scotland. Assuming balance ratios of 2.0

(AABR approach) this yields values of 324 m for the ice cap and 393 m for the corrie glacier complex. Using techniques to calculate palaeo-precipitation totals established elsewhere in Scotland (Benn and Ballantyne, 2005) yields values of 2358±337 mm a^{-1} at sea-level which are ca. 26% higher than present-day values of 1863 mm a^{-1}. The Younger Dryas palaeoenvironment in the far NW Highlands was thus wetter than today. This was even more so farther south on the Isle of Mull, indicating that a north-south gradient might have been present during the Younger Dryas. This difference might be due to less maritime conditions experienced by the central Sutherland mountains compared to the coastal areas of Mull, possibly due to scavenging of precipitation by coastal mountains.

The reconstructed ice cap in the NW Highlands is almost six times larger than hitherto assumed (Sissons, 1977), and the distribution of Younger Dryas ice masses across Scotland incorporating the data presented in this thesis is shown in Fig. 6.1. The presence of a much larger ice cap in the NW Highlands removes the previous discrepancy between comparatively small glaciers in the north and the much larger Western Highland Ice Cap in the south (Fig. 1.1). Likewise, special atmospheric conditions adduced to explain this anomaly (Sissons, 1977, 1980) do not have to be invoked. Instead, the distribution of ELAs across the ice cap and the very humid conditions confirm earlier work that indicates a westerly to southwesterly palaeo-windfield and high precipitation totals (Ballantyne, 1989, 2002a; Christiansen, 2004; Benn and Ballantyne, 2005). These suggestions and calculated palaeoclimatic variables have to be regarded as preliminary, however, since the evidence presented here needs to be extended to include data from the southern and western continuation of the ice cap.

6.3 Morphology, sedimentology and significance of "hummocky moraine"

6.3.1 Moraine types

The present study was designed to map the distribution of moraines in the NW Highlands, to investigate the processes leading to their formation and to test which, if any, of two hitherto-proposed models of moraine formation are applicable. Three types of moraines have been identified in the study area: (1) lateral moraines, (2) fluted moraines and (3) "hummocky moraine".

Lateral moraines consist of sedimentary benches and ridges that trend obliquely across the hillsides of glacial troughs at low angles. Combining their crestlines and surface trend yields clear lines with a uniform gradient that are interpreted as the former ice margin. These moraines are very useful in glacier reconstructions, especially in the lower parts of the ablation area where other information to constrain the ice surface are largely lacking.

Fluted moraines are linear ridges with a streamlined appearance that occur on the downglacier side of obstacles such as bedrock steps and corrie headwalls and are common in the accumulation area of most glaciers. Their presence indicates widespread basal sliding and flow acceleration in the

accumulation areas, strongly indicating that the ice was dominantly temperate. They also indicate that the remoulding of thick basal sediment was an important process in Younger Dryas glaciated valley landsystems in the far NW Highlands.

"Hummocky moraine" consists of discrete chains of moraine ridges and mounds that trend obliquely across the hillside and resemble arcuate shapes. Crestline bifurcations are frequent, and the moraines are closely associated with meltwater channels that either breach moraines or run parallel to them. Most moraines show a clear asymmetry with steep proximal rectilinear and gentler distal rectilinear slopes. Surveys of moraine morphology show that proximal rectilinear slopes with angles around 30° are a typical characteristic of "hummocky moraine", making up 86.4% of 118 moraines measured.

6.3.2 Genesis of "hummocky moraine"

Logging of 52 exposures in "hummocky moraines" (Appendix 1) allows their genesis to be reconstructed accurately and enables statements about the relative frequency and importance of individual processes to be made. Sedimentological logging reveals that the majority of these moraines represent terrestrial ice-contact fans, i.e. landforms produced by the stacking of supraglacially derived debris flows and fluvial sediments deposited in shallow rills or thin sheetflows at a temporarily stationary ice margin (Chapter 4.3.1). Proximal rectilinear slopes correspond to ice-contact faces that formed as the glacier retreated and withdrew its support. Distal rectilinear slopes correspond to the fan surface. A continuum of deformation structures, ranging from folds and reverse faults indicative of proglacial lateral compression to boudins and augen-like structures indicative of subglacial simple shear, has been recognised in these fans (Chapter 4.3.1.7). These structures demonstrate that the moraines were formed at the ice-margin and that glacier retreat was active, incremental and oscillatory with frequent readvances interrupting overall retreat. Plotting the frequency of individual processes based on the data set of 52 exposures demonstrates (a) that all moraines formed ice-marginally, (b) that terrestrial ice-contact fans were the norm, forming 98% of the moraines exposed in the study area and (c) that most moraines formed in a two stage process of fan formation and subsequent deformation, indicating that readvances were the norm and that the glaciers were highly dynamic. Only two exposures contained proglacially-deformed outwash sediments, suggesting that terrestrial ice-contact fans were formed almost ubiquitously. Similar findings from elsewhere in Scotland (Benn, 1990, 1992a; Bennett and Boulton, 1993a; Lukas, 2003; Golledge and Hubbard, 2005) strongly suggest that the alternating subaerial deposition by gravitational and fluvial processes was a characteristic feature of Younger Dryas ice-marginal sedimentation.

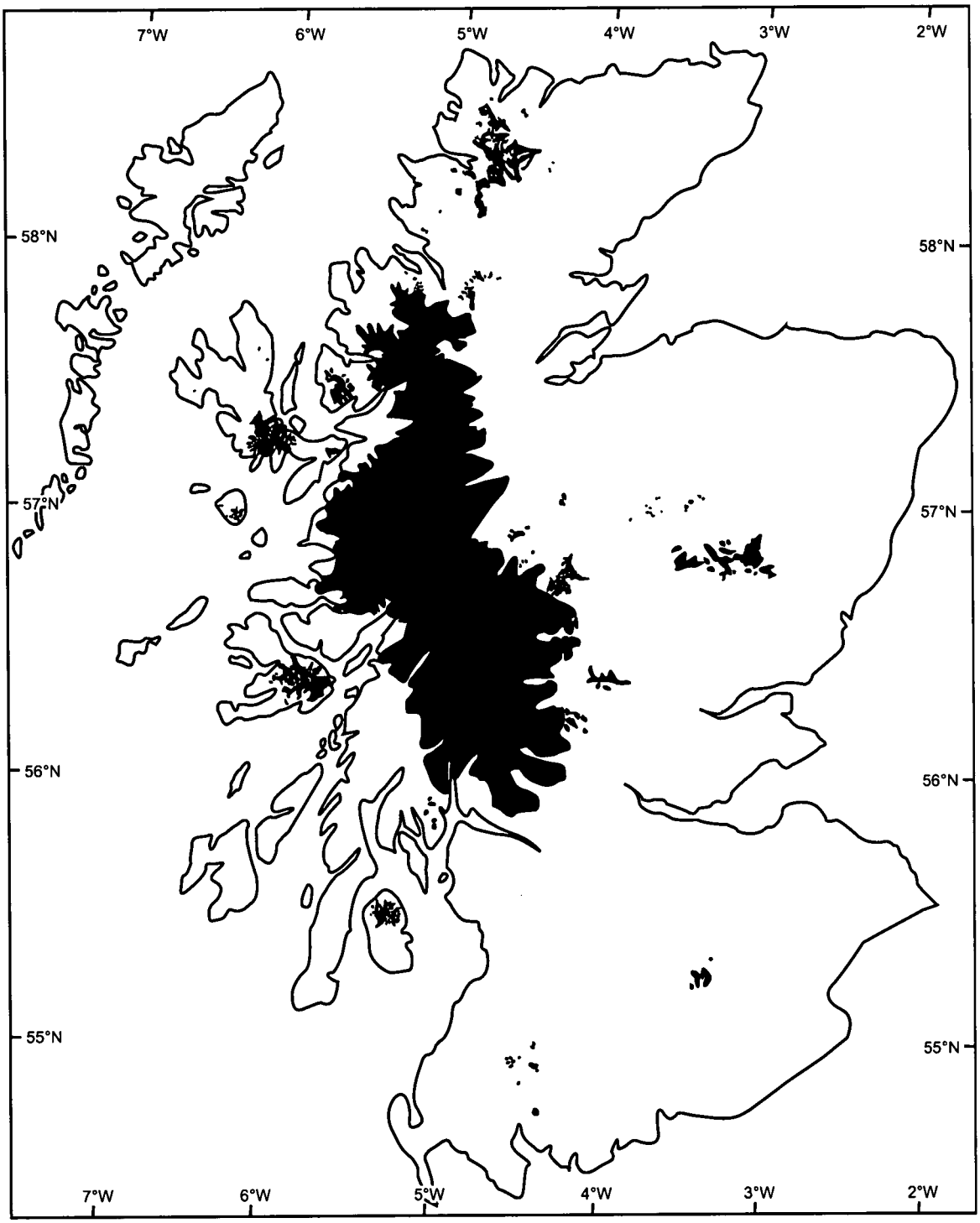


Fig. 6.1 Distribution of Younger Dryas glaciers in Scotland incorporating the evidence presented in this thesis (from various sources; cf. Fig. 1.1).

Clast shape data obtained from diamicts within "hummocky moraines" are very similar to control samples from subglacial till and distinct from scree and fluvial control samples. This indicates that the dominant transport path was subglacial. Together with evidence from roches moutonnées, ice-moulded bedrock outcrops and flutes that all indicate widespread basal sliding, this is interpreted as evidence of a dominant subglacial source of material. Models according to which the material stored

in moraines was derived primarily by reworking of paraglacial sediments (Ballantyne, 2002a, b) cannot explain the lack of systematic variation of height and thickness of debris contained within the moraines. Likewise, the data do not support the suggestion that supraglacial debris sources were dominant during the Younger Dryas in western Scotland (Eyles, 1983a).

6.3.3 Implications

6.3.3.1 Younger Dryas palaeo-glacier dynamics

The moraine sedimentology demonstrates that readvances were the norm and that the glaciers were highly dynamic. Since the moraines have been demonstrated to have formed ice-marginally, individual moraine ridges and mounds can be connected to palaeo-ice fronts, confirming earlier morphological interpretations (Benn, 1990, 1992a; Bennett, 1990; Bennett and Glasser, 1991; Bennett and Boulton, 1993a, b; McDougall, 2001; Lukas, 2003). Analyses of the spacing patterns between these palaeo-ice fronts indicates short response times with moraines having been formed at subdecadal to decadal rates with average retreat velocities between any two moraines ranging from 4 to 30 m a^{-1} (Chapter 4.4.2). This analysis supports earlier suggestions by Ballantyne (2002a) that glaciers in Scotland were close to equilibrium during the second half of the Younger Dryas. The data agree very well with the high Younger Dryas precipitation values in that the latter suggest high mass turnover rates and thus very responsive glaciers (e.g. Nesje *et al.*, 1995; Winkler, 1996b; Benn and Evans, 1998). Evidence for alteration of deposits by dead ice meltout following moraine formation is absent.

The retreat patterns of individual glaciers in the NW Highlands was found to be dependent on basin geometry, glacier size and location in relation to the dominant palaeo-wind direction. In general, smaller, steeper glaciers only show evidence of a short initial phase of retreat followed by uninterrupted decay without moraine formation whereas the larger glaciers probably modified the local climate and remained in equilibrium longer. In the larger basins, "hummocky" recessional moraines extend from the Younger Dryas maximum back into the source areas, indicating uninterrupted retreat, in some cases over considerable distances of up to 16 km. The model of a two-phased deglaciation developed by Benn *et al.* (1992) for the Isle of Skye does not appear to be fully applicable to the larger glacier basins in the far NW Highlands.

6.3.3.2 Applicability of models of moraine formation

The sedimentological data unequivocally demonstrate that the moraines formed at ice margins with proximal rectilinear slopes representing ice-contact faces, thus supporting the model of active, incremental retreat. Conversely, the sedimentological data demonstrate that proximal rectilinear slopes are not, as envisaged by proponents of the englacial thrusting model (e.g. Hambrey *et al.*, 1997; Bennett *et al.*, 1998), the surface representation of former englacial thrust planes. Other purely morphological criteria adduced in support of the thrusting model, such as the presence of reverse

bedrock slopes that might have facilitated thrusting, are also not supported by sedimentological evidence.

The englacial thrusting model is based entirely on data obtained from shallow surface exposures. Isolated clast shape measurements from such exposures that indicate subglacial transport have been used to infer the subglacial origin of entire units, although sedimentary structures necessary to support such an interpretation cannot always be observed at such a small scale. Although clast shape measurements are a powerful tool to discriminate different *transport histories* (cf. Chapters 2.1.2, 4.3.2), these data in isolation cannot be used to infer the *mode of deposition of entire units*. The englacial thrusting model is thus not in the position to provide the necessary information and resolution to formulate a representative and realistic model of moraine formation.

To conclude, the proposal of englacial thrusting as a formative process of Scottish "hummocky moraine" hinges on the morphological criterion of the presence of proximal rectilinear slopes, reverse bedrock slopes and sedimentological data extrapolated from shallow surface exposures. There is no sedimentological proof for englacial thrusting being responsible for any of the moraines summarised under "hummocky moraine" when investigated in detail. Consequently, the concept of englacial thrusting as a formative process of Scottish "hummocky moraine" has to be abandoned.

6.4 Modern analogues

6.4.1 Svalbard

6.4.1.1 Summary of results

The debris-covered termini that mark the Little Ice Age (LIA) maximum extent of three small cold-based valley glaciers in Central Spitsbergen were investigated to observe sedimentary processes at their margins and to compare them to the glaciers used in the development of the englacial thrusting model. The debris was largely transported supra- and englacially, but subglacial transport also played a role in debris-cover formation. The latter implies that temperate areas at the glacier bed must have been more extensive when the glacier was thicker during the LIA. Transfer of material was along flowlines as evident from widespread discrete sediment accumulations on the glaciers interpreted as surface crevasse fills; evidence for englacial thrusting was not found. Meltout at the front and redistribution of debris led to the emplacement of a continuous debris cover and the formation of controlled moraines at its surface.

Processes associated with the meltout of buried ice underneath the debris cover are ubiquitous. Debris flows are initiated where instabilities are formed, for example along the margins of meltwater channels. These propagate upslope along such channels leading to rapid enlargement of exposed ice faces and enhanced melting. Once debris is stripped off the ice surface, a self-reinforcing cycle is initiated resulting in rapid degradation of the ice-marginal debris-covers.

The preservation potential of material and landforms (controlled moraines) over longer timescales is very limited due to glaciofluvial reworking and removal. The observations allow the conclusion that at best a subdued outer dump moraine but more likely an uneven and thin spread of supraglacial material with little surface expression might be the resulting landform after complete de-icing of the landscape. Instead of constructional glacial landforms, meltwater channels and outwash fans are much more likely to indicate a formerly glaciated terrain.

6.4.1.2 Implications

Although the presence of buried glacier ice in moraines inferred to have formed by englacial thrusting is acknowledged by Hambrey *et al.* (1997), Bennett *et al.* (1998) and Midgley (2001), the role of dead ice meltout on landform alteration and the impact on preservation potential is generally ignored when transferring observations from modern Svalbard glaciers to Pleistocene glaciers in Scotland. Rather than being preserved as coherent englacial slabs as envisaged by the thrusting model, material within thrusts would melt out successively as the ice surface lowers, forming a veneer of supraglacial debris (Chapter 5.2.7). Constructional landforms, certainly ones with proximal rectilinear slopes, are unlikely to survive unaltered. It would therefore appear that controlled moraines observed at the margins of modern Svalbard glaciers represent a momentarily stable snapshot situation that would change when the protected buried ice was to be exposed. Thus, in addition to the criticism directed at the englacial thrusting model from the Scottish perspective, it also appears highly unrealistic and needs to be questioned on the basis of de-icing process observations. The claim that "British Loch Lomond Stadial landforms [have] a closer affinity with Arctic glaciers than those in the Alps or Iceland" (Hambrey *et al.*, 2001: 25) thus cannot be maintained.

6.4.2 Norway

Two forelands of temperate outlet glaciers of the Jostedalsbre ice cap in SW Norway were studied as a contrasting modern analogue case study. Both forelands are characterised by sequences of recessional moraines and boulder moraines that reflect the shape of the former ice margin and intervening small ice-contact outwash fans. These were formed during the Little Ice Age (LIA) and recently. Sedimentological studies of a small ice-contact outwash fan revealed deformation structures indicative of proglacial lateral compression. These structures suggest that the glacier margin oscillated during overall retreat and that the landform was formed in a two-stage process. Lichenometric dating of LIA moraines and glaciological studies on modern glaciers show that the glaciers respond in a highly active way to short-term climate forcing. Recessional moraines were formed during net retreat interrupted by smaller readvances at frequencies of between 2 and a maximum of 35 years, which corresponds to the range of response times for nearly all modern glaciers of the Jostedalsbre ice cap (Bickerton and Matthews, 1993; Nesje *et al.*, 1995; Winkler, 1996a, b; Winkler *et al.*, 1997). This

indicates a highly dynamic marginal environment. Due to the limited to absent incorporation of dead ice during moraine formation, the latter are not altered postdepositionally due to meltout and have higher preservation potential than supraglacial controlled moraines.

The Norwegian glacial landsystem is a much closer analogue to Younger Dryas glaciers in NW Scotland based on similarities in moraine planform, spacing and their formation during active, oscillatory retreat. The glaciers have short response times, high mass turnover, and they are closely coupled with climate as evident from the high precipitation totals in both systems. However, a temperate Norwegian analogue cannot explain *all* of the characteristics of Scottish Younger Dryas glaciers, i.e. moraine size and the processes leading to their formation. Pre-requisites that need to be fulfilled by a modern analogue will be considered below.

6.4.3 Synthesis

A comprehensive model of the formation and glaciological significance of "hummocky moraine" must be capable of explaining all of the evidence, including:

- the large volume of debris represented by suites of "hummocky moraine",
- how this debris was transported and elevated to the former glacier surfaces,
- the range of inferred depositional and deformational processes,
- the distinctive morphology of "hummocky moraine" and
- the highly dynamic, oscillatory behaviour of Younger Dryas outlet glaciers in the NW Highlands.

Furthermore, such a model must also be compatible with palaeoclimatic reconstructions. The Scottish Younger Dryas landsystem in the far NW Highlands, and elsewhere, contain *elements* of many different modern glacial landsystems. First, moraine planform, spacing, oscillatory retreat of glaciers with short response times, high precipitation totals and consequently high mass turnover are similar to the characteristics observed in modern-day Norway. However, moraine size and sedimentary processes cannot be reconciled with the Scottish evidence. Second, the sedimentary processes and landforms are similar to modern debris-covered glaciers, but the frontal response to climate change at these glaciers is very different from Younger Dryas glaciers in NW Scotland. Third, the reconstructed low winter temperatures are similar to those in Svalbard, but the processes of moraine formation and the slow response times of Svalbard glaciers cannot be regarded as an analogue for NW Scotland.

On an axis of temperate to cold-based glaciological boundary conditions the Younger Dryas landsystem in NW Scotland plots much closer to a temperate landsystem. This is probably because permafrost was suppressed along the west coast due to an increased precipitation compared to the

present day. High precipitation totals would have led to high mass turnover and efficient subglacial transport. An outermost narrow zone where the glacier was frozen to the bed would probably be required to elevate debris to the glacier surface in sufficient quantities near the margins to facilitate shearing, folding and thrusting. Likewise, evacuation of debris from the surface must have been very efficient, as the presence of thick debris covers would dampen glacier response to climate change (Benn *et al.*, 2003).

According to the present evidence it appears that the Scottish Younger Dryas landsystem in the far NW Highlands can be reconciled with elements of the glaciated valley landsystem (Benn *et al.*, 2003), but may be lack a direct modern analogue, a problem that does not appear to be unique in palaeo-settings (e.g. Johnson and Clayton, 2003).

6.5 Methodological implications

6.5.1 Development of genetic models

The work presented in this thesis has shown that moraines form excellent archives from which palaeo-glacier dynamics can be reconstructed. Crucially, however, it has also shown that the processes of moraine formation can only be understood through the combination of detailed geomorphological mapping and sedimentological logging. Morphological criteria alone can lead to misjudgements due to the problem of equifinality (e.g. the significance of proximal rectilinear slopes). The logging of 52 exposures has shown that detailed information about the mode of moraine formation can only be gained from larger exposures. Shallow surface exposures may allow individual lithofacies to be identified, but the exact mode of deposition and/or deformation cannot be established at this scale (Chapter 4.3.1.7, 4.4.3). Although it is realised that exposure conditions will not be as ideal everywhere, and based on the experiences gained in this study, the author would like to encourage workers seeking to elucidate the processes of landform formation to enlarge existing sections wherever possible or to create new ones. The insight gained through this exercise is far more rewarding than the frustration over missing information from shallow surface exposures (Fig. 4.21). Indeed, it is advocated here that genetic models of moraine or landform formation in general should only be proposed if backed up by detailed sedimentological work.

6.5.2 Morphostratigraphy and geomorphological mapping

Despite the paucity of sites suitable for numerical dating, unifying and formalising the coincidence of distinct landsystem contrasts with Younger Dryas limits as identified elsewhere in Scotland has provided a working template for Younger Dryas glacier reconstructions. Such an approach is useful where age constraints are not available and dateable material is sparse or absent. Since this approach is based on a large body of evidence that has been compiled and tested independently elsewhere in

Scotland over a period of > 30 years, it is suggested that it forms a good starting point for identifying possibly contemporaneous glacial limits and designing a dating programme.

However, although this approach has been confirmed to work for the study area in the NW Highlands, it is by no means suggested here that such an approach should replace the application of numerical dating techniques. To the contrary, the experience gained in this study suggests that the approach can and should be used to guide dating programmes and to identify suitable sites that can test this approach. Such a morphostratigraphical approach is most effective and remains reliable only if multiple lines of evidence converge. The sole use of a single landform assemblage, for example "hummocky moraine", as an indicator of a Younger Dryas age is not desirable as this landform assemblage has been shown not to have formed exclusively during the Younger Dryas (Clapperton *et al.*, 1975; Wilson and Evans, 2000; Everest, 2003; Everest and Golledge, 2004). Therefore, testing this approach and refining it is encouraged.

6.5.3 Optically-stimulated luminescence (OSL) dating

This study has shown that, although well-informed by sedimentological process-studies, attempts to use OSL dating to constrain the timing more closely did not work out. Despite the expected absence of zeroing problems in a supraglacial depositional setting (Chapter 3.2.2) this study has indicated that the depositional *rates* and not only the *environmental setting* have to be considered when dating glaciogenic sediments by OSL methods. One possible solution to increase the likelihood of success could be to conduct more studies characterising modern glacigenic sediments to understand zeroing rates and required transport distances in proglacial rivers, for example, so that sampling strategies can be designed accordingly (cf. Preusser, 1999; Wallinga, 2002). Since OSL-dating has proven successful on different (non-metamorphic) lithologies elsewhere in Scotland, e.g. granites (Everest and Golledge, 2004), dating programmes should aim at sediments derived from such source rocks to eliminate potential error sources. OSL dating also has the potential to date outwash terraces that are genetically linked to glacier limits, and this might help to overcome the problems of dating moraines directly. Likewise, OSL-dating of sediment retrieved by coring (such as LST 1; Chapter 3.2.1.3) might help to overcome the problems of contamination associated with the radiocarbon dating of organic sediments (cf. Preusser *et al.*, 2002; Juschus *et al.*, 2005).

6.6 Scope for further research

This section deals with problems that could not be solved in this investigation due to several reasons. Firstly, compared to other areas studied over the course of a PhD project, the present study area is very large. Secondly, and again in contrast to most other areas, very little research has been carried out in the far NW Highlands. This left a large amount of basic work to be done at the beginning of this project before detailed work could commence. This thesis is the attempt to present a balanced selection of aspects that could be tackled in the present study area, but necessarily, some aspects had to be left out due to time constraints. Thirdly, while this study has answered a number of questions, it has also generated new ones that present exciting opportunities for follow-up projects. The most interesting and important aspects will be addressed below.

6.6.1 Constraining the age of individual outlet glaciers and pre-Younger Dryas events

One of the priorities of future studies in the present study area should be to constrain the age of Quaternary glaciations in NW Scotland in general. There is evidence for glacial limits outside the Younger Dryas limits that merit more detailed investigation as these could shed light on the activity of the last ice sheet in the far NW Highlands. Furthermore, the patterns of ice sheet flow and its interactions with areas to the east (Caithness) and west (Assynt) deserve attention.

Constraining the age of the Younger Dryas ice cap further is of crucial importance. A greater density of dated sites could be achieved by, for example, coring the lochs in the area. A particularly rewarding exercise to establish more detailed chronologies of retreat across the study area might be to core lochs along the NW-SE axis of the ice cap following the principles established by Dahl *et al.* (2004) and Bakke *et al.* (2005), for example. Where bounded by bedrock thresholds, e.g. the transition from Loch More to Loch Stack, such sites are ideal as they give a good indication of the cessation of meltwater and thus suspended sediment input into these basins (Bakke *et al.*, 2005). Employing this approach near the source areas of the glaciers might help resolve the question of the timing of final deglaciation at the beginning of the Holocene. In addition, cosmogenic radionuclide (CRN) dating could likewise be attempted as sufficiently large boulders with quartz veins suitable for dating with ^{10}Be and ^{26}Al are abundant on moraine crests.

6.6.2 Modelling glacier velocities, response times and climate-glacier interactions

The detailed mapping carried out in the study area presents a good basis for testing existing numerical models (cf. Golledge and Hubbard, 2005). An exciting possibility would be to model Younger Dryas glacier velocities using the palaeo-precipitation values calculated for the whole NW Highlands ice cap. Data on glacier geometry have also been presented allowing individual cross-sections and thus mass transfer in individual glacier basins, and for the ice cap as a whole, to be calculated. Such data could

then be used to model response times of Younger Dryas palaeo-glaciers and compare these to values estimated from moraine spacings and to those calculated for modern glaciers. This way, qualitative estimates of short response times could be further scrutinised and quantified. Furthermore, it could be potentially rewarding to attempt a replication of the retreat patterns marked by individual moraine arcs. This latter aspect could be an important link in understanding the differential response of individual outlet glaciers to local short-term climate fluctuations and conditions such as snowblow, precipitation shadow effects and basin topography, for example.

6.6.3 "Hummocky moraine" formation and glacial landsystems across Scotland

This study has demonstrated that englacial thrusting is not responsible for the formation of "hummocky moraines" with proximal rectilinear slopes in the study area and that this criterion cannot be used as an indicator for palaeo-thrusting. Likewise, it is important to test other morphological predictions of the englacial thrusting model such as the predicted importance of reverse bedrock slopes which could not unequivocally be tested in the study area. To test this idea thoroughly, it would be advisable to investigate those sites where englacial thrusting has been inferred as the mechanism of moraine formation, e.g. the Valley of a Hundred Hills in Glen Torridon, Glen Geusachan in the Cairngorms or Ennerdale in the English Lake District.

This study has shown that detailed sedimentological investigations are crucial in determining the mode of moraine formation at any given site. Although the variability of facies associations is not great in the study area, more complex and variable conditions are to be expected elsewhere in Scotland.

So far, sedimentological studies serving to elucidate the formation of Scottish moraines have focused on the west coast and western Highlands of Scotland (e.g. Benn, 1990, 1992a; Benn and Evans, 1993, 1996; Phillips *et al.*, 2002; Golledge and Hubbard, 2005; Lukas, 2005, this study), but virtually nothing is known about depositional environments in the central and eastern Highlands where conditions are likely to have been somewhat more arid due to precipitation scavenging (Benn and Ballantyne, 2005). Therefore, balancing this spatial disparity and increasing the dataset of different processes contributing to the formation of "hummocky moraine" would be an interesting and necessary avenue for future sedimentological research in Scotland. Thus, further sedimentological studies are urgently required.

Furthermore, such research would enable understanding of the spatial variability of glacial landsystems during the Younger Dryas in Scotland and its implications for palaeoclimatic and palaeoglaciological conditions. In this respect, the different retreat patterns recognised across Scotland, most notably the possibility of uninterrupted active retreat throughout the Younger Dryas in the study area, deserve further attention. This thesis has highlighted the need for more detailed studies

into Scottish sediment-landform associations to understand the spatial variability of glacier dynamics as part of defining the characteristics of the Scottish Younger Dryas glacial landsystem.

REFERENCES

Aber, J.S., Croot, D.G., Fenton, M.M., 1989. *Glaciotectonic structures and landforms*. Amsterdam: Kluwer.

Agassiz, L., 1840. *Études sur les Glaciers*. Neuchâtel.

Aitken, M.J., 1998. *An introduction to optical dating*. Oxford: Oxford Science Publications.

Albertz, J., 2001. *Einführung in die Fernerkundung. Grundlagen der Interpretation von Luft- und Satellitenbildern*. 2nd ed., Darmstadt: Wissenschaftliche Buchgesellschaft.

Alley, R.B., Cuffey, K.M., Evenson, E.B., Strasser, J.C., Lawson, D.E., Larson, G.J., 1997. How glaciers entrain and transport basal sediment: physical constraints. *Quaternary Science Reviews* **16**: 1017-1038.

Anderson, D.E., 1997. Younger Dryas research and its implications for understanding abrupt climatic change. *Progress in Physical Geography* **21**: 230-249.

Anderson, R.S., 2000. A model of ablation-dominated medial moraines and the generation of debris-mantled glacier snouts. *Journal of Glaciology* **46**: 459-469.

Ashley, G.M., 1995. Glaciolacustrine environments. In: Menzies, J. (ed.), *Modern glacial environments: processes, dynamics and sediments*. Oxford: Butterworth-Heinemann, 417-444.

Atkinson, T.C., Briffa, K.R., Coope, G.R., 1987. Seasonal temperatures in Britain during the past 22,000 years reconstructed using beetle remains. *Nature* **325**: 587-592.

Auton, C.A., 1990. The middle Findhorn Valley. In: Auton, C.A., Firth, C.R. and Merritt, J.W. (eds.), *Beauly to Nairn: Field Guide*. Cambridge: Quaternary Research Association, 74-96.

Bahr, D.B., Pfeffer, W.T., Sassolas, C., Meier, M.F., 1998. Response time of glaciers as a function of size and mass balance: 1. theory. *Journal of Geophysical Research* **103**: 9777-9782.

Bakke, J., Lie, Ø., Nesje, A., Dahl, S.O., Paasche, Ø. 2005. Utilizing physical sediment variability in glacier-fed lakes for continuous glacier reconstructions during the Holocene, northern Folgefonna, western Norway. *The Holocene* **15**: 161-176.

Ballantyne, C.K., 1982. Aggregate clast form characteristics of deposits at the margins of four glaciers in the Jotunheimen Massif, Norway. *Norsk Geografisk Tidsskrift* **36**: 103-113.

Ballantyne, C.K., 1983. Precipitation gradients in Wester Ross, north-west Scotland. *Weather* **38**: 379-387.

Ballantyne, C.K., 1984. The Late Devensian periglaciation of upland Scotland. *Quaternary Science Reviews* **3**: 311-343.

Ballantyne, C.K., 1986. Protalus rampart development and the limits of former glaciers in the vicinity of Baosbheinn, Wester Ross. *Scottish Journal of Geology* **22**: 13-25.

Ballantyne, C.K., 1988. Ice-sheet moraines in southern Skye. *Scottish Journal of Geology* **24**: 301-304.

Ballantyne, C.K., 1989. The Loch Lomond Readvance on the Isle of Skye, Scotland: glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science* **4**: 95-108.

Ballantyne, C.K., 1991. Holocene geomorphic activity in the Scottish Highlands. *Scottish Geographical Magazine* **107**: 84-98.

Ballantyne, C.K., 1995. Paraglacial debris-cone formation on recently deglaciated terrain, western Norway. *The Holocene* **5**: 25-33.

Ballantyne, C.K., 1997. Periglacial trimlines in the Scottish Highlands. *Quaternary International* **38/39**: 119-136.

Ballantyne, C.K., 1998. Age and significance of mountain-top detritus. *Permafrost and periglacial processes* **9**: 327-345.

Ballantyne, C.K., 2002a. The Loch Lomond Readvance on the Isle of Mull, Scotland: glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science* **17**: 759-771.

Ballantyne, C.K., 2002b. Paraglacial geomorphology. *Quaternary Science Reviews* **21**: 1935-2017.

Ballantyne, C.K., 2003. Stac Pollaidh-a Late Devensian nuntak? A comment on Bradwell and Krabbendam (2003). *Quaternary Newsletter* **101**: 34-38.

Ballantyne, C.K., Wain-Hobson, T., 1980. The Loch Lomond Advance on the Island of Rhum. *Scottish Journal of Geology* **16**: 1-10.

REFERENCES

Ballantyne, C.K., Eckford, J.D., 1984. Characteristics and evolution of two relict talus slopes in Scotland. *Scottish Geographical Magazine* **100**: 20-33.

Ballantyne, C.K., Gray, J.M., 1984. The Quaternary of Scotland: the research contributions of J.B. Sissons. *Quaternary Science Reviews* **3**: 259-289.

Ballantyne, C.K., Kirkbride, M., 1986. The characteristics and significance of some Lateglacial protalus ramparts in upland Britain. *Earth Surface Processes and Landforms* **11**: 659-671.

Ballantyne, C.K., Kirkbride, M., 1987. Rockfall activity in upland Britain during the Loch Lomond Stadial. *Geographical Journal* **153**: 86-92.

Ballantyne, C.K., Benn, D.I., 1994. A Loch Lomond Readvance glacier in Duirinish, NW Skye. *Scottish Journal of Geology* **30**: 183-186.

Ballantyne, C.K., Harris, C., 1994. *The Periglaciation of Great Britain*. Cambridge: Cambridge University Press.

Ballantyne, C.K., McCarroll, D., Nesje, A., Dahl, S.O., 1997. Periglacial trimlines, former nunataks and the altitude of the last ice sheet in Wester Ross, northwest Scotland. *Journal of Quaternary Science* **12**: 225-238.

Ballantyne, C.K., McCarroll, D., Nesje, A., Dahl, S.O., Stone, J.O., Fifield, L.K., 1998a. High-resolution reconstruction of the last ice sheet in NW Scotland. *Terra Nova* **10**: 63-67.

Ballantyne, C.K., McCarroll, D., Nesje, A., Dahl, S.O., Stone, J.O., 1998b. The last ice sheet in north-west Scotland: Reconstructions and implications. *Quaternary Science Reviews* **17**: 1149-1184.

Barrow, G., Hinxman, L.W., Craig, E.H.C., 1913. *The geology of upper Strathspey, Gaick and the Forest of Atholl*. Memoir of the Geological Survey of Scotland. Edinburgh: HMSO.

Barsch, D., Gude, M., Mäusbacher, R., Schukraft, G., Schulte, A., 1994. Recent fluvial sediment budgets in glacial and periglacial environments, NW Spitsbergen. *Zeitschrift für Geomorphologie Supplement Band* **97**: 111-122.

Benn, D.I., 1989. Controls on sedimentation in a Late Devensian ice-dammed lake, Achnasheen, Scotland. *Boreas* **18**: 31-42.

Benn, D.I., 1990. *Scottish Lateglacial moraines: debris supply, genesis and significance*. Unpublished PhD Thesis, University of St Andrews.

Benn, D.I., 1991a. Glacial landforms and sediments on Skye. In: Ballantyne, C.K., Benn, D.I., Lowe, J.J. and Walker, M.J.C. (eds.), *The Quaternary of the Isle of Skye: Field Guide*. Cambridge: Quaternary Research Association, 35-67.

Benn, D.I., 1991b. Raised shorelines on Skye. In: Ballantyne, C.K., Benn, D.I., Lowe, J.J. and Walker, M.J.C. (eds.), *The Quaternary of the Isle of Skye: Field Guide*. Cambridge: Quaternary Research Association, 90-97.

Benn, D.I., 1992a. The genesis and significance of 'hummocky moraine': evidence from the Isle of Skye, Scotland. *Quaternary Science Reviews* **11**: 781-799.

Benn, D.I., 1992b. Scottish Landform Examples - 5. The Achnasheen terraces. *Scottish Geographical Magazine* **108**: 128-131.

Benn, D.I., 1994. Fluted moraine formation and till genesis below a temperate glacier: Slettmarkbreen, Jotunheimen, Norway. *Sedimentology* **41**: 279-292.

Benn, D.I., 1995. Fabric signature of subglacial till deformation, Breidamerkurjökull, Iceland. *Sedimentology* **42**: 735-747.

Benn, D.I., 1996. Subglacial and subaqueous processes near a glacier grounding line: sedimentological evidence from a former ice-dammed lake, Achnasheen, Scotland. *Boreas* **25**: 23-36.

Benn, D.I., 1997a. Glacier fluctuations in western Scotland. *Quaternary International* **38/39**: 137-147.

Benn, D.I., 1997b. Glacial landforms and sediments in Scotland. In: Gordon, J.E. (ed.), *Reflections on the Ice Age in Scotland*. Glasgow: Scottish Association of Geography Teachers, 152-165.

Benn, D.I., 2004. Clast morphology. In: Evans, D.J.A. and Benn, D.I. (eds.), *A practical guide to the study of glacial sediments*. London: Arnold, 78-92.

Benn, D.I., Ballantyne, C.K., 1993. The description and representation of particle shape. *Earth Surface Processes and Landforms* **18**: 665-672.

Benn, D.I., Ballantyne, C.K., 1994. Reconstructing the transport history of glacigenic sediments: a new approach based on the co-variance of clast shape indices. *Sedimentary Geology* **91**: 215-227.

REFERENCES

Benn, D.I., Ballantyne, C.K., 2004. Evidence for an independent West Drumochter icefield. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 200-209.

Benn, D.I., Ballantyne, C.K., 2005. Palaeoclimatic reconstruction from Loch Lomond Readvance glaciers in the West Drumochter Hills, Scotland. *Journal of Quaternary Science* **20**: in press.

Benn, D.I., Clapperton, C.M., 2000. Pleistocene glacitectonic landforms and sediments around central Magellan Strait, southernmost Chile: evidence for fast outlet glaciers with cold-based margins. *Quaternary Science Reviews* **19**: 591-612.

Benn, D.I., Evans, D.J.A., 1993. Glaciomarine deltaic deposition and ice-marginal tectonics: the 'Loch Don Sand Moraine', Isle of Mull, Scotland. *Journal of Quaternary Science* **8**: 279-291.

Benn, D.I., Evans, D.J.A., 1996. The interpretation and classification of subglacially-deformed materials. *Quaternary Science Reviews* **15**: 23-52.

Benn, D.I., Evans, D.J.A., 1998. *Glaciers and glaciation*. London: Arnold.

Benn, D.I., Evans, D.J.A., 2004. Introduction and rationale. In: Evans, D.J.A. and Benn, D.I. (eds.), *A practical guide to the study of glacial sediments*. London: Arnold, 1-10.

Benn, D.I., Gemmell, A.M.D., 1997. Calculating equilibrium-line altitudes of former glaciers by the balance ratio method: a new computer spreadsheet. *Glacial Geology and Geomorphology*: http://ggg.qub.ac.uk/ggg/papers/full/1997/tn011997/tn01.html.

Benn, D.I., Lehmkuhl, F., 2000. Mass balance and equilibrium-line altitudes of glaciers in high-mountain environments. *Quaternary International* **65/66**: 15-29.

Benn, D.I., Owen, L.A., 2002. Himalayan glacial sedimentary environments; a framework for reconstructing and dating the former extent of glaciers in high mountains. *Quaternary International* **97-98**: 3-25.

Benn, D.I., Lowe, J.J., Walker, M.J.C., 1992. Glacier response to climatic change during the Loch Lomond Stadial and early Flandrian: geomorphological and palynological evidence from the Isle of Skye, Scotland. *Journal of Quaternary Science* **7**: 125-144.

Benn, D.I., Kirkbride, M.P., Owen, L.A., Brazier, V., 2003. Glaciated valley landsystems. In: Evans, D.J.A. (ed.), *Glacial Landsystems*. London: Arnold, 372-406.

Benn, D.I., Evans, D.J.A., Phillips, E.R., Hiemstra, J.F., Walden, J., Hoey, T.B., 2004. The research project - a case study of Quaternary glacial sediments. In: Evans, D.J.A. and Benn, D.I. (eds.), *A practical guide to the study of glacial sediments*. London: Arnold, 209-234.

Benn, D.I., Owen, L.A., Osmaston, H.A., Seltzer.G.O., Porter, S.C., Mark, B., 2005. Reconstruction of equilibrium-line altitudes for tropical and sub-tropical glaciers. *Quaternary International* **138-139**: 8-21.

Bennett, M.R., 1990. The deglaciation of Glen Croulin, Knoydart. *Scottish Journal of Geology* **26**: 41-46.

Bennett, M.R., 1994. Morphological evidence as a guide to deglaciation following the Loch Lomond Readvance: a review of research approaches and models. *Scottish Geographical Magazine* **110**: 24-32.

Bennett, M.R., 1995. The morphology of glacially fluted terrain: examples from the Northwest Highlands of Scotland. *Proceedings of the Geologist's Association* **106**: 27-38.

Bennett, M.R., 1996. The Loch Lomond Readvance in the Cairngorm Mountains. In: Glasser, N.F. and Bennett, M.R. (eds.), *The Quaternary of the Cairngorms: Field Guide*. London: Quaternary Research Association, 54-68.

Bennett, M.R., 2001. The morphology, structural evolution and significance of push moraines. *Earth-Science Reviews* **53**: 197-236.

Bennett, M.R., 2003. Ice streams as the arteries of an ice sheet: their mechanics, stability and significance. *Earth-Science Reviews* **61**: 309-339.

Bennett, M.R., Glasser, N.F., 1991. The glacial landforms in Glen Geusachan, Cairngorms: a reinterpretation. *Scottish Geographical Magazine* **107**: 116-123.

Bennett, M.R., Boulton, G.S., 1993a. A reinterpretation of Scottish 'hummocky moraine' and its significance for the deglaciation of the Scottish Highlands during the Younger Dryas or Loch Lomond Stadial. *Geological Magazine* **130**: 301-318.

Bennett, M.R., Boulton, G.S., 1993b. Deglaciation of the Younger Dryas or Loch Lomond Stadial ice-field in the northern Highlands, Scotland. *Journal of Quaternary Science* **8**: 133-146.

Bennett, M.R., Hambrey, M.J., Huddart, D., 1997. Modification of clast shape in high-arctic glacial environments. *Journal of Sedimentary Research* **67**: 550-559.

Bennett, M.R., Hambrey, M.J., Huddart, D., Glasser, N.F., 1998. Glacial thrusting & moraine-mound formation in Svalbard & Britain: the example of Coire a' Cheud-chnoic (Valley of hundred hills), Torridon, Scotland. *Quaternary Proceedings* **6**: 17-34.

Bickerton, R.W., Matthews, J.A., 1993. 'Little Ice Age' variations of outlet glaciers from the Jostedalsbreen ice-cap, southern Norway: a regional lichenometric-dating study of ice-marginal moraine sequences and their climatic significance. *Journal of Quaternary Science* **8**: 45-66.

Birks, H.H., 1984. Late-Quaternary pollen and plant macrofossil stratigraphy at Lochan an Druim, north-west Scotland. In: Haworth, E.Y. and Lund, J.W.G. (eds.), *Lake sediments and environmental history*. Leicester: Leicester University Press, 377-405.

Birks, H.H., 2003. The importance of plant macrofossils in the reconstruction of Lateglacial vegetation and climate: examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* **22**: 453-473.

Björck, S., Walker, M.J.C., Cwynar, L.C., Johnsen, S., Knudsen, K.-L., Lowe, J.J., Wohlfarth, B., 1998. An event stratigraphy for the Last Termination in the North Atlantic region based on the Greenland ice-core record: a proposal by the INTIMATE group. *Journal of Quaternary Science* **13**: 283-292.

Björnsson, H., Gjessing, Y., Hamram, S.E., Hagen, J.O., Liestøl, O., Pålsson, F., Erlingsson, B., 1996. The thermal regime of subpolar glaciers mapped by multi-frequency radio-echo soundings. *Journal of Glaciology* **12**: 23-31.

Blümel, W.D., Eberle, J., Eitel, B., 1994. Zur jungquartären Vereisungsgeschichte und Landschaftsentwicklung in NW-Spitzbergen (Liefde-, Bock- und Woodfjord). *Zeitschrift für Geomorphologie Supplement Band* **97**: 31-42.

Bonney, T.G., 1871. On a cirque in the syenite hills of Skye. *Geological Magazine* **8**: 535-540.

Boulton, G.S., 1968. Flow tills and related deposits on some Vestspitsbergen glaciers. *Journal of Glaciology* **7**: 391-412.

Boulton, G.S., 1976. A genetic classification of tills and criteria for distinguishing tills of different origin. *Geografia* **12**: 65-80.

Boulton, G.S., 1978. Boulder shapes and grain-size distributions of debris as indicators of transport paths through a glacier and till genesis. *Sedimentology* **25**: 773-799.

Boulton, G.S., 1986. Push-moraines and glacier-contact fans in marine and terrestrial environments. *Sedimentology* **33**: 677-698.

Boulton, G.S., 1992. 'Hummocky moraine' on Rannoch Moor and in the Loch Tulla area. In: Walker, M.J.C., Gray, J.M. and Lowe, J.J. (eds.), *The South-west Scottish Highlands: Field Guide*. Cambridge: Quaternary Research Association, 50-52.

Boulton, G.S., Eyles, N., 1979. Sedimentation by valley glaciers: a model and genetic classification. In: Schlüchter, C. (ed.), *Moraines and varves*. Rotterdam: Balkema, 11-23.

Bowen, D.Q., Phillips, F.M., McCabe, A.M., Knutz, P.C., Sykes, G.A., 2002. New data for the Last Glacial Maximum in Great Britain and Ireland. *Quaternary Science Reviews* **21**: 89-101.

Bradwell, T., 2004. Annual moraines and summer temperatures at Lambatungnajökull, Iceland. *Arctic, Antarctic and Alpine Research* **36**: 502-508.

Bradwell, T., 2005. Bedrock megagrooves in Assynt, NW Scotland. *Geomorphology* **65**: 195-204.

Bradwell, T., Krabbendam, M., 2003a. Stac Pollaidh-a Late Devensian nunatak? *Quaternary Newsletter* **100**: 19-25.

Bradwell, T., Krabbendam, M., 2003b. Stac Pollaidh-a Late Devensian nunatak? A reply to Ballantyne (2003). *Quaternary Newsletter* **101**: 39-43.

Bridgland, D.R., 2000. River terrace systems in north-west Europe: An archive of environmental change, uplift and early human occupation. *Quaternary Science Reviews* **19**: 1293-1303.

Brooks, S.J., Birks, H.J.B., 2000. Chironomid-inferred late-glacial air temperatures at Whitrig Bog, south-east Scotland. *Journal of Quaternary Science* **15**: 759-764.

REFERENCES

Brooks, S.J., Birks, H.J.B., 2001. Chironomid-inferred air temperatures from Lateglacial and Holocene sites in north-west Europe: progress and problems. *Quaternary Science Reviews* **20**: 1723-1741.

Brown, I.M., 1993. Pattern of deglaciation of the last (Late Devensian) Scottish ice sheet: evidence from ice-marginal deposits in the Dee valley, northeast Scotland. *Journal of Quaternary Science* **8**: 235-250.

Chadwick, O.A., Hall, R.D., Phillips, F.M., 1997. Chronology of glacial advances in the central Rocky Mountains. *Geological Society of America Bulletin* **109**: 1443-1452.

Charlesworth, J.K., 1926. The Readvance, marginal kame-moraine of the South of Scotland, and some later stages of retreat. *Transactions of the Royal Society of Edinburgh* **55**: 25-50.

Charlesworth, J.K., 1955. Late-glacial history of the Highlands and islands of Scotland. *Transactions of the Royal Society of Edinburgh* **63**: 769-928.

Charman, D.J., 1994. Late-glacial and Holocene vegetation history of the Flow Country, northern Scotland. *New Phytologist* **127**: 155-168.

Charman, D.J., Caseldine, C., Baker, A., Gearey, B., Hatton, J., Proctor, C.J., 2001. Paleohydrological records from peat profiles and speleothems in Sutherland, northwest Scotland. *Quaternary Research* **55**: 223-234.

Christiansen, H.H., 2004. Windpolsihed boulders and bedrock in the Scottish Highlands: evidence and implications of Late Devensian wind activity. *Boreas* **33**: 82-94.

Clapperton, C.M., 1997. Greenland ice cores and North Atlantic sediments: Implications for the Last Glaciation in Scotland. In: Gordon, J.E. (ed.), *Reflections on the ice age in Scotland- an update on Quaternary studies*. Glasgow: Scottish Association of Geography Teachers, 45-58.

Clapperton, C.M., Sugden, D.E., 1977. The Late Devensian glaciation of north east Scotland. In: Gray, J.M. and Lowe, J.J. (eds.), *Studies in the Scottish Lateglacial environment*. Oxford: Pergamon, 1-14.

Clapperton, C.M., Gunson, A.R., Sugden, D.E., 1975. Loch Lomond Readvance in the eastern Cairngorms. *Nature* **253**: 710-712.

Clayton, L., 1964. Karst topography on stagnant glaciers. *Journal of Glaciology* **5**: 107-112.

Colhoun, E.A., 1988. Recent morphostratigraphic studies of the Australian Quaternary. *Progress in Physical Geography* **12**: 264-281.

Coope, G.R., Lemdahl, G., Lowe, J.J., Walkling, A., 1998. Temperature gradients in northern Europe during the last glacial-Holocene transition (14-9 ^{14}C kyr BP) interpreted from coleopteran assemblages. *Journal of Quaternary Science* **13**: 419-433.

Corfu, F., Crane, A., Moser, D., Rogers, G., 1998. U-Pb zircon systematics at Gruinard Bay, northwest Scotland: implications for the early orogenic evolution of the Lewisian complex. *Contributions to Mineralogy and Petrology* **133**: 329-345.

Curry, A.M., 1999. Paraglacial modification of slope form. *Earth Surface Processes and Landforms* **24**: 1213-1228.

Curry, A.M., 2000. Holocene reworking of drift-mantled hillslopes in the Scottish Highlands. *Journal of Quaternary Science* **15**: 529-541.

Curry, A.M., Ballantyne, C.K., 1999. Paraglacial modification of hillslope glacigenic drift. *Geografiska Annaler* **81A**: 409-419.

Dahl, S.O., Nesje, A., Øvstedal, J., 1997. Cirque glaciers as morphological evidence for a thin Younger Dryas ice sheet in east-central Norway. *Boreas* **26**: 161-180.

Dahl, S.O., Bakke, J., Lie, Ø., Nesje, A., 2003. Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: an evaluation of approaches and selection of sites. *Quaternary Science Reviews* **22**: 275-287.

Dahms, D.E., 2002. Glacial stratigraphy of Stough Creek Basin, Wind River Range, Wyoming. *Geomorphology* **42**: 59-83.

Dallmann, W.K., Kjærnet, T., Nøttvedt, A., 2001. *Geological map of Svalbard 1:100,000 - Sheet C9Q Adventdalen. Explanatory text*. Tromsø: Norsk Polarinstitutt.

Drewry, D., J., 1972. A quantitative assessment of dirt-cone dynamics. *Journal of Glaciology* **11**: 431-446.

Dyke, A.S., Evans, D.J.A., 2003. Ice-marginal terrestrial landsystems: northern Laurentide and Innuitian ice sheet margins. In: Evans, D.J.A. (ed.), *Glacial landsystems*. London: Arnold, 143-165.

Dyke, A.S., Savelle, J.M., 2000. Major end moraines of Younger Dryas age on Wollaston Pensinsula, Victoria Island, Canadian arctic: implications for palaeoclimate and for formation of hummocky moraine. *Canadian Journal of Earth Sciences* **37**: 601-619.

Eden, D.N., Marden, M., Berryman, K.R., Palmer, A.S., Cronin, S.J., 2001. Dating the culmination of river aggradation at the end of the end of last glaciation using distal tephra compositions, eastern North Island, New Zealand. *Geomorphology* **38**: 133-151.

Eitel, B., van der Borg, K., Eberle, J., Megies, H., 2002. Late Pleistocene/Early Holocene glacial history of northern Andreeland (northern Spitsbergen/Svalbard Archipelago): evidence from glacial and fluvio-glacial deposits. *Zeitschrift für Geomorphologie* **46**: 337-364.

Eklund, A., Hart, J.K., 1996. Glaciotectonic deformation within a flute from the Isfallsglaciären, Sweden. *Journal of Quaternary Science* **11**: 299-310.

Enters, D., 2000. *Primäre Sukzession auf dem Gletschervorfeld des Rieperbreen, Spitzbergen*. MSc Thesis, University of Karlsruhe, Germany.

Erikstad, L., Sollid, J.L., 1986. Neoglaciation in south Norway using lichenometric methods. *Norsk Geografisk Tidsskrift* **40**: 85-105.

Etienne, J.L., Glasser, N.F., Hambrey, M.J., 2003. Proglacial sediment-landform associations of a polythermal glacier: Storglaciären, northern Sweden. *Geografiska Annaler* **85A**: 149-164.

Etzelmüller, B., 2000. Quantification of thermo-erosion in pro-glacial areas - examples from Svalbard. *Zeitschrift für Geomorphologie* **44**: 343-361.

Etzelmüller, B., Ødegard, R.S., Vatne, G., Mysterud, R.S., Tonning, T., Sollid, J.L., 2000. Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen. *Norsk Geografisk Tidsskrift* **154**: 157-168.

Evans, D.J.A., 1999. Glacial debris transport and moraine deposition: a case study of the Jardalen cirque complex, Sogn-og-Fjordane, western Norway. *Zeitschrift für Geomorphologie* **34**: 481-503.

Evans, D.J.A., 2003a. Ice-marginal terrestrial landsystems: active temperate margins. In: Evans, D.J.A. (ed.), *Glacial landsystems*. London: Arnold, 12-43.

Evans, D.J.A. (ed.), 2003b. *Glacial landsystems*. London: Arnold.

Evans, D.J.A., 2003c. Introduction to glacial landsystems. In: Evans, D.J.A. (ed.), *Glacial landsystems*. London: Arnold, 1-11.

Evans, D.J.A., Rea, B.R., 1999. Geomorphology and sedimentology of surging glaciers: a land-systems approach. *Annals of Glaciology* **28**: 75-82.

Evans, D.J.A., Twigg, D.R., 2002. The active temperate glacial landsystem: a model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* **21**: 2143-2177.

Evans, D.J.A., Benn, D.I., 2004. Facies description and the logging of sedimentary exposures. In: Evans, D.J.A. and Benn, D.I. (eds.), *A practical guide to the study of glacial sediments*. London: Arnold, 11-51.

Evans, D.J.A., Lemmen, D.S., Rea, B.R., 1999. Glacial landsystems of the southwest Laurentide Ice Sheet: modern Icelandic analogues. *Journal of Quaternary Science* **14**: 673-691.

Evans, D.J.A., Clark, C.D., Mitchell, W.A., 2005. The last British Ice Sheet: a review of the evidence utilised in the compilation of the Glacial Map of Britain. *Earth-Science Reviews* **70**: 253-312.

Everest, J.D., 2003. *The Late Devensian deglaciation in the Cairngorm mountains, Scotland*. Unpublished PhD Thesis, University of Edinburgh.

Everest, J.D., Golledge, N.R., 2004. Dating deglaciation in Strath Spey and the Cairngorm Mountains. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 50-57.

Eybergen, F.A., 1987. Glacier snout dynamics and comtemporary push moraine formation at the Turtmann-glacier, Wallis, Switzerland. In: van der Meer, J.J.M. (ed.), *Tills and glaciotectonics*. Rotterdam: Balkema, 217-231.

Eyles, N., 1983a. Modern Icelandic glaciers as depositional models for 'hummocky moraine' in the Scottish Highlands. In: Evenson, E.B., Schlüchter, C. and Rabassa, J. (eds.), *Tills and related deposits*. Rotterdam: Balkema, 47-59.

Eyles, N., 1983b. The glaciated valley landsystem. In: Eyles, N. (ed.), *Glacial geology*. Oxford: Pergamon, 91-110.

Eyles, N., Eyles, C.H., Miall, A.D., 1983. Lithofacies types and vertical profile models. An alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. *Sedimentology* **30**: 393-410.

Eyles, N., Boyce, J.I., Barendregt, R.W., 1999. Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology* **123**: 163-174.

Finsterwalder, R., 1951. The glaciers of Jostedalsbreen. *Journal of Glaciology* **1**: 557-558.

Flint, R.F., 1971. *Glacial and Quaternary geology*. New York: Wiley & Sons.

Florineth, D., Schlüchter, C., 1998. Reconstructing the Last Glacial Maximum (LGM) ice surface geometry and flowlines in the Central Swiss Alps. *Eclogae Geologicae Helvetiae* **91**: 391-407.

Florineth, D., Schlüchter, C., 2000. Alpine evidence for atmospheric circulation patterns in Europe during the Last Glacial Maximum. *Quaternary Research* **54**: 295-308.

Forbes, J.D., 1846. Notes on the topography and geology of the Cuchullin Hills in Skye, and on traces of ancient glaciers that they present. *Edinburgh New Philosophical Journal* **40**: 76-99.

Freeman, S.R., Butler, R.W.H., Cliff, R.A., Rex, D.C., 1998. Dating mylonite evolution: an Rb-Sr and K-Ar study of the Moine mylonites, NW Scotland. *Journal of the Geological Society of London* **155**: 745-758.

French, H.M., 1996. *The periglacial environment*. 2nd ed., London: Longman.

Furbish, D.J., Andrews, J.T., 1984. The use of hypsometry to indicate long-term stability and response of valley glaciers to changes in mass transfer. *Journal of Glaciology* **30**: 199-211.

Fyfe, G., 1990. The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselkä I, southern Finland. *Boreas* **19**: 147-164.

Geirsdóttir, A., Hardardóttir, J., Sveinbjornsdóttir, A.E., 2000. Glacial extent and catastrophic meltwater events during the deglaciation of Southern Iceland. *Quaternary Science Reviews* **19**: 1749-1761.

Gemmell, A.M.D., 1973. The deglaciation of the Island of Arran. *Transactions of the Institute of British Geographers* **59**: 25-39.

Glasser, N.F., Hambrey, M.J., 2001. Styles of sedimentation beneath Svalbard valley glaciers under changing dynamic and thermal regimes. *Journal of the Geological Society of London* **158**: 697-707.

Glasser, N.F., Hambrey, M.J., 2003. Ice-marginal terrestrial landsystems: Svalbard polythermal glaciers. In: Evans, D.J.A. (ed.), *Glacial landsystems*. London: Arnold, 65-88.

Glasser, N.F., Bennett, M.R., 2004. Glacial erosional landforms: origins and significance for palaeoglaciology. *Progress in Physical Geography* **28**: 43-75.

Glasser, N.F., Hambrey, M.J., Crawford, K.R., Bennett, M.R., Huddart, D., 1998. The structural glaciology of Kongsvegen, Svalbard, and its role in landform genesis. *Journal of Glaciology* **44**: 136-148.

Glasser, N.F., Bennett, M.R., Huddart, D., 1999. Distribution of glaciofluvial sediment within and on the surface of a high arctic valley glacier: Marthabreen, Svalbard. *Earth Surface Processes and Landforms* **24**: 303-318.

Golledge, N.R., 2002. Glaci-tectonic deformation of proglacial lake sediments in the Cairngorm Mountains. *Scottish Journal of Geology* **38**: 127-136.

Golledge, N.R., 2003. A former ice-dammed lake in Glen Luibeg, Cairngorm Mountains, Scotland. *Quaternary Newsletter* **101**: 13-24.

Golledge, N.R., Hubbard, A., 2005. Evaluating Younger Dryas glacier reconstructions in part of the western Scottish Highlands: a combined empirical and theoretical approach. *Boreas* **34**: 274-286.

Gordon, J.E., 1981. Ice-scoured topography and its relationships to bedrock structure and ice movements in parts of northern Scotland and West Greenland. *Geografiska Annaler* **63A**: 331-344.

Gordon, J.E., Whalley, W.B., Gellatly, A.F., Vere, D.M., 1992. The formation of glacial flutes: assessment of models with evidence from Lyngsdalen, north Norway. *Quaternary Science Reviews* **11**: 709-731.

Graham, D.J., 2002. *Moraine mound formation during the Younger Dryas in Britain and the Neoglacial in Svalbard.* Unpublished PhD Thesis, 2 volumes, University of Wales, Aberystwyth.

Graham, D.J., Midgley, N.G., 2000a. Moraine-mound formation by englacial thrusting: the Younger Dryas moraines of Cwm Idwal, North Wales. In: Maltman, A.J., Hubbard, B. and Hambrey, M.J. (eds.), *Deformation of glacial materials.* Geological Society of London Special Publications 176, 321-336.

Graham, D.J., Midgley, N.G., 2000b. Graphical representation of particle shape using triangular diagrams: an Excel spreadsheet method. *Earth Surface Processes and Landforms* **25**: 1473-1477.

Gravenor, C.P., Kupsch, W.O., 1959. Ice disintegration features in western Canada. *Journal of Geology* **12**: 48-64.

Gray, J.M., 1982. The last glaciers (Loch Lomond Advance) in Snowdonia, N. Wales. *Geological Journal* **17**: 111-133.

Gray, J.M., 1992. Loch Etive kame and outwash terrace system. In: Walker, M.J.C., Gray, J.M. and Lowe, J.J. (eds.), *The Quaternary of the south-west Scottish Highlands: Field Guide.* Cambridge: Quaternary Research Association, 35-38.

Gray, J.M., 1997. Geomorphological change in the Scottish Late-glacial. In: Gordon, J.E. (ed.), *Reflections on the ice age in Scotland- an update on Quaternary studies.* Glasgow: Scottish Association of Geography Teachers, 95-104.

Gray, J.M., Brooks, C.L., 1972. The Loch Lomond Readvance moraines of Mull and Menteith. *Scottish Journal of Geology* **8**: 95-103.

Gray, J.M., Coxon, P., 1991. The Loch Lomond Stadial glaciation in Britain and Ireland. In: Ehlers, J., Gibbard, P.L. and Rose, J. (eds.), *Glacial deposits in Great Britain and Ireland.* Rotterdam: Balkema, 89-105.

Gripp, K., 1975. Hochsander - Satzmoräne - Endmoränenvertreter. *Zeitschrift für Geomorphologie* **19**: 490-496.

Gurnell, A.M., Clark, M.J. (eds.), 1987. *Glacio-fluvial sediment transfer - an alpine perspective.* Chichester: Wiley & Sons.

Gurnell, A.M., Edwards, P.J., Petts, G.E., Ward, J.V., 1999. A conceptual model for alpine proglacial river channel evolution under changing climatic conditions. *Catena* **38**: 223-242.

Häckel, H., 1999. *Meteorologie.* Stuttgart: Ulmer.

Hagen, J.O., Liestøl, O., Roland, E., Jørgensen, T., 1993. Glacier Atlas of Svalbard and Jan Mayen. *Norsk Polarinstitutt Meddelelser* **129**: 1-141.

Hall, A.M., Glasser, N.F., 2003. Reconstructing the basal thermal regime of an ice stream in a landscape of selective linear erosion: Glen Avon, Cairngorm Mountains, Scotland. *Boreas* **32**: 191-207.

Hall, A.M., Peacock, J.D., Connell, E.R., 2003. New data for the Last Glacial Maximum in Great Britain and Ireland: a Scottish perspective on the paper by Bowen et al. (2002). *Quaternary Science Reviews* **22**: 1551-1554.

Hambrey, M.J., Huddart, D., 1995. Englacial and proglacial glaciotectonic processes at the snout of a thermally complex glacier in Svalbard. *Journal of Quaternary Science* **10**: 313-326.

Hambrey, M.J., Glasser, N.F., 2002. Development of sediment-landform assemblages at maritime high-arctic glaciers. In: Hewitt, K., Byrne, M.-L., English, M. and Young, G. (eds.), *Landscapes of transition - landform assemblages and transformations in cold regions.* Dordrecht: Kluwer, 11-42.

Hambrey, M.J., Glasser, N.F., 2003. The role of folding and foliation development in the genesis of medial moraines: examples from Svalbard glaciers. *Journal of Geology* **111**: 471-485.

Hambrey, M.J., Ehrmann, W., 2004. Modification of sediment characteristics during glacial transport in high-alpine catchments: Mount Cook area, New Zealand. *Boreas* **33**: 300-318.

Hambrey, M.J., Huddart, D., Bennett, M.R., Glasser, N.F., 1997. Genesis of 'hummocky moraines' by thrusting of glacier ice: evidence from Svalbard and Britain. *Journal of the Geological Society of London* **154**: 623-632.

Hambrey, M.J., Bennett, M.R., Dowdeswell, J.A., Glasser, N.F., Huddart, D., 1999. Debris entrainment and transfer in polythermal valley glaciers. *Journal of Glaciology* **45**: 69-86.

Hambrey, M.J., Bennett, M.R., Glasser, N.F., Huddart, D., 2001. High-arctic glaciers as analogues for glacial landform development in Britain. *Geology Today* **17**: 24-28.

Hands, K.A., 2004. *Downwasting and supraglacial lake evolution on the debris-covered Ngozumpa Glacier, Khumbu Himal, Nepal.* Unpublished PhD Thesis, University of St Andrews.

Harker, A., 1901. Ice erosion in the Cuillin Hills, Skye. *Transactions of the Royal Society of Edinburgh* **40**: 221-252.

Harris, A.L., 1995. The nature and timing of orogenesis in the Scottish Highlands and the role of the Great Glen Fault. *Geological Association of Canada Special Paper* **41**: 65-79.

Harris, A.L., Johnson, M.R.W., 1991. Moine. In: Craig, G.Y. (ed.), *Geology of Scotland.* 3rd ed., London: The Geological Society, 87-123.

Hart, J.K., 1995. Recent drumlins, flutes and lineations at Vestari-Hagafellsjökull, Iceland. *Journal of Glaciology* **41**: 596-606.

Hart, J.K., Boulton, G.S., 1991. The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. *Quaternary Science Reviews* **10**: 335-350.

Hinchcliffe, S., Ballantyne, C.K., Walden, J., 1998. The structure and sedimentology of relict talus, Trotternish, northern Skye, Scotland. *Earth Surface Processes and Landforms* **23**: 545-560.

Hjelle, A., 1993. *Geology of Svalbard.* Oslo: Norsk Polarinstitutt.

Hodgson, D.M., 1982. *Hummocky and fluted moraines in part of north-west Scotland.* Unpublished PhD Thesis, University of Edinburgh.

Hodgson, D.M., 1986. A study of fluted moraines in the Torridon area, NW Scotland. *Journal of Quaternary Science* **1**: 109-118.

Hubbard, A., 1999. High-resolution modeling of the advance of the Younger Dryas Ice Sheet and its climate in Scotland. *Quaternary Research* **52**: 27-43.

Hubbard, B., Glasser, N.F., 2005. *Field techniques in glaciology and glacial geomorphology.* Chichester: Wiley & Sons.

Hubbard, B., Glasser, N.F., Hambrey, M.J., Etienne, J.L., 2004. A sedimentological and isotopic study of the origin of supraglacial debris bands: Kongsfjorden, Svalbard. *Journal of Glaciology* **50**: 157-170.

Hughes, P.D., 2002. Loch Lomond Stadial glaciers in the Aran and Arenig Mountains, North Wales, Great Britain. *Geological Journal* **37**: 9-15.

Humlum, O., 1985. Genesis of an imbricate push moraine, Höfdabrekkujökull, Iceland. *Journal of Geology* **93**: 185-195.

Humlum, O., 2002. Modelling late 20th-century precipitation in Nordenskiöld Land, Svalbard, by geomorphic means. *Norsk Geografisk Tidsskrift* **56**: 96-103.

Humlum, O., 2005. Holocene permafrost aggradation in Svalbard. *Geological Society of London Special Publications* **242**: 119-130.

Humlum, O., Instanes, A., Sollid, J.L., 2003. Permafrost in Svalbard: a review of research history, climatic background and engineering challenges. *Polar Research* **22**: 191-215.

Isarin, R.F.B., Bohncke, S.J.P., 1999. Mean July temperatures during the Younger Dryas in north-western and central Europe inferred from climate indicator plant species. *Quaternary Research* **51**: 158-173.

Isarin, R.F.B., Renssen, H., 1999. Reconstructing and modelling Late Weichselian climates: the Younger Dryas in Europe as a case study. *Earth-Science Reviews* **48**: 1-38.

Isarin, R.F.B., Renssen, H., Vandenberghe, J., 1998. The impact of the North Atlantic ocean on the Younger Dryas climate in north-western and central Europe. *Journal of Quaternary Science* **13**: 447-453.

Ivy-Ochs, S., Schlüchter, C., Kubik, P.W., Synal, H.-A., Beer, J., Kerschner, H., 1996. The exposure age of an Egesen moraine at Julier Pass, Switzerland, measured with cosmogenic radionuclides ^{10}Be, ^{26}Al and ^{36}Cl. *Eclogae Geologicae Helvetiae* **89**: 1049-1063.

Ivy-Ochs, S., Schlüchter, C., Kubik, P.W., Denton, G.H., 1999. Moraine exposure dates imply synchronous Younger Dryas glacier advances in the European Alps and in the southern Alps of New Zealand. *Geografiska Annaler* **81A**: 313-323.

Iwata, S., Watanabe, O., Fushimi, H., 1980. Surface morphology and in the ablation area of the Khumbu Glacier. *Seppyo (Journal of the Japanese Society of Snow and Ice)* **41**: 9-17.

Jarman, D., Lukas, S., 2005. Ben Hee. In: Cooper, R. (ed.), *Mass movements in Britain. Geological Conservation Review Series*, in press.

Johnson, M.D., Clayton, L., 2003. Supraglacial landsystems in lowland terrain. In: Evans, D.J.A. (ed.), *Glacial landsystems*. London: Arnold, 228-258.

Johnson, M.R.W., Kelley, S.P., Oliver, G.J.H., Winter, D.A., 1985. Thermal effects and timing of thrusting in the Moine thrust zone. *Journal of the Geological Society of London* **142**: 863-873.

Johnstone, G.S., Mykura, W., 1989. *British Regional Geology - The Northern Highlands of Scotland.* 4th ed., London: HMSO.

Jones, A.P., Tucker, M.E., Hart, J.K., 1999. Guidelines and recommendations. In: Jones, A.P., Tucker, M.E. and Hart, J.K. (eds.), *The description and analysis of Quaternary stratigraphic field sections*. London: Quaternary Research Association, 27-62.

Juschus, O., Preusser, F., Melles, M., Radtke, U., 2005. Applying SAR-IRSL methodology for dating fine-grained sediments from Lake El'gygytgyn, north-eastern Siberia. *Quaternary Science Reviews (Quaternary Geochronology)* **submitted**.

Justad, G.H., 1997. *Morfologiske og sedimentologiske studier av et iskontakt system ved Longyearbreen, Svalbard.* cand. scient. Thesis, University of Tromsø.

Kaplan, M.R., Kurz, M.D., Ackert Jr, R.P., Singer, B.S., Douglass, D.C., 2004. Cosmogenic nuclide chronology of millennial-scale glacial advances during O-isotope stage 2 in Patagonia. *Geological Society of America Bulletin* **116**: 308-321.

Karte, J., 1979. Räumliche Abgrenzung und regionale Differenzierung des Periglaziärs. *Bochumer Geographische Arbeiten* **35**: 1-211.

Kaser, G., Osmaston, H.A., 2002. *Tropical glaciers*. Cambridge: Cambridge University Press.

Kelletat, D., 1970. Zum Problem der Verbreitung, des Alters und der Bildungsdauer alter (inaktiver) Periglazialerscheinungen im Schottischen Hochland. *Zeitschrift für Geomorphologie Supplement Band* **14**: 510-519.

Kelly, M.A., Kubik, P.W., von Blanckenburg, F., Schlüchter, C., 2004a. Surface exposure dating of the Great Aletsch Glacier Egesen moraine system, western Swiss Alps, using cosmogenic nuclide ^{10}Be. *Journal of Quaternary Science* **19**: 431-441.

Kelly, M.A., Buoncristiani, J.F., Schlüchter, C., 2004b. A reconstruction of the last glacial maximum (LGM) ice-surface geometry in the western Swiss Alps and contiguous Alpine regions in Italy and France. *Eclogae Geologicae Helvetiae* **97**: 57-75.

Kerschner, H., Kaser, G., Sailer, R., 2000. Alpine Younger Dryas glaciers as palaeo-precipitation gauges. *Annals of Glaciology* **31**: 80-84.

King, L., Volk, M., 1994. Glaziologie und Glazialmorphologie des Liefde- und Bockfjordgebietes, NW-Spitzbergen. *Zeitschrift für Geomorphologie Supplement Band* **97**: 145-159.

Kirkbride, M., 1995. Ice flow vectors on the debris-mantled Tasman glacier, 1957- 1986. *Geografiska Annaler Series a-Physical Geography* **77A**: 147-157.

Kirkbride, M., 2000. Ice-marginal geomorphology and Holocene expansion of debris-covered Tasman Glacier, New Zealand. In: Nakawo, M., Raymond, C.F. and Fountain, A. (eds.), *Debris-Covered Glaciers*. Wallingford: IAHS-AISH Publication 264, 211-218.

Kirkbride, M.P., Warren, C.R., 1999. Tasman Glacier, New Zealand: 20th-century thinning and predicted calving retreat. *Global and Planetary Change* **22**: 11-28.

Kjær, K.H., Krüger, J., 2001. The final phase of dead-ice development: processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology* **48**: 935-952.

Kleman, J., 1992. The palimpsest glacial landscape in northwestern Sweden. *Geografiska Annaler* **74A**: 305-325.

Kleman, J., 1994. Preservation of landforms under ice sheets and ice caps. *Geomorphology* **9**: 19-32.

Knight, P.G., 1997. The basal ice layer of glaciers and ice sheets. *Quaternary Science Reviews* **16**: 975-993.

Knighton, A.D., 1998. *Fluvial forms and processes - a new perspective*. London: Arnold.

Kronberg, P., 1984. *Photogeologie. Eine Einführung in die Grundlagen und Methoden der geologischen Auswertung von Luftbildern*. Stuttgart: Enke.

Krüger, J., 1979. Structures and textures in till indicating subglacial deposition. *Boreas* **8**: 323-340.

Krüger, J., 1984. Clasts with stoss-lee form in lodgement tills: a discussion. *Journal of Glaciology* **30**: 241-243.

Krüger, J., 1993. Moraine-ridge formation along a stationary ice front in Iceland. *Boreas* **22**: 101-109.

Krüger, J., 1994. Glacial processes, sediments, landforms, and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* **21**: 1-233.

Krüger, J., 1995. Origin, chronology and climatological significance of annual-moraine ridges at Myrdalsjökull, Iceland. *The Holocene* **5**: 420-427.

Krüger, J., 1996. Moraine ridges formed from subglacial frozen-on sediment slabs and their differentiation from push moraines. *Boreas* **25**: 57-63.

Krüger, J., 1997. Development of minor outwash fans at Kötlujökull, Iceland. *Quaternary Science Reviews* **16**: 649-659.

Krüger, J., Thomsen, H.H., 1984. Morphology, stratigraphy, and genesis of small drumlins in front of the glacier Myrdalsjökull, south Iceland. *Journal of Glaciology* **30**: 94-105.

Krüger, J., Aber, J.S., 1999. Formation of supraglacial sediment accumulations on Kötlujökull, Iceland. *Journal of Glaciology* **45**: 400-402.

Krüger, J., Kjær, K.H., 1999. A data chart for field description and genetic interpretation of glacial diamicts and associated sediments - with examples from Greenland, Iceland, and Denmark. *Boreas* **28**: 386-402.

Krüger, J., Kjær, K.H., 2000. De-icing procession of ice-cored moraines in a humid, subpolar climate, Kötlujökull, Iceland. *The Holocene* **10**: 737-747.

Krzyszkowski, D., 2002. Sedimentary successions in ice-marginal fans of the Late Saalian glaciation, southwestern Poland. *Sedimentary Geology* **149**: 93-109.

Krzyszkowski, D., Zielinski, T., 2002. The Pleistocene end moraine fans: controls on their sedimentation and location. *Sedimentary Geology* **149**: 73-92.

Landvik, J.Y., Mangerud, J., Salvigsen, O., 1988. Glacial history and permafrost in the Svalbard area. *Permafrost. Fifth International Conference*. Trondheim: Tapir, 194-198.

Landvik, J.Y., Bondevik, S., Elyerhøei, A., Fjeldskaar, W., Mangerud, J., Salvigsen, O., Siegert, M.J., Svendsen, J.-I., Vorren, T.O., 1998. The Last Glacial Maximum of Svalbard and the Barents Sea area: Ice sheet extent and configuration. *Quaternary Science Reviews* **17**: 43-75.

Lawson, D.E., 1981. Distinguishing characteristics of diamictons at the margin of the Matanuska glacier, Alaska. *Annals of Glaciology* **2**: 78-84.

Lawson, D.E., 1982. Mobilization, movement and deposition of active subaerial sediment flows, Matanuska Glacier, Alaska. *Journal of Geology* **90**: 279-300.

Lawson, D.E., 1988. Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits. In: Goldthwait, R.P. and Matsch, C.L. (eds.), *Genetic classification of glacigenic deposits*. Rotterdam: Balkema, 147-169.

Lawson, T.J., 1983. *Quaternary Geomorphology of the Assynt area, N.W. Scotland*. Unpublished PhD Thesis, University of Edinburgh.

Lawson, T.J., 1986. Loch Lomond Advance glaciers in Assynt, Sutherland, and their palaeoclimatic implications. *Scottish Journal of Geology* **22**: 289-298.

Lawson, T.J., 1990. Former ice movement in Assynt, Sutherland, as shown by the distribution of glacial erratics. *Scottish Journal of Geology* **26**: 25-32.

Lawson, T.J., 1996. Glacial striae and former ice movement: the evidence from Assynt, Sutherland. *Scottish Journal of Geology* **32**: 59-65.

Lawson, T.J., Atkinson, T.C., 1995. Quaternary chronology. In: Lawson, T.J. (ed.), *The Quaternary of Assynt and Coigach: Field Guide*. Cambridge: Quaternary Research Association, 12-18.

Lehmkuhl, F., 1998. Quaternary glaciations in central and western Mongolia. *Quaternary Proceedings* **6**: 153-167.

Lemmen, D.S., England, J., 1992. Multiple glaciations and sea level changes, northern Ellesmere Island, high arctic Canada. *Boreas* **21**: 137-152.

Leonard, E.M., 1989. Climatic change in the Colordao Rocky Mountains: estimates based on modern climate at late Pleistocene equilibrium lines. *Arctic and Alpine Research* **21**: 245-255.

Liestøl, O., 1969. Glacier surges in West Spitsbergen. *Canadian Journal of Earth Sciences* **6**: 895-897.

Liestøl, O., 1993. Glaciers of Europe - glaciers of Svalbard, Norway. In: Williams, R.S. and Ferriggo, J.G. (eds.), *Satellite image atlas of glaciers of the world. USGS Professional Papers*, E127-E151.

Lillesand, T.M., Kiefer, R.W., 2000. *Remote sensing and image interpretation*. New York: Wiley & Sons.

Lindsay, R.A., Charman, D.J., Everingham, F., O'Reilly, R.M., Palmer, M.A., Rowell, T.A., Stroud, D.A., 1988. *The flow country - the peatlands of Caithness and Sutherland*. Peterborough: Nature Conservancy Council.

Linton, D.L., 1963. The forms of glacial erosion. *Transactions of the Institute of British Geographers* **33**: 1-28.

Lønne, I., 1993. Physical signatures of ice advance in a Younger Dryas ice-contact delta, Troms, northern Norway: implications for glacier-terminus history. *Boreas* **22**: 59-70.

Lønne, I., 2005. Faint traces of high Arctic glaciations: an early Holocene ice-front fluctuation in Bolterdalen, Svalbard. *Boreas* **34**: 308–323.

Lowe, J.J., Walker, M.J.C., 1997. *Reconstructing Quaternary environments*. 2nd ed., London: Longman.

Lukas, S., 2003. Scottish Landform Example No. 31. The moraines around the Pass of Drumochter. *Scottish Geographical Journal* **119**: 383-393.

Lukas, S., 2004a. The pattern of deglaciation around the Pass of Drumochter. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the Central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 68-77.

Lukas, S., 2004b. Landforms around Dalnaspidal Lodge. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the Central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 174-179.

Lukas, S., 2005. A test of the englacial thrusting hypothesis of 'hummocky' moraine formation - case studies from the north-west Highlands, Scotland. *Boreas* **34**: 287-307.

Lukas, S., Merritt, J.W., 2004. Evidence for a former ice-dammed lake in Coire Mhic-sith. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the Central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 149-158.

Lukas, S., Nicholson, L.I., Ross, F.H., Humlum, O., 2005a. Formation, meltout processes and landscape alteration of high-arctic ice-cored moraines - examples from Nordenskiöld Land, central Spitsbergen. *Polar Geography* **29**: 79-109.

Lukas, S., Spencer, J.Q.G., Robinson, R.A.J., Finch, A.A., 2005b. Problems of luminescence dating of Late Quaternary glacial sediments in the NW Scottish Highlands. *Quaternary Science Reviews (Quaternary Geochronology)* **submitted**.

Lyså, A., Lønne, I., 2001. Moraine development at a small High-Arctic valley glacier: Rieperbreen, Svalbard. *Journal of Quaternary Science* **16**: 519-529.

Major, H., Haremo, P., Dallmann, W.K., Andresen, A., 2000. *Geological map Svalbard 1:100,000 Spitsbergen - Sheet C9Q Adventdalen*. Tromsø: Norsk Polarinstitutt.

Mann, M.E., 2002. The value of multiple proxies. *Science* **297**: 1481-1482.

Marren, P.M., 2002. Glacier margin fluctuations, Skaftafellsjökull, Iceland: Implications for sandur evolution. *Boreas* **31**: 75-81.

Matthews, J.A., Dahl, S.O., Nesje, A., Berrisford, M.S., Andersson, C., 2000. Holocene glacier variations in central Jotunheimen, southern Norway, based on distal glaciolacustrine sediment cores. *Quaternary Science Reviews* **19**: 1625-1647.

McAvaney, B.J., 75 others, 2001. Model evaluation. In: Houghton, J.T. and 6 others (eds.), *Climate change 2001: The scientific basis*. Cambridge: Cambridge University Press, 473-523.

McCarroll, D., Rijsdijk, K.F., 2003. Deformation styles as a key for interpreting glacial depositional environments. *Journal of Quaternary Science* **18**: 473-489.

McCarroll, D., Ballantyne, C.K., Nesje, A., Dahl, S.O., 1995. Nunataks of the last ice sheet in northwest Scotland. *Boreas* **24**: 305-323.

McDougall, D.A., 2001. The geomorphological impact of Loch Lomond (Younger Dryas) Stadial plateau icefields in the central Lake District, northwest England. *Journal of Quaternary Science* **16**: 531-544.

Meierding, T.C., 1982. Late Pleistocene glacial equilibrium-line in the Colordao Front range: a comparison of methods. *Quaternary Research* **18**: 289-310.

Merritt, J.W., Coope, G.R., Taylor, B.J., Walker, M.J.C., 1990. Late Devensian organic deposits beneath till in the Teith Valley, Perthshire. *Scottish Journal of Geology* **26**: 15-24.

Midgley, N.G., 2001. *Moraine-mound development in Britain and Svalbard - The development of 'hummocky moraine'*. Unpublished PhD Thesis, Liverpool John Moores University.

Mitchell, W.A., Lukas, S., 2004. Drumochter Mast. In: Lukas, S., Merritt, J.W. and Mitchell, W.A. (eds.), *The Quaternary of the Central Grampian Highlands: Field Guide*. London: Quaternary Research Association, 163-170.

Moar, N.T., 1969. A radiocarbon-dated pollen diagram from north-west Scotland. *New Phytologist* **68**: 209-214.

Morrocco, S.M., 2003. Optically-stimulated luminescence dating of plateau surface erosion and windblown sand resedimentation in the Highlands of Scotland. *Quaternary Newsletter* **101**: 75-77.

Mottershead, D.N., 1978. High altitude solifluction and post-glacial vegetation, Arkle, Sutherland. *Transactions of the Botanical Society of Edinburgh* **43**: 17-24.

Mottershead, D.N., White, I.D., 1969. Some solifluction terraces in Sutherland. *Transactions of the Botanical Society of Edinburgh* **40**: 604-620.

Murray, A.S., Wintle, A.G., 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. *Radiation Measurements* **32**: 57-73.

Murray, A.S., Wintle, A.G., 2003. The single aliquot regenerative dose protocol: potential for improvements in reliability. *Radiation Measurements* **37**: 377-381.

Murray, N.A., Bonsall, C., Sutherland, D.G., Lawson, T.J., Kitchener, A.C., 1993. Further radiocarbon determinations on reindeer remains of Middle and Late Devensian age from the Creag nan Uamh caves, Assynt, NW Scotland. *Quaternary Newsletter* **70**: 1-10.

Nesje, A., 1992. Topographical effects on the equilibrium line altitude on glaciers. *GeoJournal* **27**: 383-391.

Nesje, A., Dahl, S.O., 2000. *Glaciers and environmental change*. London: Arnold.

Nesje, A., Johannessen, T., Birks, H.J.B., 1995. Briksdalsbreen, western Norway: climatic effects on the terminal response of a temperate glacier between AD 1901 and 1994. *The Holocene* **5**: 343-347.

Oerlemans, J., 2001. *Glaciers and climate change*. Rotterdam: Balkema.

Ohmura, A., Kasser, P., Funk, M., 1992. Climate at the equilibrium line of glaciers. *Journal of Glaciology* **38**: 397-411.

Orvis, K.H., Horn, S.P., 2000. Quaternary glaciers and climate on Cerro Chirripo, Costa Rica. *Quaternary Research* **54**: 24-37.

Osmaston, H.A., 2005. Estimates of glacier equilibrium line altitudes by the Area x Altitude, the Area x Altitude Balance Ratio and the Area x Altitude Balance Index methods and their validation. *Quaternary International* **138-139**: 22-31.

Østrem, G., Dale Selvig, K., Tandberg, K., 1988. *Atlas over breer i Sør-Norge. - NVE Hydrologisk avdeling, Meddelse, 61, 1-247*

Owen, L.A., Derbyshire, E., 1989. The Karakorum glacial depositional system. *Zeitschrift für Geomorphologie Supplement Band* **76**: 33-73.

Park, R.G., Stewart, A.D., Wright, D.T., 2002. The Hebridean terrane. In: Trewin, N.H. (ed.), *The geology of Scotland*. London: Geological Society of London, 45-80.

Peach, B.N., Horne, J., 1892. The ice-shed in the North-West Highlands during the maximum glaciation. *Report of the British Association for 1892*: 720.

Peach, B.N., Horne, J., Gunn, W., Clough, C.T., Hinxman, L.W., Teall, J.J.H., 1907. *The geological structure of the northwest Highlands of Scotland*. Memoir of the Geological Survey, UK. Edinburgh.

Peacock, J.D., 1970. Some aspects on the glacial geology of west Inverness-shire. *Bulletin of the Geological Survey of Great Britain* **33**: 43-56.

Peacock, J.D., Cornish, R. (eds.), 1989. *Glen Roy area: Field Guide*. Cambridge: Quaternary Research Association.

Pelto, M.S., 2000. Mass balance of adjacent debris-covered and clean glacier ice in the North Cascades, Washington. In: Nakawo, M., Raymond, C.F. and Fountain, A. (eds.), *Debris-Covered Glaciers*. Wallingford: IAHS-AISH Publication 264, 35-42.

Penck, A., Brückner, E., 1901/1909. *Die Alpen im Eiszeitalter*. Leipzig: Tauchnitz.

Pennington, W., 1977. Lake sediments and the Lateglacial environment in northern Scotland. In: Gray, J.M. and Lowe, J.J. (eds.), *Studies in the Scottish Lateglacial environment*. Oxford: Pergamon Press, 119-141.

Pennington, W., Haworth, E.Y., Bonny, A.P., Lishman, J.P., 1972. Lake sediments in northern Scotland. *Philosophical Transactions of the Royal Society of London* **B264**: 191-294.

Phillips, E.R., Evans, D.J.A., Auton, C.A., 2002. Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. *Sedimentary Geology* **149**: 157-182.

Porter, S.C., 1975. Equilibrium-line altitudes of Late Quaternary glaciers in the southern Alps, New Zealand. *Quaternary Research* **5**: 27-47.

Preusser, F., 1999. Lumineszenzdatierung fluviatiler Sedimente - Beispiele aus der Schweiz und Norddeutschland. *Kölner Forum für Geologie und Paläontologie* **3**: 1-62.

Preusser, F., 2003. IRSL dating of K-rich feldspars using the SAR protocol: comparison with independent age control. *Ancient TL* **21**: 17-23.

Preusser, F., 2004. Towards a chronology of the Late Pleistocene in the northern Alpine Foreland. *Boreas* **33**: 195-210.

Preusser, F., Radies, D., Matter, A., 2002. A 160,000-year record of dune development and atmospheric circulation in southern Arabia. *Science* **296**: 2018-2020.

Proctor, C.J., Baker, A., Barnes, W.L., 2002. A three thousand year record of North Atlantic climate. *Climate Dynamics* **19**: 449-454.

Proctor, C.J., Baker, A., Barnes, W.L., Gilmour, M.A., 2000. A thousand year speleothem proxy record of North Atlantic climate from Scotland. *Climate Dynamics* **16**: 815-820.

Rahmstorf, S., 2002. Ocean circulation and climate during the past 120,000 years. *Nature* **419**: 207-214.

Rapson, S.C., 1985. Minimum age of corrie moraine ridges in the Cairngorm Mountains, Scotland. *Boreas* **14**: 155-159.

Rastas, J., Seppälä, M., 1981. Rock jointing and abrasion forms on roches moutonnées, SW Finland. *Annals of Glaciology* **2**: 159-163.

Rea, B.R., Evans, D.J.A., 1996. Scottish Landform Examples: 14 - Landscapes of areal scoruing in N.W. Scotland. *Scottish Geographical Magazine* **112**: 47-50.

Rea, B.R., Evans, D.J.A., 2003. Plateau icefield landsystems. In: Evans, D.J.A. (ed.), *Glacial Landsystems*. London: Arnold, 407-431.

Rea, B.R., Whalley, W.B., Evans, D.J.A., Gordon, J.E., McDougall, D.A., 1998. Plateau icefields: geomorphology and dynamics. *Quaternary Proceedings* **6**: 35-54.

Rea, B.R., Whalley, W.B., Dixon, T., Gordon, J.E., 1999. Plateau icefields as contributing areas to valley glaciers and the potential impact on reconstructed ELAs: a case study from the Lyngen Alps, north Norway. *Annals of Glaciology* **28**: 97-102.

Read, H.H., 1931. *The geology of central Sutherland*. Edinburgh: HMSO.

Read, H.H., Phemister, J., Ross, G., 1926. *The Geology of Strath Oykell and lower Loch Shin*. Edinburgh: HMSO.

Reading, H.G. (ed.), 1996. *Sedimentary environments: processes, facies and stratigraphy*. 3rd ed., Oxford: Blackwell.

Renssen, H., Podzun, R., Vandenberghe, J., Isarin, R.F.B., Jacob, D., 2001. Simulation of the Younger Dryas climate in Europe using a regional climate model nested in an AGCM: Preliminary results. *Global and Planetary Change* **30**: 41-57.

Richards, B.W.M., Benn, D.I., Owen, L.A., Rhodes, E.J., Spencer, J.Q.G., 2000. Timing of late Quaternary glaciations south of Mount Everest in the Khmubu Himal, Nepal. *Geological Society of America Bulletin* **112**: 1621-1632.

Rijsdijk, K.F., 2001. Density-drive deformation structures in glacigenic consolidated diamicts: Examples from Traeth y Mwnt, Cardiganshire, Wales U.K. *Journal of Sedimentary Research* **71**: 122-135.

Robertson-Rintoul, M.S.E., 1986. A quantitative soil-stratigraphic approach to the correlation and dating of post-glacial river terraces in Glen Feshie, Western Cairngorms. *Earth Surface Processes and Landforms* **11**: 605-617.

Robinson, M., Ballantyne, C.K., 1979. Evidence for a glacial readvance pre-dating the Loch Lomond Advance in Wester Ross. *Scottish Journal of Geology* **15**: 271-277.

Rose, J., 1987. Drumlins as part of a bedform continuum. In: Menzies, J. and Rose, J. (eds.), *Drumlin Symposium*. Rotterdam: Balkema, 103-116.

Rose, J., Lowe, J.J., Switsur, R., 1988. A radiocarbon date on plant detritus beneath till from the type area of the Loch Lomond Readvance. *Scottish Journal of Geology* **24**: 113-124.

Salt, K.E., Ballantyne, C.K., 1997. The structure and sedimentology of relict talus, Knockan, Assynt. *Scottish Geographical Magazine* **113**: 82-89.

Schellmann, G., Radtke, U., 2004. A revised morpho- and chronostratigraphy of the Late and Middle Pleistocene coral reef terraces on Southern Barbados. *Earth-Science Reviews* **64**: 157-187.

Sellier, D., Lawson, T.J., 1998. A complex slope failure on Beinn nan Cnaimhseag, Assynt, Sutherland. *Scottish Geographical Magazine* **114**: 85-93.

Sharp, M., 1984. Annual moraine ridges at Skàlafellsjökull, south-east Iceland. *Journal of Glaciology* **30**: 82-93.

Sharp, M., Dowdeswell, J.A., Gemmell, J.C., 1989. Reconstructing past glacier dynamics and erosion from glacial geomorphic evidence: Snowdon, North Wales. *Journal of Quaternary Science* **4**: 115-130.

Shroder, J.F., Bishop, M.P., Copland, L., Sloan, V.F., 2000. Debris-covered glaciers and rock glaciers in the Nanga Parbat Himalaya, Pakistan. *Geografiska Annaler* **82A**: 17-31.

Sigmond, E.M.O., Gustavson, M., Roberts, D., 1984. *Berggrunnskart over Norge - M 1: 1 million*. Oslo: Norges Geologiske Undersokelske.

Simpson, J.B., 1933. The Lateglacial readvance moraines of the Highland border west of the River Tay. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **57**: 633-645.

Sissons, J.B., 1967. *The evolution of Scotland's scenery*. Edinburgh: Oliver & Boyd.

Sissons, J.B., 1972. The last glaciers in part of the south east Grampians. *Scottish Geographical Magazine* **88**: 168-181.

Sissons, J.B., 1974. A lateglacial ice cap in the central Grampians, Scotland. *Transactions of the Institute of British Geographers* **62**: 95-114.

Sissons, J.B., 1976. *The Geomorphology of the British Isles: Scotland*. London: Methuen.

Sissons, J.B., 1977. The Loch Lomond Readvance in the northern mainland of Scotland. In: Gray, J.M. and Lowe, J.J. (eds.), *Studies in the Scottish Lateglacial environment*. Oxford: Pergamon Press, 45-59.

Sissons, J.B., 1979a. The Loch Lomond Stadial in the British Isles. *Nature* **280**: 199-203.

Sissons, J.B., 1979b. The Loch Lomond Advance in the Cairngorm Mountains. *Scottish Geographical Magazine* **95**: 66-82.

Sissons, J.B., 1980. Palaeoclimatic inferences from Loch Lomond Advance glaciers. In: Lowe, J.J., Gray, J.M. and Robinson, J.E. (eds.), *Studies in the Lateglacial of North-west Europe*. Oxford: Pergamon Press, 31-44.

Sissons, J.B., 1982. A former ice-dammed lake and associated glacier limits in the Achnasheen area, central Ross-shire. *Transactions of the Institute of British Geographers* **7**: 98-116.

Sissons, J.B., Grant, A.J.H., 1972. The last glaciers in the Lochnagar area, Aberdeenshire. *Scottish Journal of Geology* **8**: 85-93.

Sissons, J.B., Walker, M.J.C., 1974. Lateglacial site in the central Grampian Highlands. *Nature* **249**: 822-824.

Sissons, J.B., Sutherland, D.G., 1976. Climatic inferences from former glaciers in the south-east Grampian Highlands. *Journal of Glaciology* **17**: 325-346.

Sissons, J.B., Dawson, A.G., 1981. Former sea-levels and ice limits in part of Wester Ross, northwest Scotland. *Proceedings of the Geologist's Association* **92**: 115-124.

Skinner, B.J., Porter, S.C., 1987. *Physical geology*. Chichester: Wiley & Sons.

Sletten, K., Lyså, A., Lønne, I., 2001. Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard. *Boreas* **30**: 272-284.

Small, R.J., 1983. Lateral moraines of Glacier de Tsidjiore Nouve: form, development and implications. *Journal of Glaciology* **29**: 250-259.

Small, R.J., 1987. Moraine sediment budgets. In: Gurnell, A.M. and Clark, M.J. (eds.), *Glacio-fluvial sediment transfer: an Alpine perspective*. Chichester: Wiley & Sons, 165-197.

Sneed, E.D., Folk, R.L., 1958. Pebbles in the lower Colorado River, Texas, a study in particle morphogenesis. *Journal of Geology* **66**: 114-150.

Soper, N.J., Harris, A.L., Strachan, R.A., 1998. Tectonostratigraphy of the Moine Supergroup: a synthesis. *Journal of the Geological Society of London* **155**: 13-24.

Sørbel, L., Tolgensbakk, J., Hagen, J.O., Høgvard, K., 2001. *Geomorphological and Quaternary geological map of Svalbard 1:100,000 - Sheet C9Q Adventdalen. Explanatory text*. Tromsø: Norsk Polarinstitutt.

Spedding, N., 2000. Hydrological controls on sediment transport pathways: implications for debris-covered glaciers. In: Nakawo, M., Raymond, C.F. and Fountain, A. (eds.), *Debris-Covered Glaciers*. Wallingford: IAHS-AISH Publication 264, 133-142.

Spedding, N., Evans, D.J.A., 2002. Sediments and landforms at Kvíárjökull, southeast Iceland: A reappraisal of the glaciated valley landsystem. *Sedimentary Geology* **149**: 21-42.

Spencer, J.Q.G., Owen, L.A., 2004. Optically stimulated luminescence dating of Late Quaternary glaciogenic sediments in the upper Hunza valley: validating the timing of glaciation and assessing dating methods. *Quaternary Science Reviews* **23**: 175-191.

Stoker, M., Bradwell, T., 2005. The Minch palaeo-ice stream, NW sector of the British-Irish ice sheet. *Journal of the Geological Society of London* **163**: 423-428.

Stokes, C.R., Clark, C.D., 2001. Palaeo-ice streams. *Quaternary Science Reviews* **20**: 1437-1457.

Strachan, R.A., Harris, A.L., Fettes, D.J., Smith, M., 2002. The Highland and Grampian terranes. In: Trewin, N.H. (ed.), *The geology of Scotland*. London: Geological Society of London, 81-148.

Sugden, D.E., 1968. The selectivity of glacial erosion in the Cairngorm Mountains, Scotland. *Transactions of the Institute of British Geographers* **45**: 75-92.

Sugden, D.E., 1970. Landforms of deglaciation in the Cairngorm Mountains, Scotland. *Transactions of the Institute of British Geographers* **51**: 201-219.

Sugden, D.E., 1974. Deglaciation of the Cairngorms and its wider implications. In: Caseldine, C.J. and Mitchell, W.A. (eds.), *Problems of the deglaciation of Scotland*. St Andrews: Department of Geography, 17-28.

Sugden, D.E., 1980. The Loch Lomond Advance in the Cairngorms (a reply to J.B. Sissons). *Scottish Geographical Magazine* **96**: 18-19.

Sugden, D.E., Watts, S.H., 1977. Tors, felsenmeer and glaciation in northern Cumberland Peninsula, Baffin Island. *Canadian Journal of Earth Sciences* **14**: 2817-2823.

Sutherland, D.G., 1984a. The Quaternary deposits and landforms of Scotland and the neighbouring shelves: a review. *Quaternary Science Reviews* **3**: 157-254.

Sutherland, D.G., 1984b. Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond Stadial. *Quaternary Science Reviews* **3**: 291-309.

Svendsen, J.-I., Mangerud, J., 1997. Holocene glacial and climatic variations on Spitsbergen, Svalbard. *The Holocene* **7**.

Tarasov, L., Peltier, W.R., 2005. Arctic freshwater forcing of the Younger Dryas cold reversal. *Nature* **435**: 662-665.

Thomas, G.S.P., Montague, E., 1997. The morphology, stratigraphy and sedimentology of the Carstairs esker, Scotland, U.K. *Quaternary Science Reviews* **16**: 661-674.

Thorp, P.W., 1986. A mountain icefield of Loch Lomond Stadial age, western Grampians, Scotland. *Boreas* **15**: 83-97.

Thorp, P.W., 1991. The glaciation and glacial deposits of the western Grampians. In: Ehlers, J., Gibbard, P.L. and Rose, J. (eds.), *Glacial deposits in Great Britain and Ireland*. Rotterdam: Balkema, 137-149.

Tipping, R., 1989. Palynological evidence for the extent of the Loch Lomond Readvance in the Awe valley and adjacent areas, SE Highlands. *Scottish Journal of Geology* **25**: 325-337.

Tipping, R. (ed.), 2003. *The Quaternary of Glen Affric and Kintail: Field Guide*. London: Quaternary Research Association.

Tipping, R., Carter, S., Haggart, B.A., 1994. Late Quaternary valley floor evolution in the lower reaches of Carradale, Kintyre. *Scottish Journal of Geology* **30**: 131-145.

Tipping, R., Tisdall, E., Davies, A., 2003. Delayed deglaciation from the Loch Lomond Readvance: radiocarbon dated sequences from West Glen Affric. In: Tipping, R. (ed.), *The Quaternary of Glen Affric and Kintail: Field Guide*. London: Quaternary Research Association, 85-92.

Tonning, T., 1996. *Bruk av eit geografisk informasjonssystem på glasiologiske problemstilingar: døme fra Longyearbreen og Larsbreen på Svalbard*. cand. scient. Thesis, University of Olso.

Torsnes, I., Rye, N., Nesje, A., 1993. Modern and Little Ice Age equilibrium-line altitudes on outlet valley glaciers from jostedalsbreen, western Norway: an evaluation of different approaches to their calculation. *Arctic and Alpine Research* **25**: 106-116.

van Breemen, O., Aftalion, M., Johnson, M.R.W., 1979. Age of the Loch Borrolan Complex, Assynt and late movements along the Moine Thrust Zone. *Journal of the Geological Society of London* **16**: 489-495.

van der Meer, J.J.M., 1983. A recent drumlin with fluted surface in the Swiss Alps. In: Evenson, E.B. (ed.), *Tills and related deposits*. Rotterdam: Balkema, 105-109.

van der Meer, J.J.M., Kjær, K.H., Krüger, J., 2001. Subglacial water-escape structures and till structures, Sléttjökull, Iceland. *Journal of Quaternary Science* **14**: 191-205.

van der Wateren, F.M., 1995. Structural geology and sedimentology of push moraines - processes of soft sediment deformation in a glacial environment and the distribution of glaciotectonic styles. *Mededelingen Rijks Geologische Dienst* **54**: 1-168.

van der Wateren, F.M., 1999. Structural geology and sedimentology of the Heiligenhafen till section, Northern Germany. *Quaternary Science Reviews* **18**: 1625-1639.

van der Wateren, F.M., Kluiving, S.J., Bartek, L.R., 2000. Kinematic indicators of subglacial shearing. In: Maltman, A.J., Hubbard, B. and Hambrey, M.J. (eds.), *Deformation of glacial materials. Geological Society of London Special Publication 176*, 259-278.

Vasari, Y., 1977. Radiocarbon dating of the Lateglacial and early Flandrian vegetational succession in the Scottish Highlands and the Isle of Skye. In: Gray, J.M. and Lowe, J.J. (eds.), *Studies in the Scottish Lateglacial environment*. Oxford: Pergamon Press, 143-162.

Walker, M.J.C., 1975a. Late Glacial and Early Postglacial environmental history of the Central Grampian Highlands, Scotland. *Journal of Biogeography* **2**: 265-284.

Walker, M.J.C., 1975b. A pollen diagram from the Pass of Drumochter, central Grampian Highlands, Scotland. *Transactions of the Botanical Society of Edinburgh* **42**: 335-343.

Walker, M.J.C., Ballantyne, C.K., Lowe, J.J., Sutherland, D.G., 1988. A reinterpretation of the Lateglacial environmental history of the Isle of Skye, Inner Hebrides, Scotland. *Journal of Quaternary Science* **3**: 135-146.

Wallinga, J., 2002. Optically stimulated luminescence dating of fluvial deposits: a review. *Boreas* **31**: 303-322.

Walther, M., 1984. *Geomorphologische Untersuchungen zum Spätglazial und Frühholozän in den Cuillin Hills (Insel Skye, Schottland)*. Unpublished PhD Thesis, Freie Universität Berlin.

Walton, E.K., Oliver, G.J.H., 1991. Lower Palaeozoic - stratigraphy. In: Craig, G.Y. (ed.), *Geology of Scotland*. 3rd ed., London: The Geological Society, 161-228.

Ward, R.C., Robinson, M., 1990. *Principles of Hydrology*. 3rd ed., London: McGraw Hill.

Watson, R.A., 1980. Landform development on moraines of the Klutlan Glacier, Yukon Territory, Canada. *Quaternary Research* **14**: 293-300.

Weertman, J., 1961. Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. *Journal of Glaciology* **3**: 965-978.

White, I.D., Mottershead, D.N., 1972. Past and present vegetation in relation to soilfluction on Ben Arkle, Sutherland. *Transactions of the Botanical Society of Edinburgh* **41**: 475-489.

Whittington, G.W., Hall, A.M., 2002. The Tolsta Interstadial, Scotland: correlation with D-O cycles GI-8 to GI-5? *Quaternary Science Reviews* **21**: 901-915.

Wilson, P., 2002. Morphology and significance of some Loch Lomond Stadial moraines in the south-central Lake District, England. *Proceedings of the Geologist's Association* **113**: 9-21.

Wilson, S.B., Evans, D.J.A., 2000. Scottish Landform Example No. 24: Coire a' Cheud-chnoic, the 'hummocky moraine' of Glen Torridon. *Scottish Geographical Journal* **116**: 149-158.

Winkler, S., 1996a. Frührezente und rezente Gletscherstandssschwankungen in Ostalpen und West-/Zentralnorwegen. Ein regionaler Vergleich von Chronologie, Ursachen und glazialmorphologischen Auswirkungen. *Trierer Geographische Studien* **15**: 1-579.

Winkler, S., 1996b. Front variations from outlet glaciers of Jostedalsbreen, western Norway, during the 20th century. *Norges Geologiske Undersøkelse Bulletin* **431**: 33-47.

Winkler, S., Haakensen, N., 1999. Kritische Überprufung der Möglichkeit zur Prognose des Gletscherverhaltens auf der Grundlage von Modellierungen - dargestellt anhand von regionalen Beispielen aus Norwegen. *Petermanns Geographische Mitteilungen* **143**: 291-304.

Winkler, S., Nesje, A., 1999. Moraine formation at an advancing temperate glacier: Brigsdalsbreen, western Norway. *Geografiska Annaler* **81A**: 17-30.

Winkler, S., Haakensen, N., Nesje, A., Rye, N., 1997. Glaziale Dynamik in Westnorwegen - Ablauf und Ursachen des aktuellen Gletschervorstoßes am Jostedalsbreen. *Petermanns Geographische Mitteilungen* **141**: 43-63.

Wohlfarth, B., Skog, G., Possnert, G., Holmquist, B., 1998. Pitfalls in the AMS radiocarbon-dating of terrestrial macrofossils. *Journal of Quaternary Science* **13**: 137-145.

Woodward, J., Murray, T., McCaig, A., 2002. Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard. *Journal of Quaternary Science* **17**: 201-209.

Woodward, J., Murray, T., McCaig, A., 2003. Reply: Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard. *Journal of Quaternary Science* **18**: 99-100.

Young, J.A.T., 1974. Ice wastage in Glenmore, upper Spey valley, Inverness-shire. *Scottish Journal of Geology* **10**: 147-157.

Young, J.A.T., 1978. The landforms of upper Strathspey. *Scottish Geographical Magazine* **94**: 76-94.

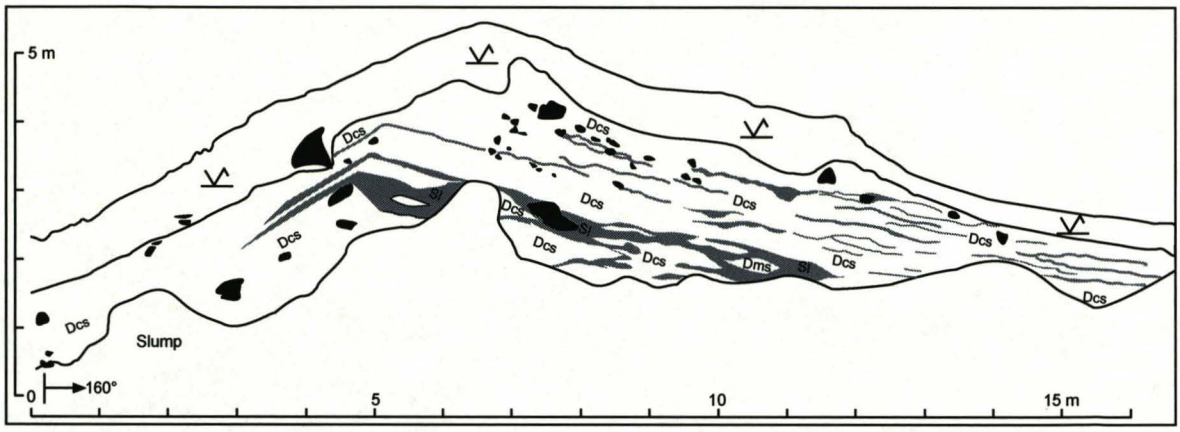
Ziaja, W., 2001. Glacial recession in Sørkappland and Central Nordenskiöldland, Spitsbergen, Svalbard, during the 20th century. *Arctic, Antarctic and Alpine Research* **33**: 36-41.

Zielinski, T., van Loon, A.J., 2000. Subaerial terminoglacial fans III: overview of sedimentary characteristics and depositional model. *Geologie en Mijnbouw* **79**: 93-107.

APPENDIX 1 DATABASE OF EXPOSURES IN "HUMMOCKY MORAINES" IN THE NW SCOTTISH HIGHLANDS

Introduction

This appendix describes the characteristics of ice-marginal "hummocky" moraines in the NW Highlands of Scotland and interprets them in a wider sedimentological context that aims to (a) elucidate the modes of moraine formation, (b) give an overview of the frequency of individual processes and (c) address the variability of their internal architecture. To make the results of this study transparent and testable, the exact location of each exposure, its approximate size and – where possible and useful – a log or an annotated photograph is presented in a database format. In order to give a quick overview of the genetic processes and their variability, descriptions and interpretations are kept short. Detailed information on lithofacies associations and the reasoning behind the interpretations is given in Chapters 4.3 and 4.4. The moraines are classified according to the criteria outlined in Table 4.1 in order to give an overview of the frequency of genetic processes recognised across the study area (see also Fig. 4.21 for a frequency distribution and Chapter 4.4.2 and 4.4.3 for implications derived from these data).



Section ID	Grid reference	Type and size of exposure
SLK 1	NC 39333 34445	Artificial, ca. 16 m x 5 m

Lithofacies description

See Chapter 4.3.1.1

Interpretation and classification (Chapter 4.3.1.7)

Undeformed terrestrial ice-contact fan formed at a stationary ice margin

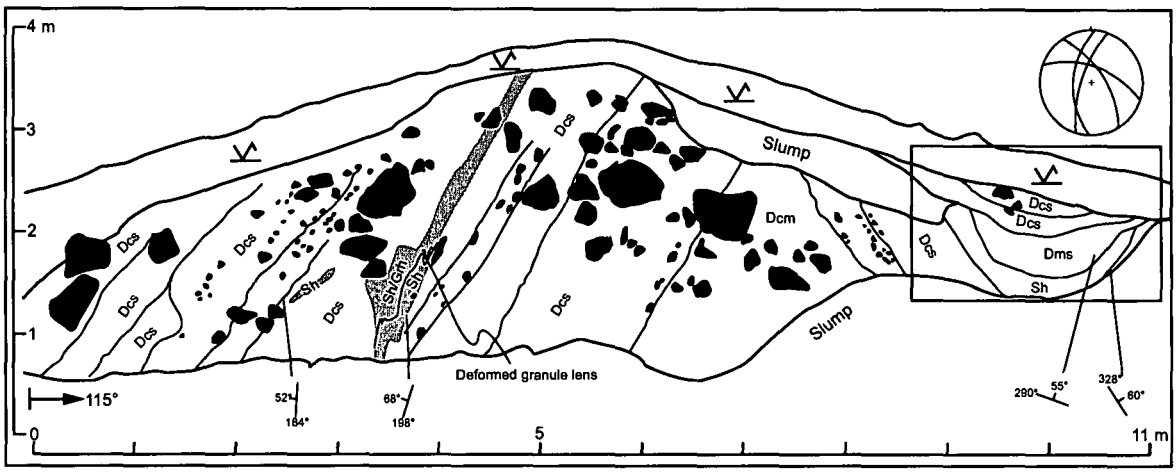
Section ID	Grid reference	Type and size of exposure
SLK 2	NC 40616 35646	Artificial, ca. 8 m x 3 m,

Lithofacies description

Alternating gently ESE-dipping units of Dcs, Dms and very fine to medium sand (Sm, Sl, Sh). Diamicts are easy to excavate, maximum a-axis length of clasts 0.7 m, cf. clast shape SLK2-1. Sand lenses show larger deformation structures (folds) and water escape structures.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan



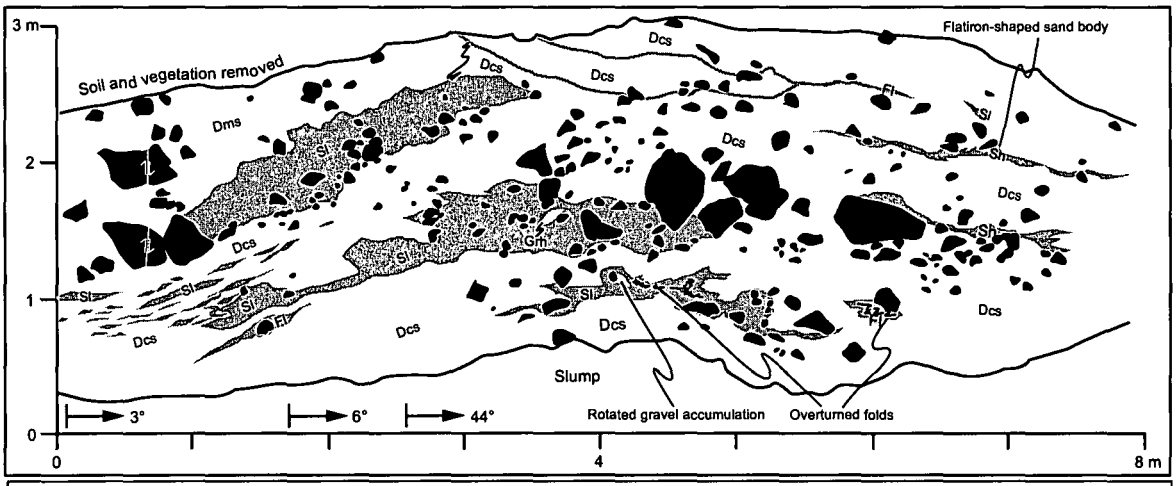
Section ID	Grid reference	Type and size of exposure
SLK 3	NC 43850 38270	Artificial, ca. 11 m x 4 m

Lithofacies description

See Chapter 4.3.1.3

Interpretation and classification (Chapter 4.3.1.7)

Dislocated terrestrial ice-contact fan, oversteepened during readvance



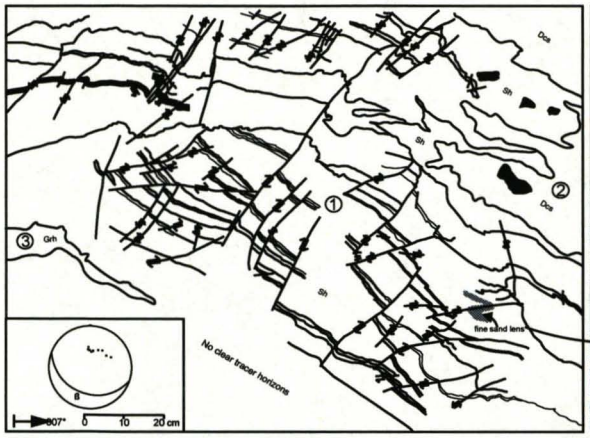
Section ID	Grid reference	Type and size of exposure
SLK 4	NC 44220 38710	Artificial, ca. 8 m x 3 m

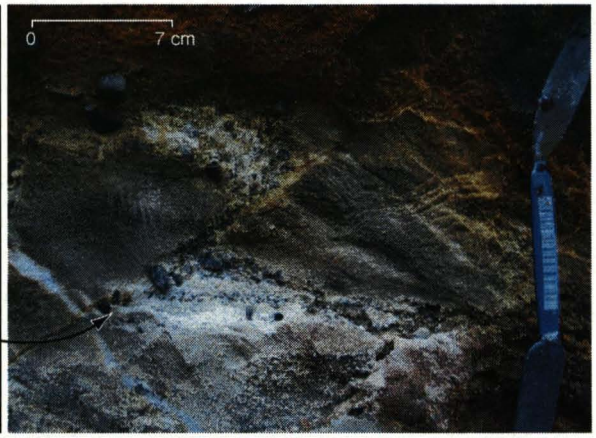
Lithofacies description

See Chapter 4.3.1.4

Interpretation and classification (Chapter 4.3.1.7)

Deformed and partly overridden terrestrial ice-contact fan





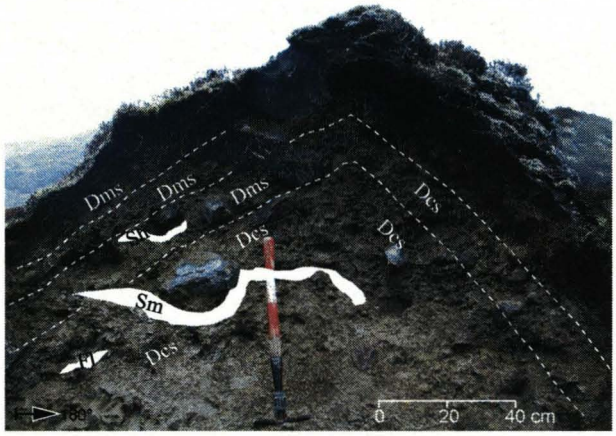
Section ID	Grid reference	Type and size of exposure
SLK 5	NC 44317 39171	Artificial, ca. 11 m x 4 m

Lithofacies description

See Chapter 4.3.1.6

Interpretation and classification (Chapter 4.3.1.7)

Deformed outwash fan



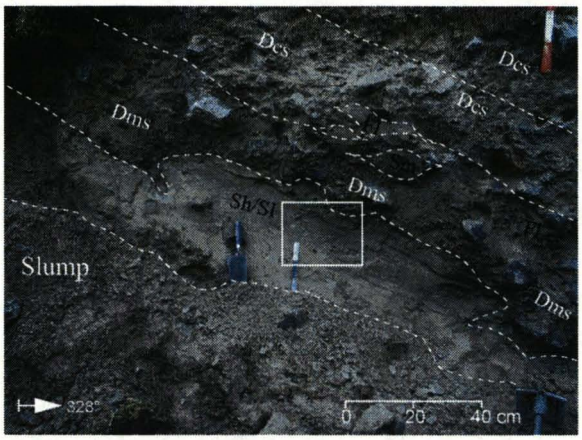
Section ID	Grid reference	Type and size of exposure
SLK 6	NC 36154 42032	Natural, ca. 3 m x 1.5 m

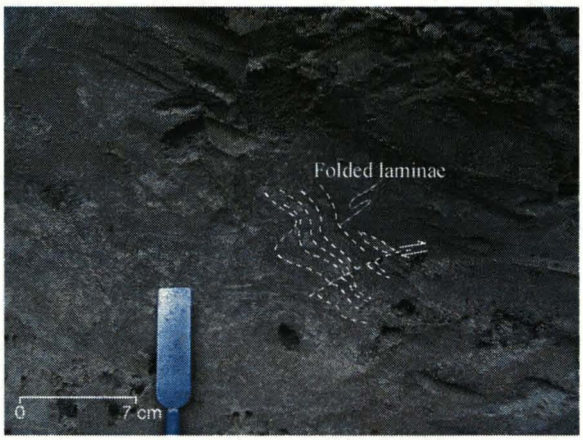
Lithofacies description

Alternating units of Dcs and Dms, easy to excavate, interbedded with Sm and thin Fl units. The latter show folding. Maximum a-axis length of clasts 0.4 m. Moraine surface and stratification of diamict units is asymmetrical, although individual units are parallel to each other. The sand units partly disrupt this stratification. Section is perpendicular to crestline (98°).

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan





Section ID	Grid reference	Type and size of exposure
SLK 7	NC 35990 41950	Natural, ca. 11 m x 2 m

Lithofacies description

Gently, conformably dipping, alternating units of silty to very fine sandy Dcs and Dms, easy to excavate, interbedded with Sm, Sh and thin Fl units. The latter show folding and signs of localised liquefaction. Prominent 0.4-m-thick unit of silt and very fine sand which is extensively folded and shows frequent reverse faults. Maximum a-axis length of clasts 1.0 m. Moraine surface displays classical fan shape, section is nearly parallel to crestline (~340°). The sand units partly disrupt this stratification.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan (apex location)

Section ID	Grid reference	Type and size of exposure
SLK 8	NC 35954 42014	Artificial, ca. 3 m x 1 m

Lithofacies description

Exposure in distal part of moraine, olique to crestline (235°). Alternating units of Dcs, easy to excavate, interbedded with Grm, GrO, Sm and thin Fl units. The latter two lithofacies show folding. Maximum a-axis length of clasts 0.4 m. The units conformably and gently dip downvalley.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

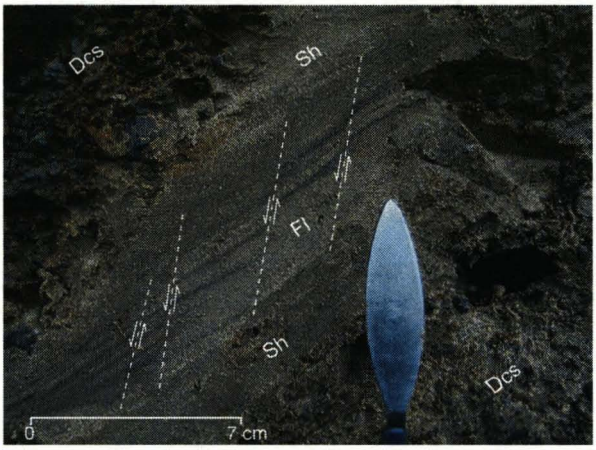
Section ID	Grid reference	Type and size of exposure
SLK 9	NC 40169 41290	Natural, ca. 1 m x 0.5 m, no photograph available

Lithofacies description

Small exposure in distal side of moraine with broad top. Gently, conformably dipping, alternating units of silty to very fine sandy Dcs, easy to excavate, interbedded with Sh and Fl units. Extensive signs of liquefaction and hydrofracturing where diamict breaks through sorted-sediment units. The latter display small-scale folding. Maximum a-axis length of clasts is 1.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan



Section ID	Grid reference	Type and size of exposure
SLK 10	NC 39792 41268	Natural, ca. 12 m x 2 m

Lithofacies description

Exposure in distal part of moraine, olique to crestline. Alternating, steeply dipping units of Dcs and Dms, easy to excavate, interbedded with Sh units up to 0.6 m thick. Close-up above shows en echelon normal faults in one of these sand bodies. Localised folding of unit boundaries, liquefaction and hydrofracturing evident throughout smaller parts of exposure cleaned. Maximum a-axis length of clasts 0.8 m.

Interpretation and classification (Chapter 4.3.1.7)

Dislocated terrestrial ice-contact fan

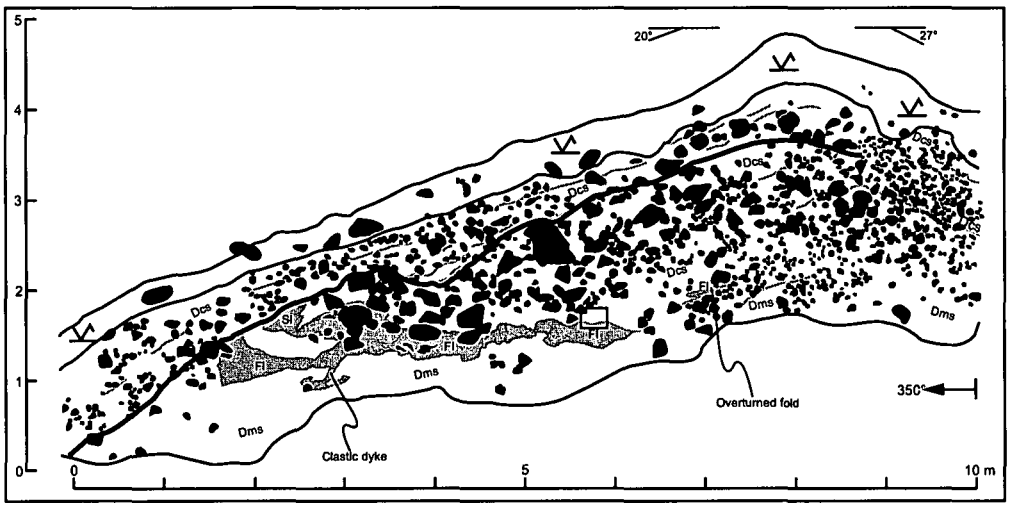
Section ID	Grid reference	Type and size of exposure
SLK 11	NC 38489 34789	Artificial, ca. 0.8 m x 0.4 m

Lithofacies description

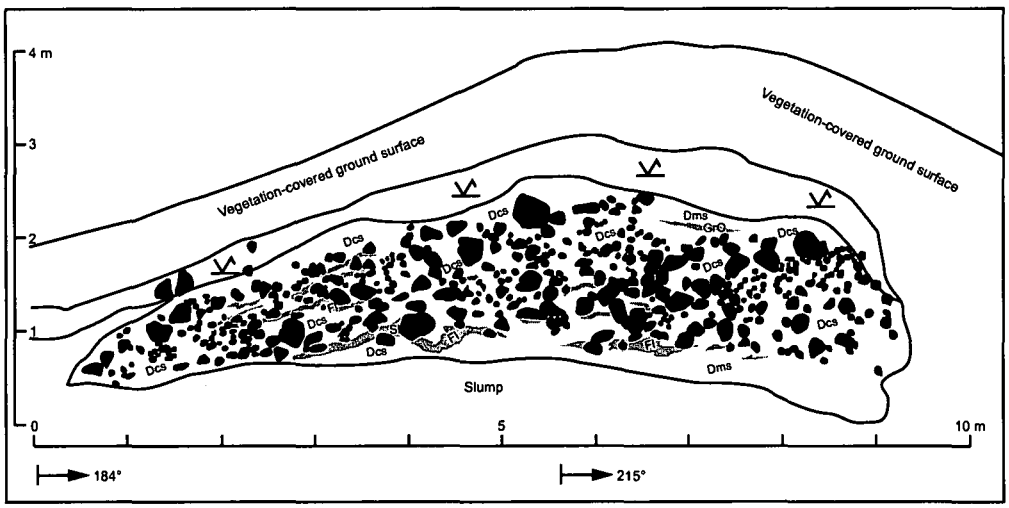
Exposure in distal part of asymmetrical moraine, perpendicular to crestline. Gently and conformably dipping units of Dcs, easy to excavate. No sorted sediment layers found, thus no statement about presence or absence of deformation possible. Maximum a-axis length of clasts 0.8 m.

Interpretation and classification (Chapter 4.3.1.7)

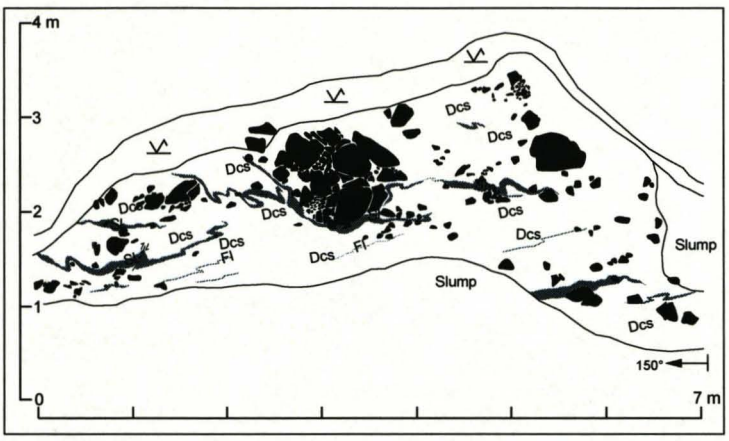
Terrestrial ice-contact fan, deformation history uncertain



Section ID	Grid reference	Type and size of exposure
SLK 12	NC 35820 34650	Artificial, ca. 10 m x 5 m
Lithofacies description		
See Chapter 4.3.1.2		
Interpretation and classification (Chapter 4.3.1.7)		
Heavily deformed terrestrial ice-contact fan		



Section ID	Grid reference	Type and size of exposure
SLK 13	NC 35750 34770	Artificial, ca. 10 m x 4 m
Lithofacies description		
See Chapter 4.3.1.2		
Interpretation and classification (Chapter 4.3.1.7)		
Deformed terrestrial ice-contact fan		



Section ID	Grid reference	Type and size of exposure
SLK 14	NC 38383 41438	Natural, ca. 7 m x 4 m

Lithofacies description

Large exposure through whole width of moraine with broad top. Gently, conformably dipping, alternating units of silty to very fine sandy Dcs, easy to excavate, interbedded with heavily folded and faulted Sh and Fl units. Extensive signs of liquefaction and hydrofracturing where diamict breaks through sorted-sediment units. Maximum a-axis length of clasts is 1.5 m. A fractured bedrock raft is located in the centre of the exposure.

Interpretation and classification (Chapter 4.3.1.7)

Heavily deformed terrestrial ice-contact fan

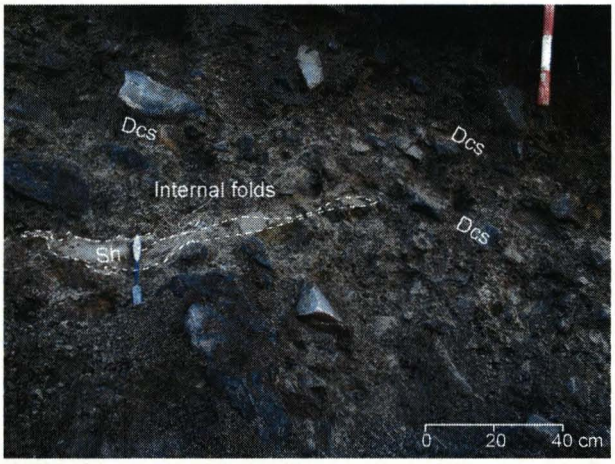
Section ID	Grid reference	Type and size of exposure
SLK 15	NC 39498 34825	Artificial, ca. 0.6 m x 0.5 m

Lithofacies description

Small exposure in distal side of asymmetrical moraine perpendicular to crestline. Gently, conformably dipping units of silty to very fine sandy Dcs, easy to excavate. No sorted sediment units found during excavation, thus no statement about presence or absence of deformation possible. Maximum a-axis length of clasts is 1.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain



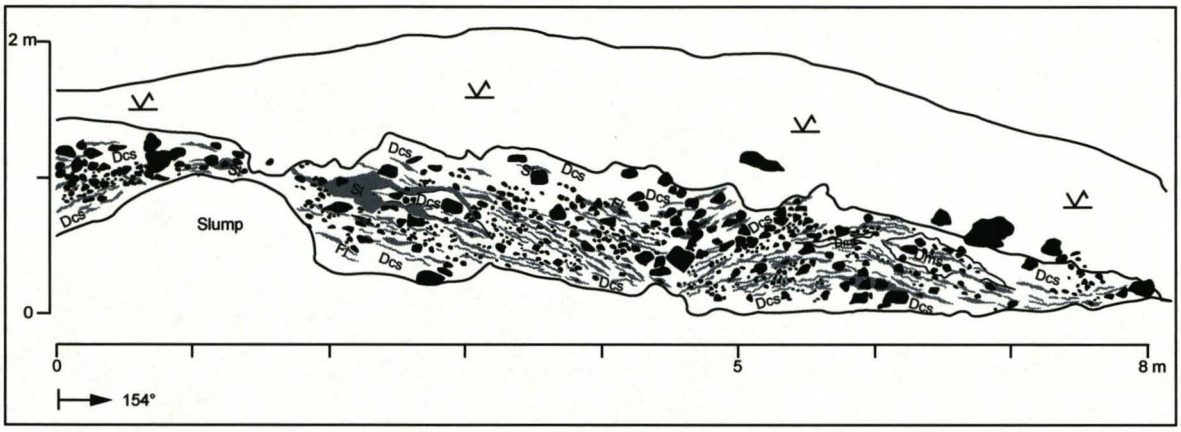
Section ID	Grid reference	Type and size of exposure
SLK 16	NC 40072 35435	Artificial, ca. 12 m x 3 m

Lithofacies description

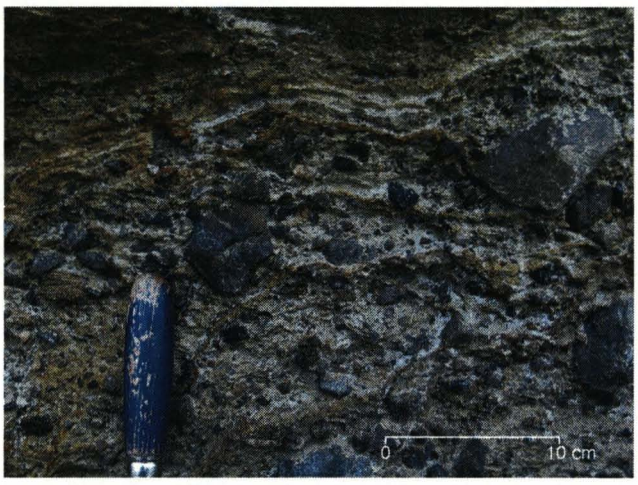
Detail of larger exposure in apex of fan-shaped moraine; exposure perpendicular to crestline. Gently, conformably dipping units of silty to very fine sandy Dcs, easy to excavate. Thin (< 1 cm thick) fine sandy laminae occur throughout the diamict units; contact between Dcs and prominent well-sorted very fine sand layer (pictured) is folded. Sand layer is also internally folded. Maximum a-axis length of clasts is 0.5 m.

Interpretation and classification (Chapter 4.3.1.7)

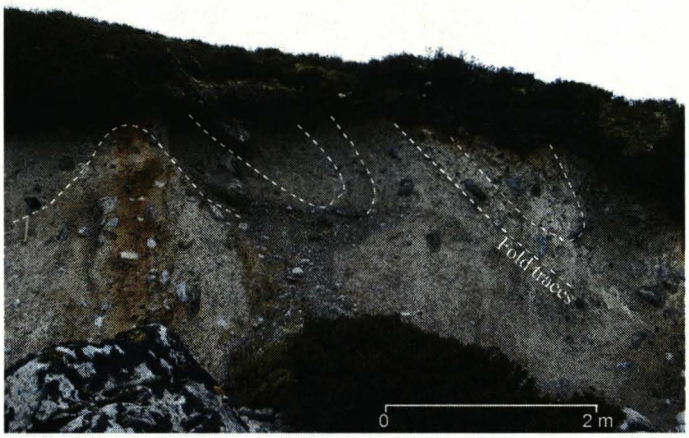
Deformed terrestrial ice-contact fan



Section ID	Grid reference	Type and size of exposure
SLK 17	NC 39180 24950	Natural, ca. 8 m x 2 m
Lithofacies description		
See Chapter 4.3.1.5		
Interpretation and classification (Chapter 4.3.1.7)		
Overridden terrestrial ice-contact fans		



Section ID	Grid reference	Type and size of exposure
SLK 18	NC 41808 22907	Natural, ca. 6 m x 2 m
Lithofacies description		
Shore exposure of alternating units of silty to very fine sandy Dcs and Dms, easy to excavate. Sorted sediment units (Fl, Sh, Sm, Sl) bifurcate around inclusions of diamicton (picture), which themselves form boudins. Maximum a-axis length of clasts is 0.6 m.		
Interpretation and classification (Chapter 4.3.1.7)		
Overridden terrestrial ice-contact fans		



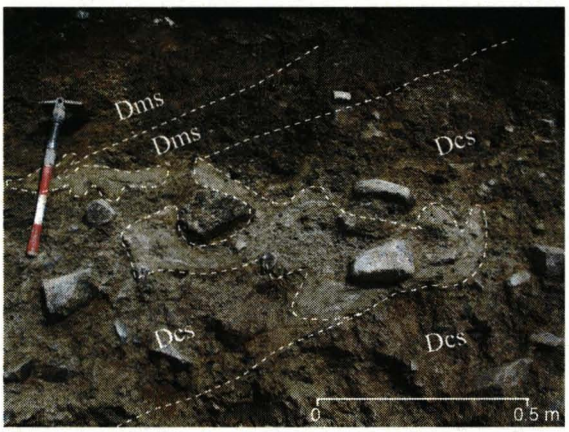
Section ID	Grid reference	Type and size of exposure
SLK 19	NC 40298 22855	Natural, ca. 50 m x 5 m

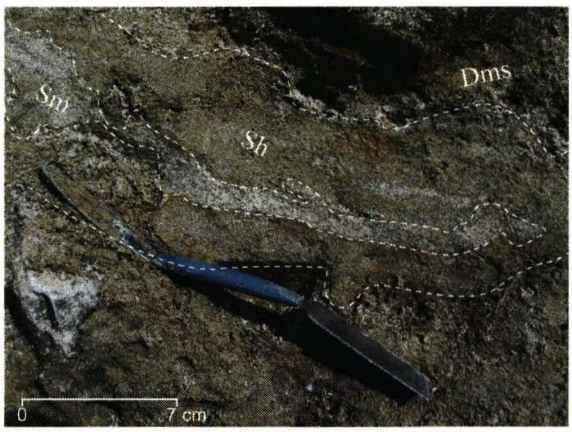
Lithofacies description

Exposure in moraines marking the Younger Dryas maximum. Alternating Dcs and Dms, partly very difficult, partly very easy to excavate. Interbeds of sorted sediments (Sl, Sh, Sm, Fl) that exhibit similar boudin structures to SLK 17 and 18. All the units are folded on a large scale, and, in places, fold limbs have been overturned by significant lateral compression.

Interpretation and classification (Chapter 4.3.1.7)

Overridden terrestrial ice-contact fans





Section ID	Grid reference	Type and size of exposure
SLK 20	NC 51055 34883	Artificial, ca. 7 m x 2 m

Lithofacies description

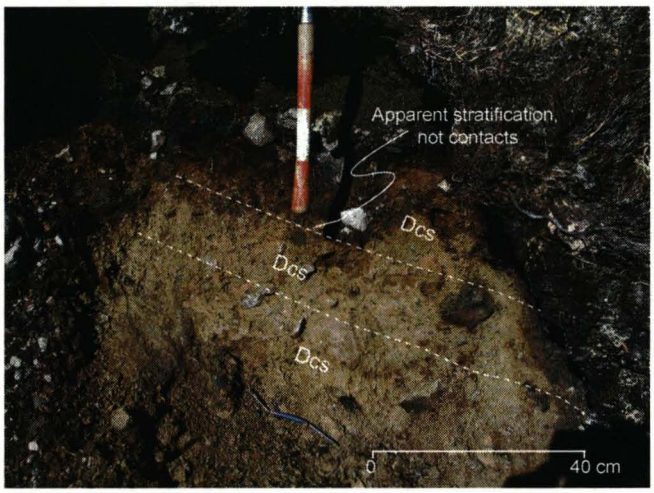
Exposure of alternating, gently and conformably dipping units of silty to very fine sandy Dcs and Dms, easy to excavate. Contacts with sorted sediment interbeds (Fl, Sh, Sm, Sl) are folded (picture). Stratification within some sorted units is also deformed (picture). Frequent signs of liquefaction where diamict units penetrate sorted-sediment units. Maximum a-axis length of clasts is 1.5 m.

Interpretation and classification (Chapter 4.3.1.7)

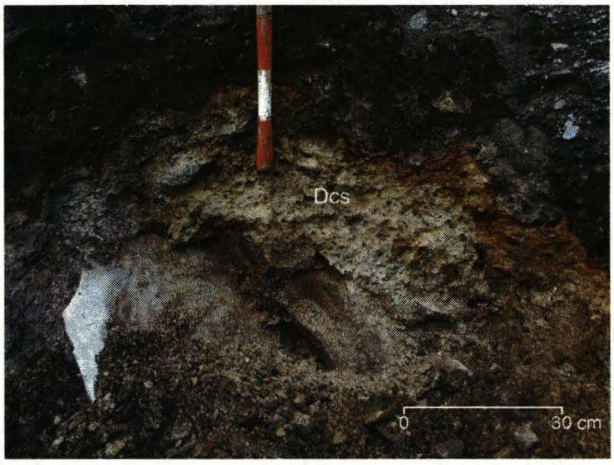
Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 21	NC 51502 35120	Artificial, ca. 3 m x 2 m
Lithofacies description		
Exposure in central part of an asymmetrical moraine. Alternating, conformably and gently dipping Dcs and Dms, easy to excavate. Interbeds of sorted sediments (Sl, Sh, Sm, Fl) exhibit folding and signs of liquefaction. Maximum a-axis length 0.8 m.		
Interpretation and classification (Chapter 4.3.1.7)		
Deformed terrestrial ice-contact fan		

Section ID	Grid reference	Type and size of exposure
SLK 22	NC 51622 35097	Artificial, ca. 5 m x 3 m
Lithofacies description		
Exposure in distal part of an asymmetrical moraine. Units of conformably and gently sloping Dcs, easy to excavate. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.0 m.		
Interpretation and classification (Chapter 4.3.1.7)		
Terrestrial ice-contact fan, deformation history uncertain		



Section ID	Grid reference	Type and size of exposure
SLK 23	NC 41117 35820	Artificial, ca. 7 m x 2 m
Lithofacies description		
Exposure in distal part of an asymmetrical moraine. Units of conformably and gently sloping Dcs, easy to excavate. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 0.7 m.		
Interpretation and classification (Chapter 4.3.1.7)		
Terrestrial ice-contact fan, deformation history uncertain		



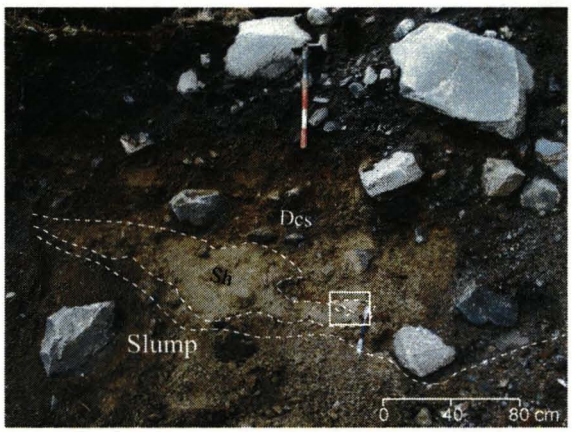
Section ID	Grid reference	Type and size of exposure
SLK 24	NC 41383 36064	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure in central part of symmetrical moraine. Units of conformable Dcs, which are easy to excavate, are subhorizontal. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.5 m. Note large clast in lower left of picture.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain



Section ID	Grid reference	Type and size of exposure
SLK 25	NC 41753 36522	Artificial, ca. 4 m x 2 m

Lithofacies description

Exposure in distal part of an asymmetrical moraine. Units of conformably and gently sloping Dcs, easy to excavate. Large lens of well-sorted Sh shows signs of folding and liquefaction (pictures). Maximum a-axis length 0.8 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

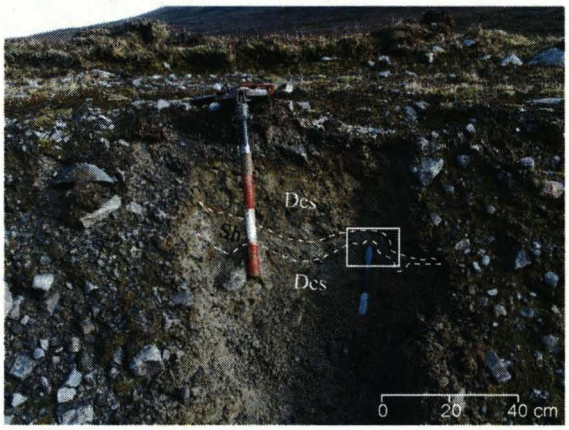
Section ID	Grid reference	Type and size of exposure
SLK 26	NC 42232 37014	Artificial, ca. 1.5 m x 1 m

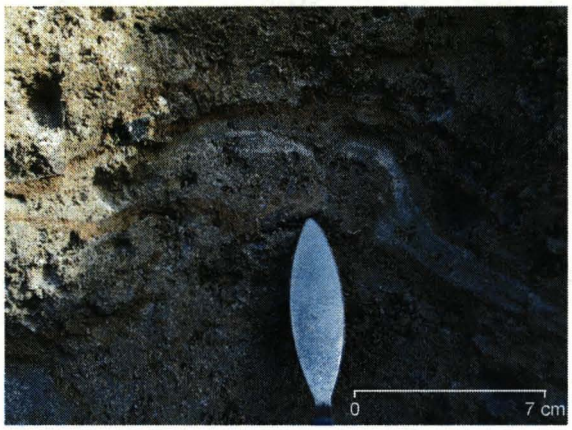
Lithofacies description

Exposure near crestline of asymmetrical moraine. Units of conformable Dcs, which are easy to excavate, are subhorizontal. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 3 m.

Interpretation and classification (cf. Chapter 3.1)

Terrestrial ice-contact fan, deformation history uncertain





Section ID	Grid reference	Type and size of exposure
SLK 27	NC 42293 37072	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure near crestline of moraine; morphology obscured by gravel extraction activity. Units of conformably and gently sloping silty to very fine sandy Dcs dominant; these are easy to excavate. Large lens of well-sorted Sh shows signs of folding (pictures). Outsized clasts of granule and medium gravel size occur throughout this unit. Maximum a-axis length 1.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

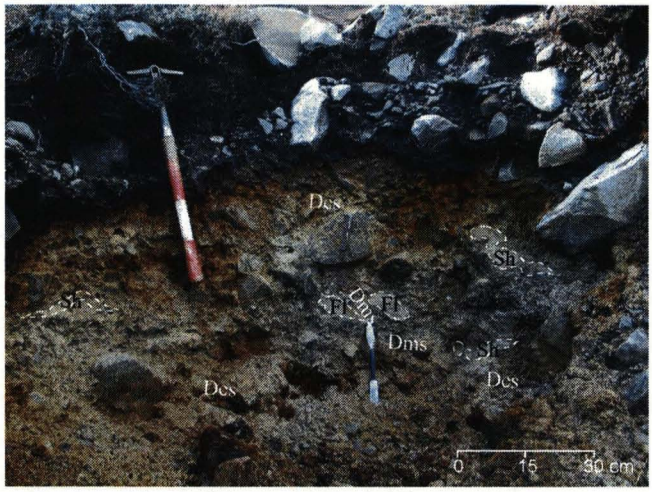
Section ID	Grid reference	Type and size of exposure
SLK 28	NC 42769 37376	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure near crestline of moraine; morphology obscured by gravel extraction activity. Units of conformable Dcs and Dms, which are easy to excavate, are subhorizontal. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.5 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain



Section ID	Grid reference	Type and size of exposure
SLK 29	NC 43595 37961	Artificial, ca. 3 m x 1 m

Lithofacies description

Exposure near crestline of asymmetrical moraine. Units of conformably and gently sloping silty to very fine sandy Dcs and Dms, which are easy to excavate, occur. Smaller lenses of well-sorted Sh and Fl show folding (picture). In one location, an Fl lens is cut by Dms (hydrofracturing). Maximum a-axis length 0.7 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 30	NC 44268 39017	Artificial, ca. 3 m x 1.5 m

Lithofacies description

Exposure near crestline of moraine; morphology obscured by gravel extraction activity. Subhorizontal units of conformable Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.1 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 31	NC 44187 38581	Artificial, ca. 1 m x 0.5 m

Lithofacies description

Exposure near crestline of moraine; morphology obscured by gravel extraction activity. Subhorizontal units of conformable very fine sandy Dcs, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not found, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 0.8 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 32	NC 36215 47621	Natural, ca. 10 m x 5 m

Lithofacies description

Exposure through whole width of an asymmetrical moraine. Units of conformably and steeply (~25°) sloping sorted sediments ranging from laminated silts and very fine sand to openwork gravel occur throughout; diamicts are absent. The finer-grained units show folding and normal and reverse faulting (picture).

Interpretation and classification (Chapter 4.3.1.7)

Deformed outwash sediments/fan

Section ID	Grid reference	Type and size of exposure
SLK 33	NC 36198 47790	Natural, ca. 7 m x 3 m

Lithofacies description

Exposure in distal half of asymmetrical moraine. Subhorizontal units of conformable Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments (Sh, Sm, Fl) show some small-scale folds and reverse faults. Maximum a-axis length of boulders in diamict units is 2.5 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 34	NC 39027 41704	Artificial, ca. 3 m x 1 m

Lithofacies description

Exposure in distal part of an asymmetric moraine. Subhorizontal units of conformable very fine sandy Dcs, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not observed, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 2.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 35	NC 38900 41415	Artificial, ca. 3 m x 1 m

Lithofacies description

Exposure in central part of an asymmetric moraine. Subhorizontal units of conformable very fine sandy Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not observed, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.3 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 36	NC 39013 417800	Artificial, ca. 2.5 m x 1.5 m

Lithofacies description

Exposure through distal part of an asymmetrical moraine. Alternating units of Dcs and Dms dip gently and conformably to the left and are easy to excavate. Interbeds of sorted sediments (Sh, Sm, Fl) show larger folds (picture) and some reverse faults. Maximum a-axis length of boulders in diamict units is 0.9 m.

Interpretation and classification (Chapter 4.3.1.7)

Heavily deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 37	NC 36700 35029	Natural, ca. 1 m x 0.5 m

Lithofacies description

Exposure in distal half of asymmetrical moraine. Subhorizontal units of conformable Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not observed, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length of boulders in diamict units is 0.6 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 38	NC 38312 35850	Artificial, ca. 3 m x 2 m

Lithofacies description

Exposure in central part of an asymmetric moraine. Subhorizontal units of conformable very fine sandy Dcs, which are easy to excavate, occur throughout. Contacts of sorted sediment interbeds (Sh, Fl) with surrounding diamicts are folded. Maximum a-axis length 1.2 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 39	NC 38613 41500	Artificial, ca. 3 m x 1 m

Lithofacies description

Exposure in proximal part of an asymmetric moraine. Alternating, gently dipping units of conformable very fine sandy Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not observed, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.3 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 40	NC 35615 34823	Natural, ca. 5 m x 3 m

Lithofacies description

Exposure in distal part of an asymmetric moraine. Alternating, gently dipping units of conformable very fine sandy Dcs and Dms, which are easy to excavate, occur throughout. Contacts of sorted sediment interbeds (Sh, Sm) with surrounding diamicts are folded. Maximum a-axis length 2.2 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 41	NC 39105 32513	Artificial, ca. 7 m x 1 m

Lithofacies description

Exposure near surface of flat-topped moraine. Alternating units of Dcs and Dms dip gently and conformably to the right and are easy to excavate. Interbeds of sorted sediments (Sh, Sm, Fl) show folds and reverse faults. Maximum a-axis length of boulders in diamict units is 3.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 42	NC 30905 41465	Natural, ca. 2 m x 1 m

Lithofacies description

Exposure in distal half of asymmetrical moraine. Subhorizontal units of conformable Dcs and Dms, which are easy to excavate, occur throughout. Contacts of sorted sediment interbeds (Sh, Sm) with surrounding diamicts are folded. Maximum a-axis length of boulders in diamict units is 1.7 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 43	NC 30832 41212	Natural, ca. 3 m x 2 m

Lithofacies description

Exposure in distal part of an asymmetric moraine. Gently, conformably sloping units of very fine sandy Dcs and Dms, which are easy to excavate, occur throughout. A prominent bed of undisturbed Fl occurs in the lower right part of the exposure indicating absence of postdepositional deformation. Maximum a-axis length 1.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan

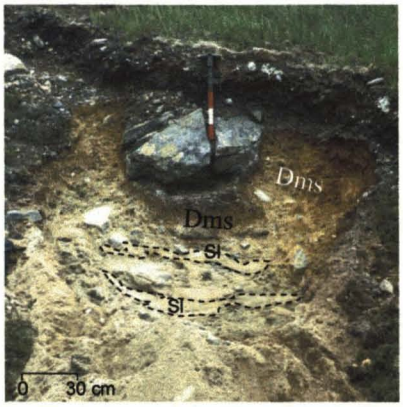
Section ID	Grid reference	Type and size of exposure
SLK 44	NC 30727 40972	Natural, ca. 4 m x 3 m

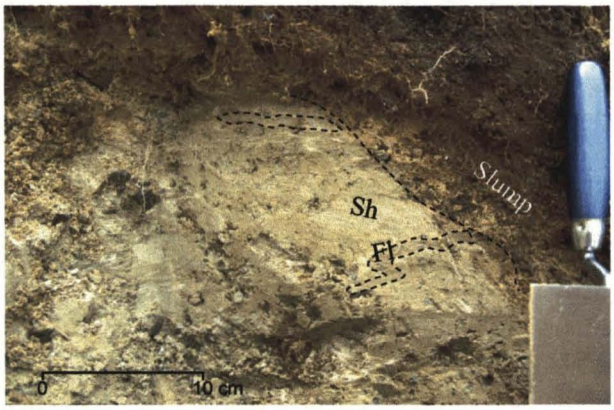
Lithofacies description

Exposure in proximal part of an asymmetric moraine. Alternating, gently dipping units of conformable very fine sandy Dcs and Dms, which are easy to excavate, occur throughout. Interbeds of sorted sediments were not observed, hence no statement about presence or absence of deformation structures possible. Maximum a-axis length 1.5 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain





Section ID	Grid reference	Type and size of exposure
SLK 45	NC 38427 28125	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure near crestline of an asymmetric moraine. Alternating, gently dipping units of conformable very fine sandy Dms, which are easy to excavate, observed. Contacts of sorted sediment interbeds (Sh, Sm, Fl) with surrounding diamicts are folded (picture); sorted sediment units also show internal deformation (picture). Maximum a-axis length 1.2 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

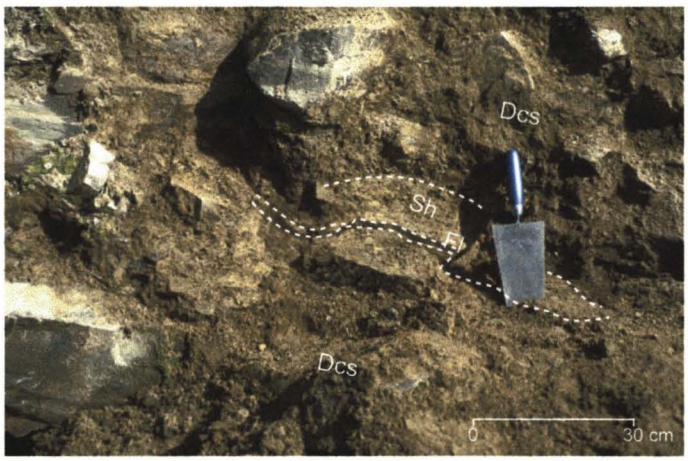
Section ID	Grid reference	Type and size of exposure
SLK 46	NC 33894 31214	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure near crestline of asymmetrical moraine. Alternating units of Dcs and Dms are subhorizontal and easy to excavate. Interbeds of sorted sediments were not found, hence no statements about presence or absence of deformation structures possible. Maximum a-axis length of boulders in diamict units is 1.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain



Section ID	Grid reference	Type and size of exposure
SLK 47	NC 34087 31152	Artificial, ca. 2 m x 1 m

Lithofacies description

Exposure in distal half of asymmetrical moraine. Subhorizontal units of conformable Dcs and Dms, which are easy to excavate, occur throughout. Contacts of sorted sediment interbeds (Sh, Sm) with surrounding diamicts are folded. Maximum a-axis length of boulders in diamict units is 2.5 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 48	NC 40237 41358	Natural, ca. 12 m x 1.5 m

Lithofacies description

Exposure in basal part through whole width of an asymmetric moraine. Alternating, gently dipping units of conformable very fine sandy Dcs and Dms, which are easy to excavate, observed. Contacts of sorted sediment interbeds (Sh, Sm, Fl) with surrounding diamicts are folded as are the units internally. Maximum a-axis length of clasts contained in diamicts is 2.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 49	NC 45530 26072	Artificial, ca. 6 m x 2 m

Lithofacies description

Exposure near crestline of an asymmetric moraine; upper metre removed for gravel extraction. Alternating, gently dipping units of conformable very fine sandy Dcs and Dms, which are easy to excavate, dominate section. Contacts of sorted sediment interbeds (Sm, Fl) with surrounding diamicts are folded. Maximum a-axis length of clasts contained in diamicts is 2.5 m.

Interpretation and classification (Chapter 4.3.1.7)

Deformed terrestrial ice-contact fan

Section ID	Grid reference	Type and size of exposure
SLK 50	NC 40821 44673	Artificial, ca. 5 m x 1.5 m

Lithofacies description

Exposure near basal part and through whole width of asymmetrical moraine. Alternating units of Dcs and Dms are subhorizontal and easy to excavate. Interbeds of sorted sediments were not found, hence no statements about presence or absence of deformation structures possible. Maximum a-axis length of boulders in diamict units is 1.3 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 51	NC 40513 41340	Artificial, ca. 2 m x 1.5 m

Lithofacies description

Exposure near basal part of asymmetrical moraine. Alternating units of Dcs and Dms are subhorizontal and easy to excavate. Interbeds of sorted sediments were not found, hence no statements about presence or absence of deformation structures possible. Maximum a-axis length of boulders in diamict units is 0.6 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan, deformation history uncertain

Section ID	Grid reference	Type and size of exposure
SLK 52	NC 3425 5273	Artificial, ca. 7 m x 3 m

Lithofacies description

Exposure through whole width of symmetrical moraine. Alternating units of Dcs and Dms are subhorizontal and easy to excavate. Interbeds of sorted sediments are restricted to thin veneers of silts and occasional Sl layers. no evidence of deformation structures. Maximum a-axis length of boulders in diamict units is 3.0 m.

Interpretation and classification (Chapter 4.3.1.7)

Terrestrial ice-contact fan

APPENDIX 4 EXAMPLES OF MAPS PRODUCED FROM AERIAL PHOTOGRAPHS

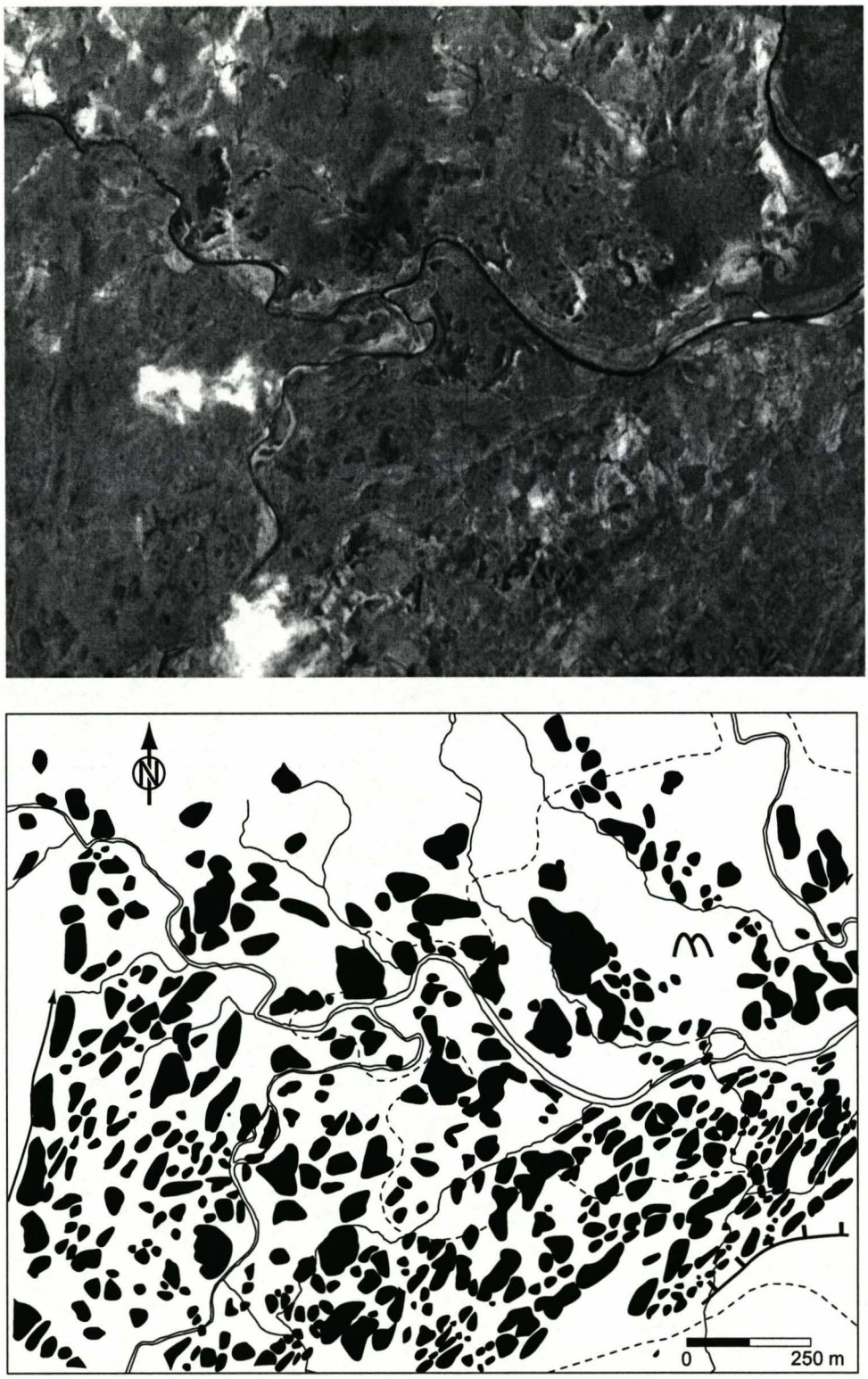


Fig. A4.1 Example of geomorphological map produced from aerial photograph 1989080 (RCAHMS Edinburgh, Crown copyright) using the approach described in Chapter 2. For key see Appendix 2.

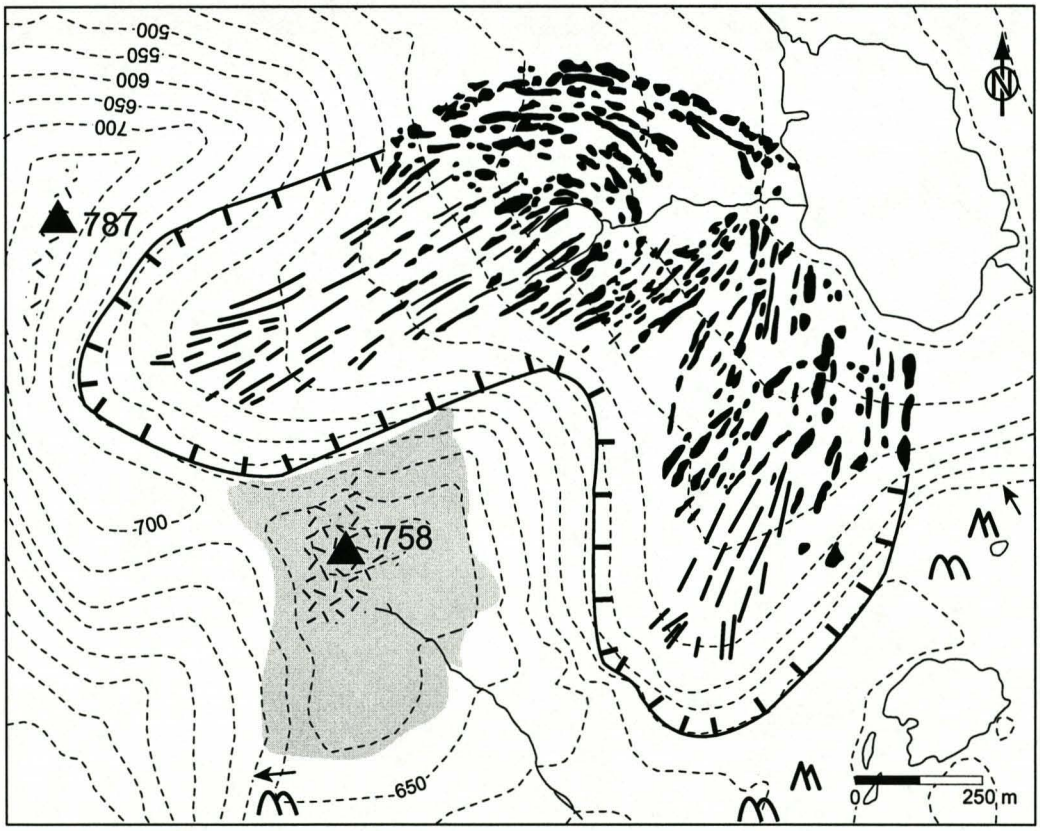


Fig. A4.2 Example of geomorphological map produced from aerial photograph 3788088 (RCAHMS Edinburgh, Crown copyright). For key see Appendix 2.

Fig. A4.3 Overview aerial photograph of the Longyeardalen catchment (cf. Fig. 5.1). The extent of the debris covers of the three glaciers studied can be clearly seen.

APPENDIX 5 PEER-REVIEWED PUBLICATIONS

Paper 1

Lukas, S. 2005. A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland. *Boreas* **34**: 287-307.

Paper 2

Lukas, S., Nicholson, L.I., Ross, F.H. and Humlum, O. 2005a. Formation, meltout processes and landscape alteration of high-arctic ice-cored moraines – examples from Nordenskiöld Land, central Spitsbergen. *Polar Geography* **29(2)**: 79-109.

Paper 3

Lukas, S., Spencer, J.Q.G., Robinson, R.A.J. and Finch, A.A. 2005b. Problems of luminescence dating of Late Quaternary sediments in the NW Scottish Highlands. *Quaternary Science Reviews (Quaternary Geochronology)*, **submitted**.

A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland

SVEN LUKAS

BOREAS

Lukas, S. 2005 (August): A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland. *Boreas*, Vol. 34, pp. 287–307. Oslo. ISSN 0300-9483.

The melt-out of material contained within englacial thrust planes has been proposed to result in the formation of stacked moraine sequences with characteristic proximal rectilinear slopes. This model has been applied to explain the formation of Scottish Younger Dryas ice-marginal ('hummocky') moraines on the basis of these morphological characteristics. However, no sedimentological data exist to support this proposal. This article reviews hitherto proposed models of 'hummocky' moraine formation and presents detailed geomorphological and sedimentological results from the NW Scottish Highlands with the aims of reconstructing the dynamics of Younger Dryas glaciers and of testing the applicability of the englacial thrusting model. Exposures demonstrate that moraines represent terrestrial ice-contact fans throughout, with a variety of postdepositional deformation structures being identified in most cases, indicating that glacier retreat was incremental and oscillatory; proximal rectilinear slopes are interpreted as ice-contact faces formed after ice support was withdrawn during retreat. This evidence strongly suggests a temperate glacier regime and short glacier response times similar to those in present-day SW Norway or Iceland. It contradicts the thrusting model and the proposal that Svalbard might form a suitable analogue for Younger Dryas moraines in Scotland.

Sven Lukas (e-mail: SL33@st-andrews.ac.uk), School of Geography and Geosciences, University of St Andrews, Irvine Building, North Street, St Andrews, KY16 9AL, Scotland; *received 15th December 2004, accepted 8th February 2005.*

Ice-marginal moraines, defining the extent of glacier ice at a given time, form the basis for establishing detailed chronologies of ice retreat. The retreat history can range from relatively short time periods like the Little Ice Age (e.g. Bickerton & Matthews 1993; Winkler *et al.* 1997) to longer-lasting glacial stages (e.g. Penck & Brückner 1901/1909; Liedtke 1981; Benn & Owen 2002). Moraines also form the most important basis of glacier reconstruction in terms of establishing the extent and equilibrium line altitude (ELA) of former glaciers (e.g. Benn & Evans 1998; Nesje & Dahl 2000; Ballantyne 2002).

Observations of moraine formation at modern glacier margins, where the sedimentological end product can be clearly linked to the climatic and glaciological regime have led to an advanced understanding of the links between glacier dynamics and ice-marginal deposition (e.g. Sharp 1984; Humlum 1985; Boulton 1986; Eybergen 1987; Krüger 1994, 1996; Matthews *et al.* 1995; Winkler & Nesje 1999; Kjær & Krüger 2001). Applied to a palaeo-setting, their internal architecture, if sufficiently exposed, can yield important information on the nature of moraine formation and, conversely, enables a reconstruction of palaeoglaciological variables of former glaciers, e.g. their dynamics and thermal regime (cf. Benn & Evans 1998; Evans & Benn 2004; Lukas & Benn submitted).

Detailed studies have so far focused on large Pleistocene moraine belts and their structural geology to elucidate the processes of their formation (e.g. Aber *et al.* 1989; van der Wateren 1995), but small moraines formed by Pleistocene valley glaciers have been relatively neglected, partly because of their assumed 'poor preservation potential' (Bennett 2001: p. 231). In the Scottish context, the transfer of knowledge gained in modern environments is unfortunately limited to one detailed investigation of both moraine geomorphology and sedimentology of ice-marginal moraines on the Isle of Skye (Benn 1990, 1992; Benn *et al.* 1992). All other studies to date have largely been restricted by limited exposure conditions (Bennett & Boulton 1993; Hambrey *et al.* 1997; Bennett *et al.* 1998; Lukas 2003). This study is the first detailed contribution that integrates moraine morphology and sedimentology to reconstruct glacier dynamics in the Scottish context since the seminal work by Benn (1990, 1992). On a global scale, it demonstrates how sedimentological data from small moraines can effectively be used to decipher a signal of past glacier dynamics and hence – indirectly – a palaeoclimatic and palaeoglaciological signature contained within glacigenic sediments.

The moraines in Scotland formed during the Younger Dryas are potentially very valuable terrestrial archives, as they represent one of the few cases where glacier dynamics of relatively small former ice caps can be studied and their response be linked to palaeoclimate proxies (e.g. Brooks & Birks 2000; Ballantyne 2002).

Taylor & Francis
Taylor & Francis Group

DOI: 10.1080/03009480510013042

Scottish 'hummocky' moraines: a review of existing models and problems

Model I: Incremental, active retreat and modern analogues

During the 1960s and 1970s, 'hummocky moraine', so-termed because of its apparently chaotic appearance when viewed from the ground, was interpreted as the product of widespread *in situ* ice stagnation (e.g. Sissons 1967, 1974, 1979), a notion informed by interpretations from the margins of the Laurentide Ice Sheet (Hoppe 1952; Sissons 1967) that is apparently still valid today (Eyles *et al.* 1999). In Scotland, the distribution of this landform assemblage has been used extensively to define the maximum extent of Younger Dryas (Loch Lomond Stadial) glaciers (e.g. Sissons 1974; Ballantyne 1989).

Eyles (1983) was the first to conclude that Scottish 'hummocky moraine' might instead reflect the product of controlled and uncontrolled deposition of supraglacial debris from incrementally stagnating margins of active glaciers as observed in Iceland. Crucially, however, no sedimentological data were provided from Scotland and, thus, hummocky moraine was not largely accepted as a product of 'active' incremental retreat.

Using detailed sedimentological analyses and geomorphological mapping, Benn (1990, 1992) recognized that moraines are commonly aligned in individual chains, which he interpreted as ice-marginal moraine fragments formed during incremental 'active' (as opposed to stagnating or 'passive') ice retreat (Fig. 1A–C). At about the same time, Bennett & Glasser (1991) and Bennett & Boulton (1993) independently arrived at the same conclusion using geomorphological evidence from aerial photographs. To date, the following sedimentary processes leading to the formation of Scottish ice-marginal ('hummocky') moraines have been recognized: (a) sedimentation along stationary ice margins leading to the formation of terrestrial ice-contact fans (Benn 1990, 1992; Mitchell & Lukas 2004); (b) ice-marginal pushing of glacigenic material to form a push moraine (Benn 1990, 1992); (c) dumping of supraglacial and englacial material following *in situ* ice stagnation forming unorganized, 'chaotic' moraines (Eyles 1983; Benn 1990, 1992); (d) subglacial moulding, i.e. formation of flutes, megaflutes and drumlins (Hodgson 1986; Benn 1992; Bennett 1995; Wilson & Evans 2000). With the exception of (c) and (d), actively formed recessional moraines (the majority of Scottish hummocky moraine) have been used to reconstruct the retreat patterns of glaciers across Scotland in many studies since (Benn 1990, 1992; Bennett & Glasser 1991; Bennett & Boulton 1993; Lukas 2003, 2004).

Eyles (1983) argued that Icelandic temperate glaciers might be regarded as a modern analogue. In Iceland, hummocky moraine forms as a result of melt-out and transfer of supraglacial debris from stagnant ice bodies into topographic troughs resulting in topographic inversion and a 'chaotic' appearance (Eyles 1983; Kjær & Krüger 2001). Sedimentological studies in Scotland have not been able to confirm the importance of this process, and apart from a few isolated moraines there is

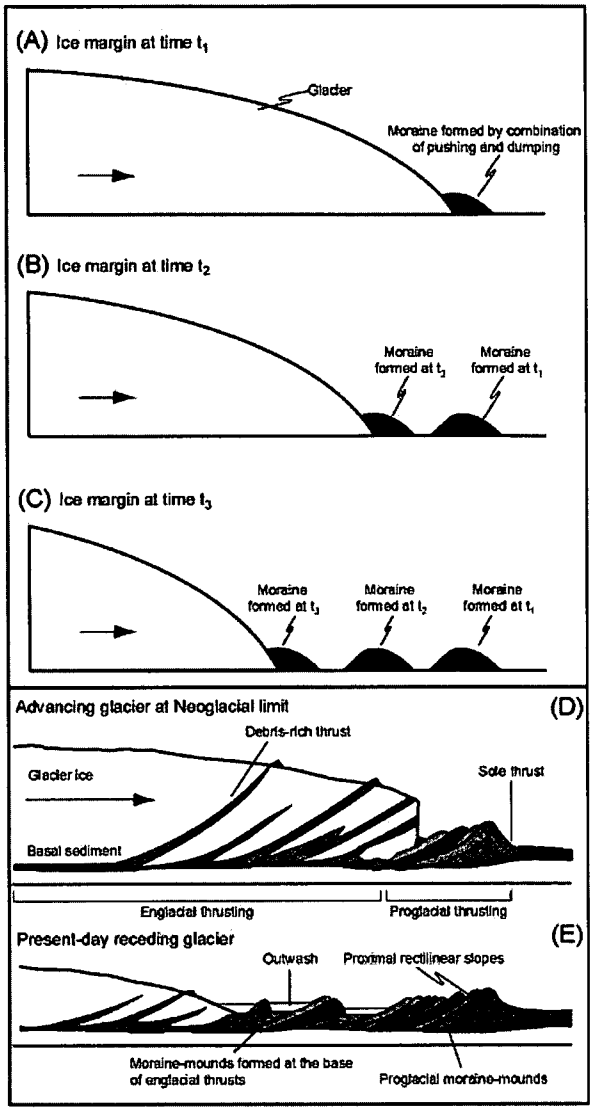


Fig. 1. A.–C. Conceptual sketch of moraine formation according to the incremental 'active' retreat model without incorporation of dead ice bodies or significant stagnation (for explanation, see text). D. Sketch of englacial and proglacial thrusting as deduced for polythermal glaciers on Svalbard. Stresses are propagated into the foreland along a basal décollement surface or sole thrust. E. Lowering of englacial debris after ice retreat to form a sequence of stacked moraines with characteristic proximal rectilinear slopes that dip upglacier. Note the inferred lack of alteration in the mounds due to dead ice melt-out ((D) and (E) redrawn from Bennett *et al.* 1998. Copyright John Wiley and Sons Limited. Reproduced with permission).

very little evidence for dead ice melt-out (Benn 1990, 1992). Apart from the brief note by Benn (1992: p. 784) that the erosional and depositional evidence on Skye indicates that the ice was 'wet-based throughout', little is known about the former thermal regime and dynamics of the glaciers.

Model II: Englacial and proglacial thrusting

The search for modern analogues was recently taken up again by Hambrey *et al.* (1997) and Bennett *et al.* (1998). Englacial and proglacial thrusting, described from polythermal surging and non-surging glaciers on Svalbard, is thought to be responsible for the formation of 'stacked' moraines with characteristic proximal rectilinear slopes. According to this model, englacial thrust planes form at the transition from warm-based to cold-based ice near the snout as a result of flow compression. Basal sediment is apparently elevated along these thrust planes into an englacial position where it forms englacial debris septa (Fig. 1D). During glacier retreat, this material is inferred to melt-out without alteration to leave a sequence of stacked moraines with characteristic proximal rectilinear slopes (Fig. 1E; Bennett *et al.* 1998). Stress propagation into the foreland induces stacking of proglacial material to a series of moraines (Fig. 1A, B). Sedimentological evidence in support of their model is restricted to shallow surface exposures, but it was stated that 'each mound is composed of a single facies or facies association' (Hambrey *et al.* 1997: p. 624; Bennett *et al.* 1998: p. 19).

This work highlights the morphological similarity of ice-marginal moraines at ice margins in Svalbard and one site in the Scottish Highlands (Hambrey *et al.* 1997, 2001; Bennett *et al.* 1998). This is notably the presence of proximal rectilinear slopes, interpreted as the surface representation of former thrust planes, and their uniform dip direction giving suites of moraines a 'stacked' appearance. Only sparse evidence from shallow exposures in moraine crests displaying 'uniform' facies is used. No detailed description or interpretation of the complete internal architecture of a moraine has ever been presented in support of this model (e.g. Bennett *et al.* 1998; Graham & Midgley 2000; Midgley 2001; Graham 2002). As a logical consequence of this morphological similarity, however, it has been proposed that Svalbard may form a good analogue for Upland Britain during the Younger Dryas (Hambrey *et al.* 1997, 2001; Bennett *et al.* 1998).

Implications and model applicability

The first model of active incremental ice retreat implies that one arc of moraine ridges and mounds reflects one former ice-marginal position formed during 'active' or oscillatory retreat (Fig. 1A–C). The second model implies that several moraines can form in, and consequently represent, one glacial event as they result from a combination of the melt-out of englacial material contained in thrusts and material thrust up in the foreland into a series of stacked moraines (Fig. 1D, E).

It has yet not been possible to thoroughly test geomorphologically and sedimentologically which of the two models applies to the Scottish palaeo-context. Although the model of incremental retreat is often preferred over the thrusting hypothesis on morphological grounds (e.g. the presence of bifurcating moraines), this has rarely been backed up by sufficient sedimentological evidence (e.g. Wilson & Evans 2000; McDougall 2001; Ballantyne 2002; Wilson 2002; Lukas 2003).

Recent re-evaluation of the structures interpreted as thrusts at Kongsvegen, Svalbard, has led to the conclusion that the role of thrusting may have been overestimated (Woodward *et al.* 2002, 2003). As a consequence, Woodward *et al.* (2002: p. 207) conclude that the model of englacial thrusting as an explanation of British 'hummocky moraine' formed during the Younger Dryas 'also must be questioned'. Similarly, the landforms in Glen Torridon interpreted as a result of thrusting by Hambrey *et al.* (1997) and Bennett *et al.* (1998) have been more convincingly re-interpreted as a palimpsest landscape in which older moraines have been overprinted by flutes during the Younger Dryas (Wilson & Evans 2000).

A thorough test of the applicability of the two models introduced above is thus necessary to advance our understanding of the nature of glaciation during the Younger Dryas in Scotland.

The purpose of this article is, first, to present detailed sedimentological and geomorphological evidence from moraines formed during the Younger Dryas in the far NW Highlands of Scotland to increase the limited data set of ice-marginal moraine sedimentology; second, to test which (if any) of the two models is applicable to that area, and, third, to briefly examine the spectrum of possible modern analogues that might describe the Scottish palaeo-environment best.

Methodology

A combination of geomorphological field mapping at a scale of 1 : 10 000 and aerial photograph interpretation at a scale of 1 : 25 000 was used to produce high-resolution geomorphological maps and to determine the precise location and planform of glacial, in particular ice-marginal, landforms (cf. Lukas 2003). Sedimentological logging of available sections on square millimetre paper and on overlays of enlarged photomosaics of the cleaned sections, or a combination of the two, was used to ensure maximum planimetric accuracy of the final logs. Individual sedimentary units were identified on the basis of their visual physical properties

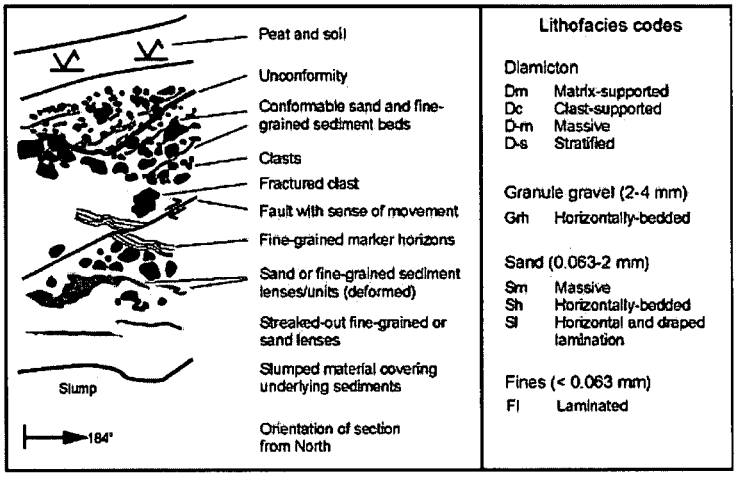


Fig. 2. Lithofacies codes and symbols employed in the section logs of this article.

including grain size range, compaction, sedimentary structures and clast shape following the guidelines detailed by Evans & Benn (2004) and logged utilizing a slightly modified version of the lithofacies code introduced by Eyles *et al.* (1983) (cf. Evans & Benn 2004; Fig. 2). The dip and strike of selected units, fold axes, fault planes and rectilinear proximal slopes was measured using a Recta compass-clinometer.

Study area

The study area comprises *c.* 1000 km^2 of the NW Scottish Highlands in the county of Sutherland (58°5′–58°29′N; 4°58′–4°34′W; Fig. 3). The eastern and larger part of the area is underlain by Moine psammitic and pelitic schists, which contain frequent quartz veins (Johnstone & Mykura 1989). To the north and west, these rocks have been transported over unmoved Lewisian Gneiss ('basement rocks') to form a series of nappes bounded by thrust planes grouped together in the Moine Thrust Zone (Johnstone & Mykura 1989). Bedrock is at or near the surface in *c.* 70–80% of the area and its lithology and structure exerts a large influence on the distribution of glacial erosional and depositional landforms (e.g. Bradwell 2005). All moraines dealt with in this article are formed on and from meta-sedimentary Moine rocks. Depth to bedrock from the bottom of exposures is generally 1 to 5 m.

Moraine characteristics

Isolated moraine mounds and ridges are widespread features that are usually confined to the valley bottoms and the lower *c.* 50–100 m of adjacent slopes. These moraines are usually aligned in chains that can be connected to form continuous arcs and have been interpreted as individual palaeo-ice margins elsewhere (cf. Benn 1990, 1992; Benn *et al.* 1992; Bennett & Boulton 1993; McDougall 2001; Lukas 2003, 2004). These arcs in turn have close spacings of only a few metres to tens of metres and sometimes appear 'stacked'.

A survey of the form and symmetry patterns of moraines has shown that the larger number of moraine ridges and mounds displays a clear asymmetry with a clear steeper proximal and gentler distal slope. Due to blanket peat cover, however, the exact planform of individual moraines (fan or ridge-shape) cannot be reconstructed with confidence. Of 118 moraines surveyed, 86.4% contain clear proximal rectilinear slopes. The distribution of slope angles (Fig. 4) is similar to those reported from Svalbard and other isolated sites throughout upland Britain, with 43.2% of slopes in the classes 28–33° (cf. Bennett *et al.* 1998; Graham & Midgley 2000; Graham 2002). As detailed sedimentological evidence from natural and artificially created exposures is widely available, contrasting to most other areas, this presents ideal conditions for testing which model best explains the formation of Scottish 'hummocky moraine'.

Seven examples of exposures through the whole width of moraines within an area covered by a Younger Dryas ice cap are presented below. These examples are representative of 49 sections described so far and demonstrate the range of facies and internal architecture found in ice-marginal moraines throughout the study area.

Case study 1: Terrestrial ice-contact fan

Description

Moraine ridge SLK 1 (NC 393 344; 215 m a.s.l.) is located at the foot of the western slopes of Ben Hee, where it forms part of a larger series of moraines stacked on the mountainside. In the valley bottom,

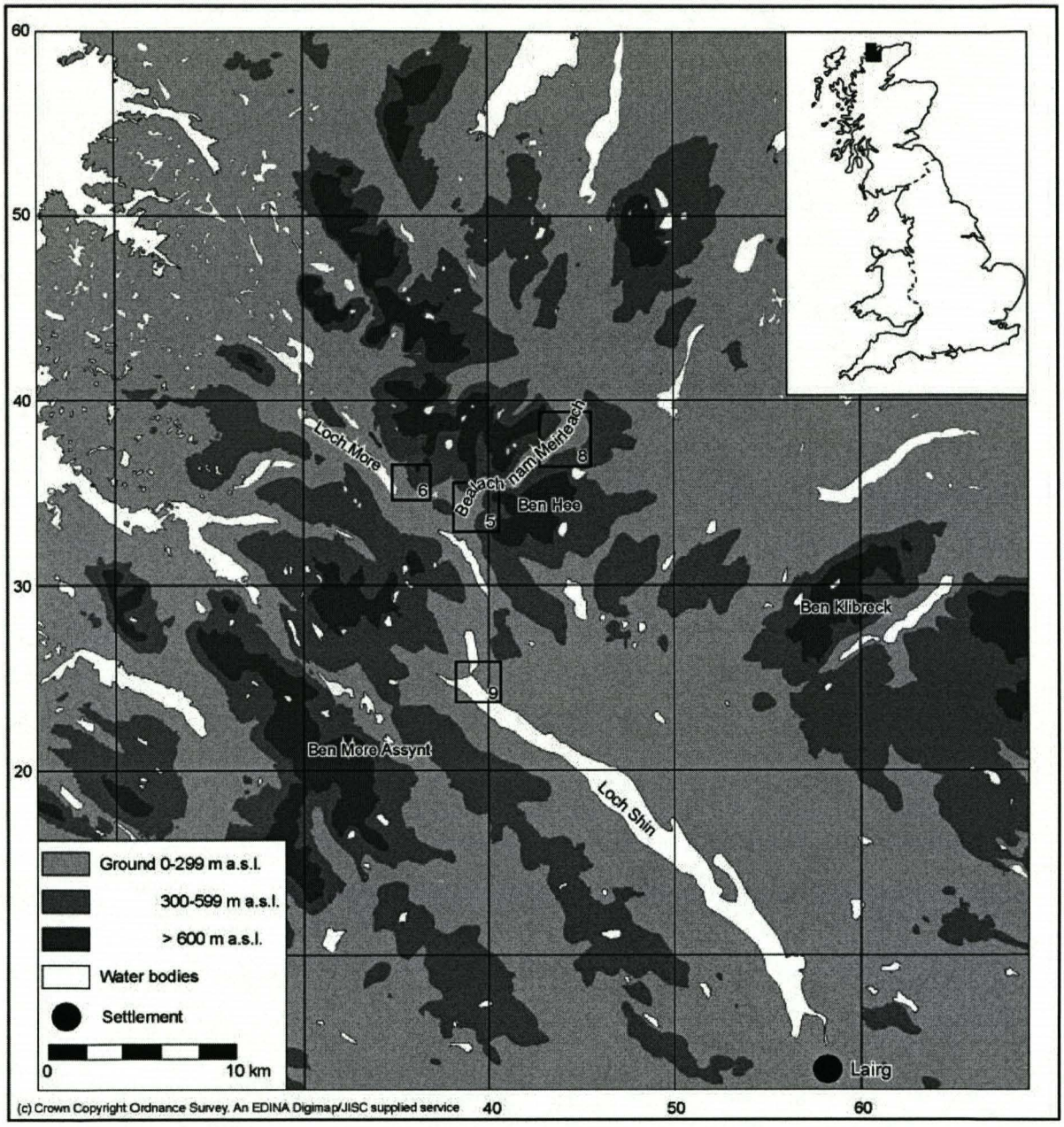


Fig. 3. Location of the study area in NW Scotland showing places mentioned in the text; numbered frames mark the location of detailed geomorphological maps.

a series of recessional moraines can be found. Both suites of moraines are partly breached and/or bounded by glacial meltwater channels (Fig. 5A). The moraine is *c.* 6 m high, 10 m wide, 30 m long and displays an asymmetrical cross-profile with a steeper ice-proximal rectilinear slope (*c.* 32°) pointing NNW and a gentler distal rectilinear face (*c.* 16°) dipping SSE. This latter gentle surface grades into the steep ice-proximal slope of the next moraine upslope that shows a similar cross-section and is of similar dimensions. The overall direction of ice retreat is reconstructed to have been towards the N and NNW in this area (Fig. 5A–C).

A trackside pit provides excellent exposure through the whole moraine perpendicular to its crestline. Sediment characteristics allow grouping into two facies associations, those on the left (NW) side and those on

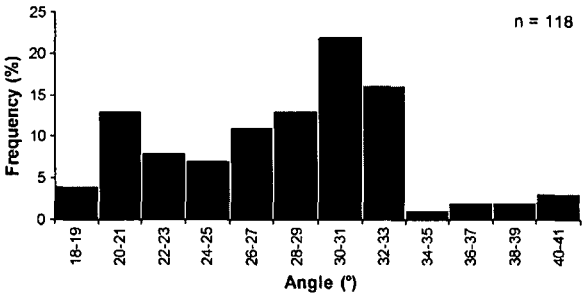


Fig. 4. Distribution of slope angles of proximal rectilinear slopes of 118 moraines in the NW Highlands. Only 16 of the 118 moraines (= 13.6%) measured displayed an irregular, undulating slope. Note the peak of slope angles around 30°. The classes follow the approach by Bennett *et al.* (1998) and Graham & Midgley (2000).

the right (SE) side. The latter consists of alternating units of matrix-supported to clast-supported stratified diamictons that reach thicknesses up to 1.1 m with the thickest units towards the top near the crestline (Fig. 5C). These diamict units often taper out over a few decimetres in a distinct wedge shape. Individual units are often separated by fine to medium sand units that are commonly between 0.01 and 0.1 m (Fig. 5D) but can reach thicknesses of 0.7 m, especially nearer the crestline of the moraine. Few interspersed fine to coarse gravel and occasional pebble-sized particles occur throughout the sequence near the crestline. Thinner beds of sand often bend around the underside of embedded larger clasts. The sorting within the sand units increases while their thickness decreases towards the right (SSE) (cf. Fig. 5C, D). In a similar fashion, the sand units show fining from predominantly medium and coarse sand near the crestline to silty very fine sand towards the right. On the far right-hand side of the section, internal stratification within the finer-grained units is apparent and reveals a sequence of gently dipping, subparallel, 2 to 5-mm-thick beds of silt and silty very fine sand unconformably overlain by conformable horizontal units of fine to medium sand and clast-supported diamicton (Fig. 5D). Despite these small-scale variations, all of the units described above are inclined subparallel to the surface slope of the moraine (*c.* 16°).

In a zone that starts at the crestline and steeply dips to the NNW (left) by *c.* 32°, the dip angle and dip direction of units changes abruptly over the space of 0.1 m and individual units interdigitate. The layer within which the units dip more steeply is parallel to the left (NNW) slope of the moraine and of an almost constant thickness of *c.* 1.2 m. This side of the section, in contrast to the right, shows steeply (*c.* 32°) inclined layers of clast-supported stratified diamicton that are intercalated with very thin continuous sand layers which taper out in sharp wedges in a downslope direction (Fig. 5C).

Interpretation

The alternation of diamict and sorted units on the right side of the moraine together with their uniform and gentle dip is consistent with an interpretation as supraglacial debris flow units and fluvial 'wash' horizons, respectively. Distinct wedge shapes at the downslope ends of diamict units are interpreted as flow noses indicative of the deposition in slow coherent debris flows (cf. Lawson 1982, 1988; Benn 1992). The alternating deposition of such units can be explained by a combination of two factors, namely a stationary ice margin and differences in water content, the latter of which determines whether gravitational or fluvial processes dominate in fan formation at any given time (cf. Krzyszkowski & Zielinski 2002). Laminations within the sand units and their laterally discontinuous fashion are compatible with deposition in sheets and shallow channels or rills, respectively (cf. Lawson 1988). Thin, inclined beds of silt and silty very fine sand on the far right side are interpreted as a puddle fill that then got topped by slightly coarser sand during a subsequent sheet wash event. The fine internal stratification of the sand units, preservation of the underlying delicate silt layers and limited bed thickness suggests slow sedimentation rates, and is compatible with deposition in a supraglacial setting on a relatively gentle fan surface (cf. Gripp 1975; Krüger 1997; Krzyszkowski & Zielinski 2002). The decrease in thickness and grain size within the sand units can be explained by waning flow velocity leading to downfan fining, a phenomenon commonly observed in ice-marginal fans (Zielinski & van Loon 2000). Downwarping of thinner sand beds underneath clasts is interpreted as evidence of clasts sliding and falling from the ice surface onto the former fan surface causing localized deformation.

The steeper slope on the left side is interpreted as the former ice-contact face that formed during collapse of the proximal part of the units as support by the ice was withdrawn during ice retreat. Limited compaction of the diamict units, clear stratification and interdigitation of sand and diamict units within a narrow, well-defined zone all support this interpretation. The lack of collapse features in this zone suggests that no dead ice bodies were present at the time of glacier retreat and that this ice-contact face is merely the result of withdrawal of support by the ice margin.

The moraine is thus interpreted as a terrestrial ice-contact fan following Benn (1992), Benn & Evans (1998), Zielinski & van Loon (2000) and Krzyszkowski & Zielinski (2002). This form is a scaled version of those classed as Type A (mass flow deposits – dominated ice marginal fans) by Krzyszkowski & Zielinski (2002), in that it bears striking similarities in its sedimentological composition and characteristics, but is about an order of magnitude smaller. The morphological asymmetry thus reflects a steeper proximal ice-contact

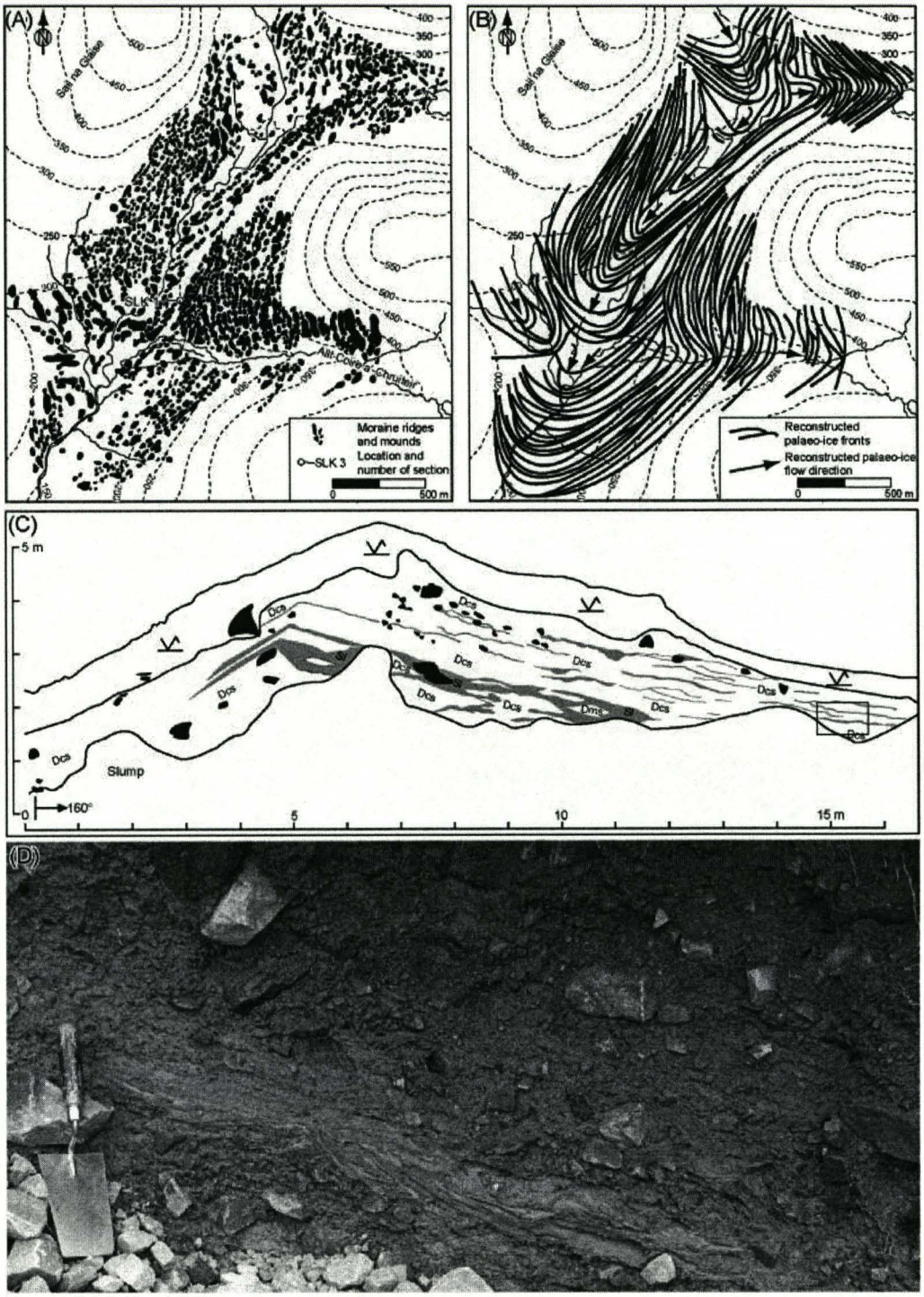


Fig. 5. A. Geomorphological map of the southern part of Bealach nam Meirleach showing 'stacked' moraines on the SW slopes of Ben Hee and location of SLK 1. B. Reconstruction of palaeo-ice fronts and ice-flow directions. C. Section log of section SLK 1; for key, see Fig. 2. D. Close-up photograph of units at the distal side shown by frame in (C).

slope with material at the angle of repose and a much gentler fan surface representing the depositional slope.

Case study 2: Deformed ice-contact fans formed in succession

In the lower reaches of Coire Eas na Maoile (Fig. 6A, B), two asymmetrical moraine ridges contain exposures through almost their whole width and height, initiated by fluvial undercutting. The setting and quality of the exposures presents a rare opportunity to study the sedimentology of two moraines and test whether they formed in succession or during one event.

Description of moraine SLK 13

The first moraine (NC 3575 3477; 72 m a.s.l.; SLK 13) is asymmetrical and bounded by rectilinear slopes with gentle (15°) northern distal and steeper (32°) southern proximal gradients. An enlarged exposure reveals alternating units of clast-supported, stratified diamictons (A) and interbedded units of (B) fine sand to granule-sized gravel, (C) openwork granule to coarse gravel, and (D) well-sorted very fine to fine sand with occasional silt layers (Fig. 6C). The rightmost 3 m of the moraine is not exposed and the top is partly obliterated by vegetation and peat. In general, these sediments are similar to those described in Case study 1. Unit (A) contains subrounded clasts with a-axes up to 0.8 m that contain numerous facets. Layers of this unit attain thicknesses of 0.5 m and are particularly well developed in the left (northern) part of the exposure, where they dip at *c.* 15° subparallel to the moraine surface. Interbeds of sorted sediments (B–D) are rare and only reach a thickness of 1–2 cm. A discontinuous layer of alternating units B and D is exposed in the lower parts of the sequence and shows signs of gentle folding that is clearly visible at the contact with surrounding layers of A and B (Fig. 6C). Lenses and more continuous layers of units B and D that display folding occur scattered throughout the sequence.

Interpretation of moraine SLK 13

The succession of stratified diamict and sorted finer-grained units is fully compatible with an interpretation as a terrestrial ice-contact fan formed at a temporarily stationary ice margin during overall southward retreat. Gravel and granule units suggest that flow events larger than those of the Hochsander-type introduced above occurred occasionally. However, likewise, these gravel lenses might represent lag bodies created by winnowing of fines during clean water runoff off the glacier surface (Lawson 1988). Deformation structures within the finer-grained unit (D) are indicative of proglacial push from the right (south). This episode of lateral compression is inferred to have occurred after fan formation and initial retreat during a small-scale readvance of the glacier. This is supported by the largely unaltered fan morphology with a clear proximal rectilinear ice-contact face and a distal fan surface, but also by the fact that the sedimentary units are largely undisturbed. Similar relationships between fan formation along a stationary ice margin and proglacial deformation during a subsequent readvance were noted for much larger ice-contact fans in Poland by Krzyszkowski (2002) and by Benn (1992) in terrestrial ice-contact fans on the Isle of Skye.

Description of moraine SLK 12

This moraine (NC 3582 3465; 75 m a.s.l.; SLK 12) merges directly upvalley with the proximal (right) slope of SLK 13 described above (Fig. 6A). It too displays a clear asymmetry with a gentle distal rectilinear face dipping to the left (north) by *c.* 20° and a steeper proximal rectilinear face dipping by *c.* 27° to the right (south). Three lithofacies units can be distinguished (Fig. 6D):

(1) A lower sandy-gravelly, clast-supported, stratified diamicton with clast a-axes $\leq$0.4 m. This diamict unit is of medium compactness due to interlocking clasts within clast clusters, voids in which are sometimes infilled by finer material. Laterally discontinuous lenses of sorted silty to very fine sandy, matrix-supported, stratified diamicton with most clasts in the granule and fine gravel fraction and maximum thicknesses of 0.1 m enhance the stratified appearance of the lower diamict which dips by *c.* 10° towards the left (north).

(2) More prominent, but discontinuous, layers of massive very fine to medium sand with occasional coarse sand grains reach thicknesses up to 0.1 m. They occur throughout unit (1) and contain numerous deformation structures, e.g. small-scale overturned folds and undulating contacts with the surrounding diamicton (Figs 6D, 7).

(3) A unit of massive very fine sandy silt can be traced through much of the lower part of the exposure. It reaches a maximum thickness of 0.6 m and contains numerous signs of pervasive deformation on a variety of scales (Figs 6D, 7). The smallest examples are 5-cm-wide dykes of sand pointing upwards into the surrounding diamicton by as much as 0.1 m. Internally, such dykes show widespread wavy laminations that are frequently disrupted, particularly around clasts. Small-scale reverse faults with a maximum displacement of 2 cm are frequent at the contact with the lower diamict. At the larger scale, the silt layer is partly folded and bifurcates towards the lower left side (north) of the section. The upper branch is heavily disturbed into an overturned fold.

The lower part of the moraine is cut unconformably by an upper sandy clast-supported, stratified diamicton (unit 4 in Fig. 6D) which extends into the lower

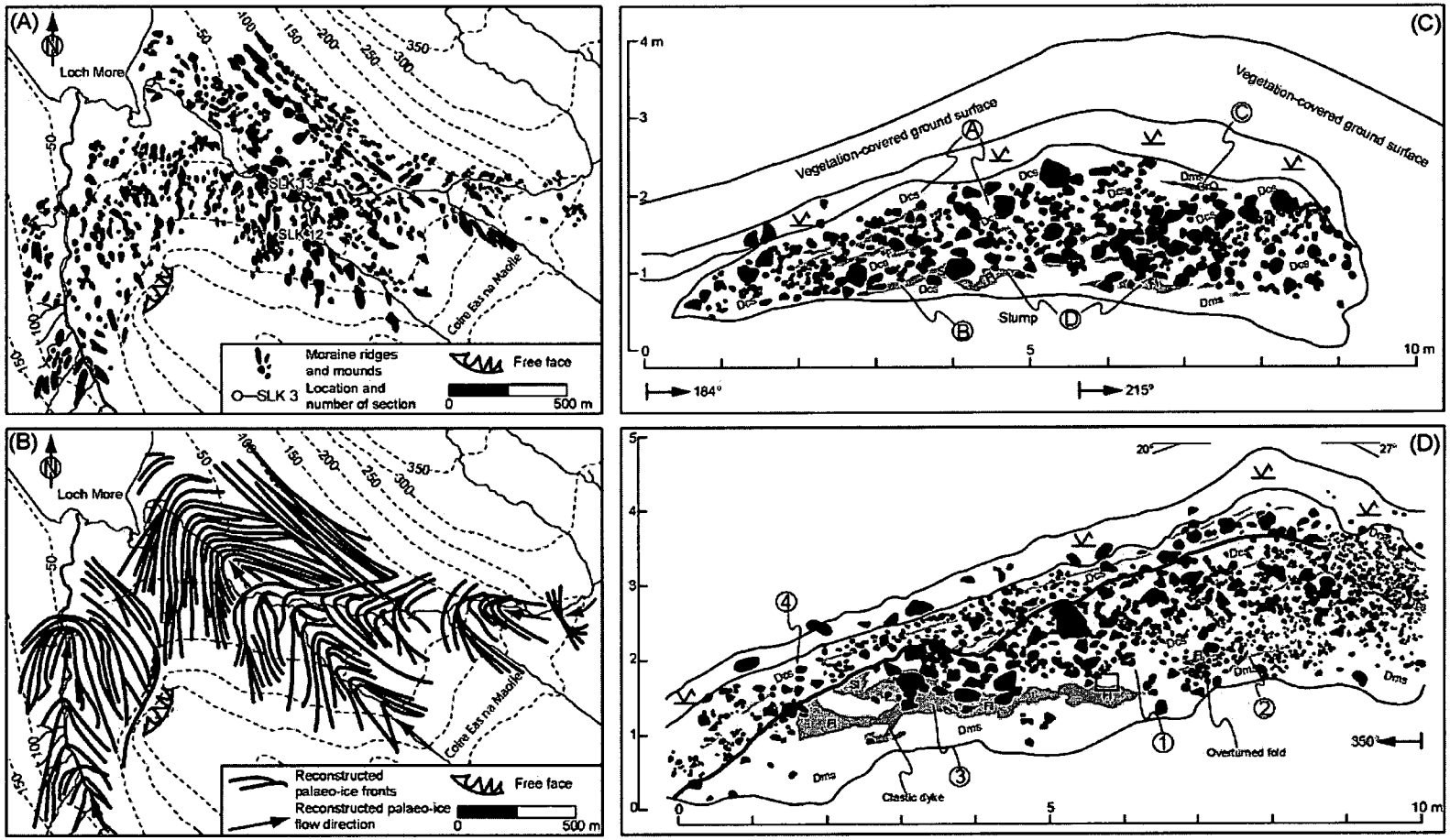


Fig. 6. A. Geomorphological map of Coire Eas na Maoile showing moraine mounds and ridges and the locations of SLK 13 and SLK 12. B. Reconstruction of palaeo-ice fronts and ice-flow directions. C. Sedimentary log of section SLK 13; for key, see Fig. 2. D. Sedimentary log of section SLK 12; for key, see Fig. 2. Circled numbers and letters refer to units mentioned in the text.

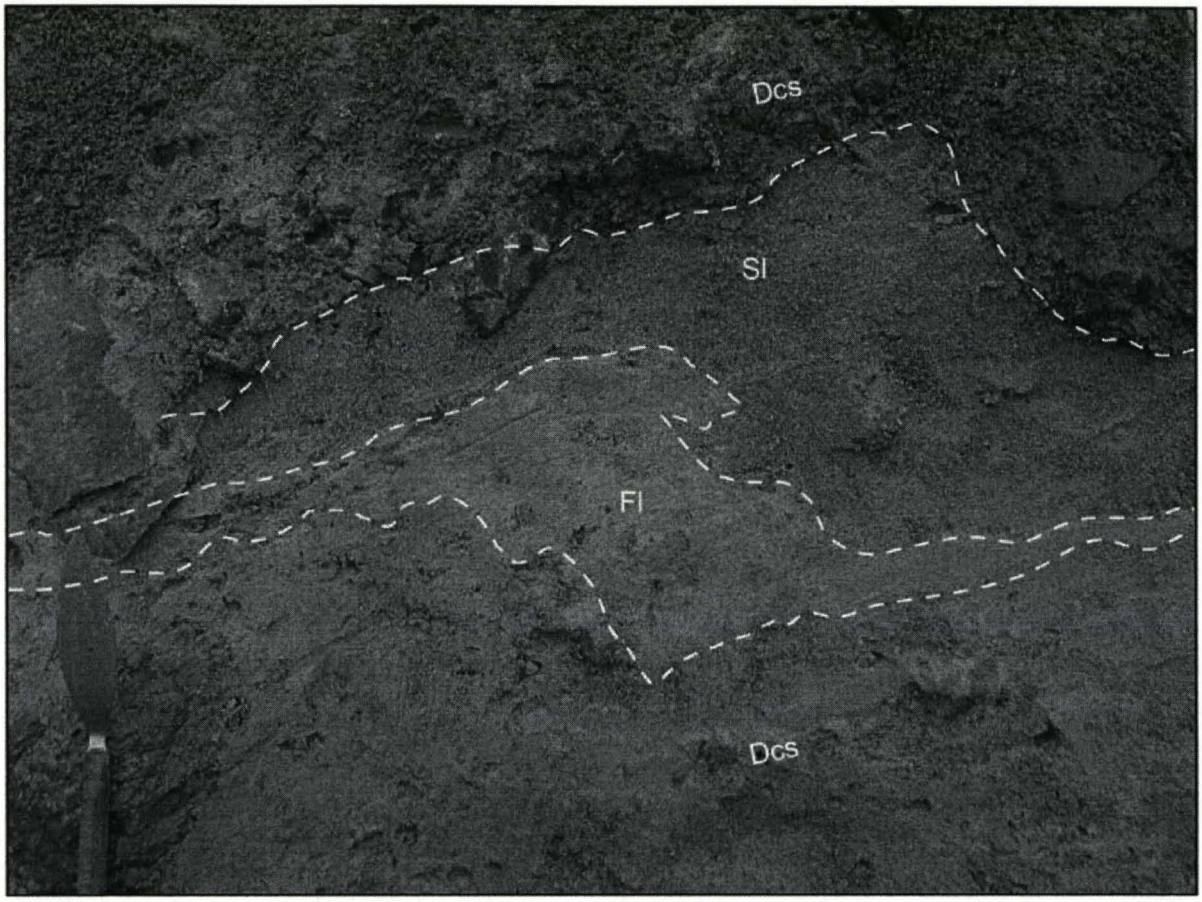


Fig. 7. Close-up photograph of overturned fold at the deformed boundary between diamicton (Dcs) and underlying laminated sand (Sl) and silt (Fl) layers. Note the occurrence of thin silt and fine sand lenses in the lower diamict. Location given by frame in Fig. 9. The upper metal part of the scraper is 7 cm long.

diamicton in a series of channel and half-channel structures. This erosive unconformity is most obvious where the upper diamicton truncates the branches of unit (3). Part of the overturned fold described above is folded back on itself along this contact. The upper diamict unit dips by *c.* 8–10° towards the left (north); deviations from the angle of dip of the underlying diamict are apparent.

Interpretation of moraine SLK 12

The facies in the lower part of this moraine are interpreted as supraglacial debris flows and fluvial deposits, respectively, and are indicative of a terrestrial ice-contact fan formed at a stationary ice margin during overall retreat. Interlocking of boulders is inferred to result from blocks sliding and toppling off the ice front and successive blocks becoming trapped on their stoss side. The infill by relatively coarse sediments (no silt) is interpreted as a result of trapping of sediment transported in shallow flows in the lee of such boulders.

The deformation structures are indicative of proglacial deformation and are interpreted as evidence of a glacier readvance into this fan following formation and initial retreat. The continuous silt layer has probably acted as a décollement surface during push as inferred from the concentrations of deformation structures within, and in the near vicinity of, this layer. In particular, the numerous injected layers of sand into the surrounding diamict units are indicative of liquefaction and hydrofracturing and are interpreted as water escape structures. They form as a result of compression under impeded drainage conditions (cf. Rijsdijk 2001; van der Meer *et al.* 1999) or due to differential loading, such as during deposition of debris flow on top of a thin aquifer (e.g. Lawson 1988). It appears that the surrounding diamict and silt was sufficiently dense to over-pressurize water contained within thin sand lenses in their vicinity to cause widespread hydrofracturing.

The water escape structures are comparatively small and support an interpretation of localized stress transmission during a short-lived readvance.

The upper diamicton that truncates the top left of this prominent silt layer is interpreted as a later debris flow event that also partly eroded the existing fan surface. Basal traction gravels and erosive scours are sedimentary features compatible with such an interpretation (cf. Lawson 1981, 1982, 1988; Benn 1992; Krzyszkowski & Zielinski 2002). This addition of new material is likely to have caused some of the surficial folding of the upper parts of the silt layers as a response to drag forces exerted by cohesive debris flows.

This terrestrial ice-contact fan thus formed by a similarly two-staged process as SLK 13 described above. In this case, however, the readvance was longer-lived, causing severe disruption of the sediment pile along a basal décollement surface. Additional debris flows were also able to modify the pre-existing fan surface.

Significance of SLK 12 and SLK 13

The sedimentary architecture of both moraines demonstrates that each moraine represents a terrestrial ice-contact fan and is the product of a two-staged process of (1) fan formation and (2) subsequent deformation during a readvance. This indicates an active, incremental and oscillatory retreat mode. Both moraines also display a marked asymmetry with steep proximal and gentle distal slopes, the proximal slope reflecting the former ice-contact face with material at the angle of repose. Thus, these moraines demonstrate that the moraines formed sequentially in two separate episodes of fan formation and deformation. This evidence is clearly incompatible with the englacial thrusting model, which interprets rectilinear ice-proximal slopes as thrust planes rather than angle of repose debris slopes, and argues that multiple moraines form contemporaneously during one glacial event.

Case study 3: Dislocated ice-contact fans

Description

This moraine complex at NC 4385 3827 in the northern parts of Bealach nam Meirleach (230 m a.s.l.; SLK 3) consists of three individual moraine ridges with rounded crestlines linked by peat-infilled basins (Fig. 8A). The complex is *c.* 7 m high at its highest point, up to 10 m wide and 12 m long. Most of the moraine complex has been disturbed by quarrying activity so that data on the cross-sectional shape (e.g. asymmetry) cannot be presented. A section orientated subparallel to the crestline exposes the uppermost 2 m of the main mound.

Main section. – Three lithofacies units can be distinguished (Fig. 8C). (1) A succession of sandy clast-supported, stratified diamicton units that attain maximum thicknesses of *c.* 1.5 m and dip by up to 52° towards the left (west to WNW). Contacts between individual beds are largely conformable. Intercalated within the diamictons is unit (2), a *c.* 0.4-m-thick very fine to coarse sand that frequently contains lenses of very fine sand to silt with occasional fine gravel and few outsized clasts. A sharp basal contact striking 198° and dipping towards the WNW by 68° separates it from a unit of clast-supported, stratified diamicton. An isolated lens of horizontally stratified granule gravel is embedded within this unit and shows signs of soft sediment deformation (Fig. 8C). Deformation structures occur in a *c.* 15-cm-wide zone, visible in lenses of silt that are attenuated downwards. Unit (3) is a loose, clast-supported, massive diamicton with a maximum thickness of 1.3 m. It is overlain by a subhorizontal loose, sandy, clast-supported, stratified diamicton that attains maximum thicknesses of 0.7 m, contains subrounded and subangular clasts with a-axes up to 0.5 m and forms a continuous carapace on top of the sediments exposed in the pit described below.

Pit section. – In a smaller pit dug both parallel and at right angles to the crestline to supplement the observations, four units can be distinguished (Fig. 8C). Unit (A) corresponds to (1) described above and caps the whole frontal wall. (B) is a matrix-supported stratified diamicton with a maximum thickness of 0.8 m and few clasts with a-axes <3 cm interspersed in a medium to coarse sandy matrix. Subhorizontal stratification is apparent through discrete layers of medium sand-sized particles. A few deformation structures, best described as flexures, occur in the upper left-hand part of this unit. Contacts with the embracing units are sharp and at least the lower one could be described as erosional due to the presence of channel structures. (C) is a 0.6-m-thick body of planar cross-bedded fine to medium sand that contains some deformation structures, e.g. high-angle reverse faults with displacements of up to 2 cm. A sharp contact is formed with the underlying unit (D), a sandy, clast-supported, stratified diamicton that is exposed at the bottom. These four units are also exposed in the sidewalls, where they dip more steeply towards the NNE (Fig. 8C). Here, unit (A) tapers out after *c.* 0.1 m from the left side of the right (southern) wall, the remainder being occupied by units (B) and (C), which both contain deformation structures. The upper contact of unit (B) with (A) is sharp and relatively straight; the lower contact by contrast is highly undulating and contains several smaller fold traces. Steeply-inclined planar cross beds occur within this unit, and the lower contact strikes 290° and dips with 55° to the NE. Unit (C) contains both low-angle normal and reverse faults cut steeply dipping planar cross beds, and high-angle reverse fault sets subparallel to the cross beds that are displaced by between <1 cm and 4 cm. Normal *en echelon* high-angle fault sets strike 328° and are inclined between 40° and 60° towards the

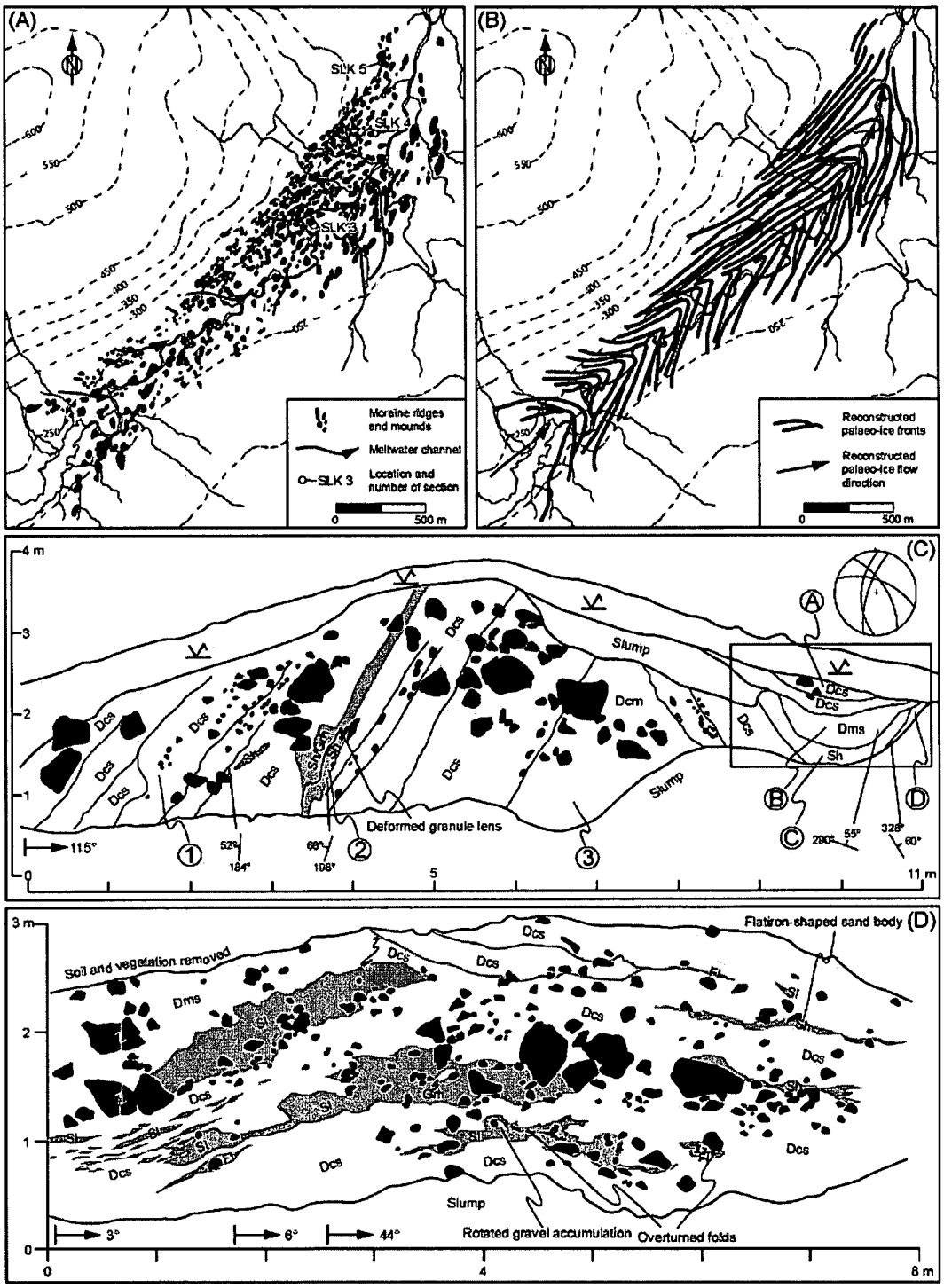


Fig. 8. A. Geomorphological map of the northern part of Bealach nam Meirleach showing moraine mounds and ridges and the locations of SLK 3 to SLK 5. B. Reconstruction of palaeo-ice fronts and ice flow directions. C. Sedimentary log of section SLK 3; for key, see Fig. 2. The location of the frame shows the location of the pit section described in the text. D. Sedimentary log of section SLK 4; for key, see Fig. 2. Circled numbers and letters refer to units mentioned in the text.

NE and show similar displacement. Their spacing between 0.5 and 2 cm gives a fragmented appearance to individual inter-fault blocks.

Interpretation

The sedimentological composition of this moraine is compatible with an interpretation as a terrestrial ice-contact fan. However, the steep inclination of the units (in the left part of the section to the WNW, in the pit towards the NE) cannot be explained in terms of the primary depositional slope. The wealth of deformational information contained within unit (C) and the sharp contact between sandy and diamicton units in the main section interpreted as a high-angle normal fault – based on the presence of drag features in silt lenses in the hanging wall (Fig. 8C) – indicate ice pushing into the sediment pile, resulting in a dislocation of the entire fan. Different scales of deformation can be distinguished – small-scale faults and folds as recognized in unit (C) and large-scale steepening of the entire landform.

This evidence suggests a readvance of the glacier into the previously formed terrestrial ice-contact fan which was probably more sustained than in the cases reported above. The deformation history contained within the sediments indicates a two-stage process: First, the smaller-scale compressional reverse fault sets and large-scale steepening of the units were formed by forward push. Second, a subjugate set of normal faults developed in both parts of the section and is interpreted as relaxation response following steepening of the ice-contact fan by the glacier readvancing from a southern direction (cf. Humlum 1985). It appears to mark a stage following glacier readvance during which slippage occurred in the sediment pile as a response to the considerable oversteepening of a much shallower depositional slope of *c.* 10–20° to the present *c.* 60°. The exact direction of ice push cannot be reconstructed due to the limited extent of the exposure but ice flow from the SE quadrant as suggested by structural data (Fig. 8C) is fully compatible with the available geomorphological evidence from the surroundings (Fig. 8B).

Case study 4: Partly overridden and glaciotectonized ice-contact fans

Description

A moraine ridge (NC 4422 3871; 220 m a.s.l.; SLK 4), *c.* 4 m high, 15 m wide and 45 m long, with a straight crestline orientated NW–SE (126°), has been excavated for gravel extraction along a width of 6 m. Owing to removal of the upper metre of sediment and peat, the original asymmetrical cross-profile with a steeper side to the left (SW) and a gentler slope on the right (NE) of this moraine is not discernible on the sedimentary log. The sediments exposed in this ridge can be grouped into two distinct associations, one in the left third and one occupying two-thirds of the right-hand side (Fig. 8D), the junction of which coincides with the crestline. The units on the right represent facies similar to those reported above and consist of alternating loose clast-supported, stratified diamicts and sorted sediments, mostly sand and very fine silt, that both dip gently by about 11° to the right (NE). The diamict units reach thicknesses up to 1.0 m, with only occasional thin stringers of silt or very fine sand, while larger boulders with a-axes $\leq$0.5 m are frequently found within the diamict units. Finer-grained sediment units in the right-hand part of the section display numerous deformation structures like small-scale overturned folds with a concentration of the most severe deformation structures near the bottom of the exposure (Fig. 8D). Here, a formerly subhorizontal bed of gravel accumulated in the lee of a clast has been rotated and now dips by about 40° to the right (NE). Near the right-hand surface, a sand bed with a distinct flatiron shape can be found underneath a clast with a flat underside, and this in turn is underlain by a continuous 10-cm-thick layer of very fine sandy silt that shows localized folding in the vicinity of the clast.

In the left third of the section, the same lithofacies units are capped by 30 cm of matrix-supported, stratified diamicton (Fig. 8D). Clasts in the diamict units are distinctly fractured by reverse faults that are filled with silty very fine sand. All three lithofacies units consist of elongated and attenuated lenses and contain flame structures throughout, undulate across the width of the left-hand side and dip by about 29° to the left (SSW). One isolated sand body contains a massive silt bed with overturned folds that display small reverse faults in their hinge-lines (Fig. 8D).

Interpretation

The alternation of the three lithofacies units described above is compatible with an interpretation as terrestrial ice-contact fan deposits. The left-hand side of the exposure contains evidence for much more widespread and severe deformation that is not readily reconcilable with proglacial compression. The presence of boudins and flame structures, fractured rock with intervening silt layers and a gently undulating appearance in this part of the moraine is interpreted as subglacially sheared during partial overriding. Stresses transmitted into the distal (right) part of the fan were large enough to create the deformation structures observed there. These in turn are compatible with proglacial deformation caused by ice pushing into, and partly overriding, this fan from the left (SW) during a readvance, causing widespread folding. This lateral compression also led to the rotation of previously gently inclined units, i.e. the gravel accumulation in the lee of a clast in the lower part. Rotation of one clast near the right-hand

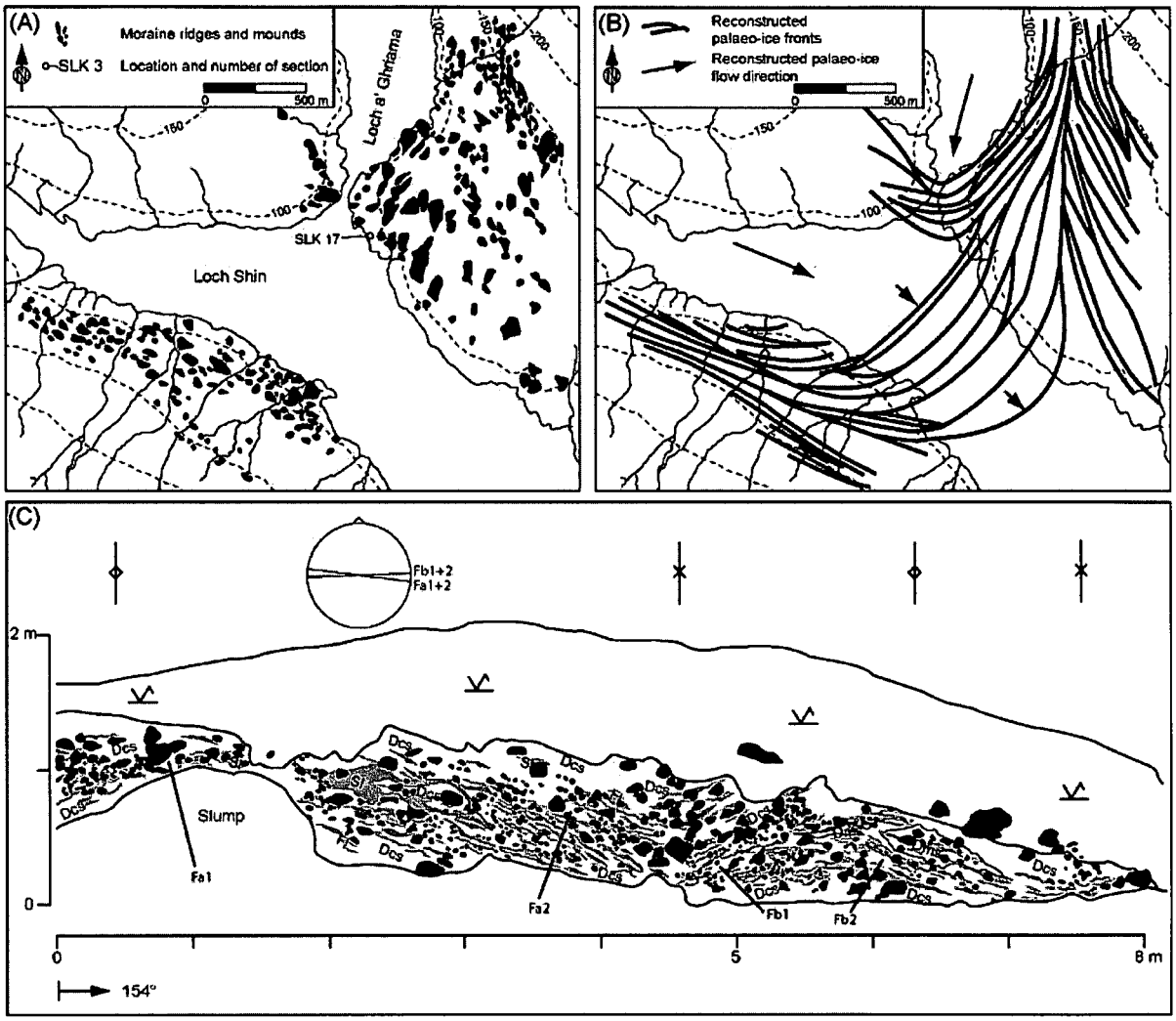


Fig. 9. A. Geomorphological map of the northern part of Loch Shin showing subdued moraine mounds and ridges and the location of SLK 17. B. Reconstruction of palaeo-ice fronts and ice flow directions. C. Sedimentary log of section SLK 17; for key, see Fig. 2. Approximate position of synclinal and anticlinal fold axial planes is shown as viewed.

surface that ploughed material up in front of it to form the flatiron shaped sand body was possible by localized décollement within the underlying silt layer. Close to the crestline, a variety of both ductile and brittle deformation structures indicates the location of the transition zone between subglacial and proglacial deformation.

Case study 5: Completely overridden ice-contact fans

Description

Moraines along the eastern shore of Loch Shin deviate from the general character of those described above in that they are only between 2 and 3 m high, up to 100 m long and 30 m wide; they show an almost symmetrical, regular profile with maximum slope angles of 12°, giving them a subdued or smoothed appearance. They form discrete moraines that are somewhat wider than elsewhere in the area and appear to be much more continuous features transverse to former ice flow. Depth to bedrock cannot be established due to obliteration by beach gravel and the loch level. Unfortunately, this area is largely forested, hindering clear mapping of smaller moraines to be carried out on the ground and from aerial photographs (Fig. 9A, B).

Exposures in these moraines along the Loch shore (NC 3918 2495; 100 m a.s.l.; SLK 17) reveal three alternating lithofacies units that are exposed along a width of *c.* 8 m (Fig. 9C). Unit (1) is a loose, stratified, clast-supported diamict (Dcs) that reaches a maximum

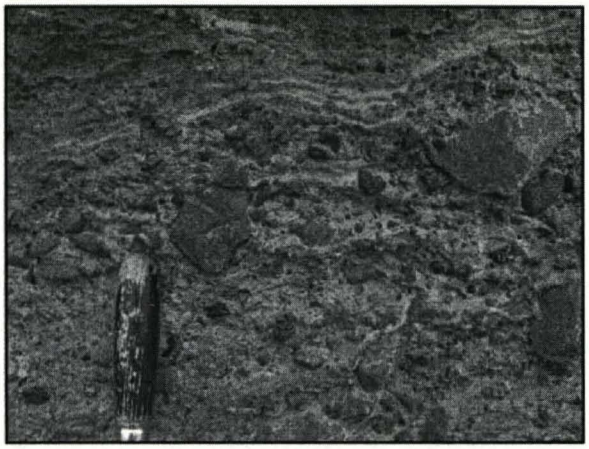


Fig. 10. Close-up photograph of glaciotectonized debris flow and fluvial wash units exposed along the eastern shore of Loch Shin. The wooden handle of the trowel is 10 cm long.

thickness of 0.2 m and frequently contains clasts with maximum a-axes lengths of 0.4 m. Unit (2) is a slightly silty fine sand (Sl, Fl) that contains occasional outsized granule clasts, exhibits wavy laminations and reaches a maximum thickness of 0.1 m. Both units exhibit frequent lateral thickness changes and take the general shape of elongated or streaked-out lenses or pods. Frequently, beds of unit (2) bifurcate around clasts or lenses of unit (1), showing partitioning and rejoining, while individual beds appear attenuated, tapering out within the space of a few millimetres on either side. Although discontinuous, individual lenses of the same unit are laterally aligned while the units themselves appear to be subparallel to each other (Fig. 10). These units form a series of long-wavelength, low-amplitude anticlines and synclines which do not reflect the surface of the moraines (Fig. 9C). Fold axis measurements shown in the inset of Fig. 9C indicate an east–west orientation, roughly perpendicular to the section wall.

Interpretation

The sedimentary composition of the units and their pattern of alternation suggest that they formed as a series of terrestrial ice-contact fans during glacier retreat. The wealth of deformation structures can best be interpreted as the result of two processes. The long-wavelength, low-amplitude folding is indicative of larger-scale proglacial push, as recognized in larger exposures elsewhere (e.g. van der Wateren 1999). Proglacial push leading to this deformation was directed from the north (Fig. 9C). This implies glaciers sourced in the mountains and fits the evidence in the surrounding area well (Fig. 9A, B). The units themselves contain ubiquitous features indicative of subglacial simple shear and can hence clearly be distinguished from the lateral compression typical of proglacial environments (e.g. Hart & Boulton 1991; van der Wateren 1995, 1999; Benn & Evans 1996, 1998; Benn & Clapperton 2000; van der Wateren *et al.* 2000; Golledge 2002; McCarroll & Rijsdijk 2003). In particular, units that bifurcate around clasts or inclusions have also been observed in glaciotectonized sediments by Benn & Evans (1996). In summary, the subdued character of the moraine ridges and the out-of-phase folding of the underlying sediment is due to a three-stage process of successive fan formation, proglacial folding and subsequent overriding, either during the same, or a later, readvance.

Case study 6: Deformed outwash sediments

Description

This moraine (NC 44317 39171; 217 m a.s.l.; SLK 5) is *c.* 5 m high, 20 m long, 40 m wide and contains an exposure in a subsidiary ridge in its left-hand part (Fig. 8A, B). The ice-proximal slope is irregular and on the left (SE) in this case, while the distal slope dips by about 25° to the right (NW). Three units can be distinguished (Fig. 11). (1) Alternating very fine and fine sand layers which are horizontally stratified and contain numerous lenses of fine to coarse sand with interspersed fine gravel and very fine sand and silt couplets. The couplets usually reach thicknesses between 1 and 5 mm, while the sand layers can extend to thicknesses of >5 cm. High- and low-angle reverse and – to a lesser extent – normal faults occur frequently with the silt and very fine sand couplets acting as marker horizons (Fig. 11). The mean principal strike of the fault sets is towards the ESE (111°) with a mean dip of 35° towards the SW (Beta-axis: 193°; Fig. 11). Displacement of beds is commonly between 2 and 9 cm, with a maximum of 25 cm. Unit (1) is overlain by a clast-supported, stratified diamicton (2) with a fine to medium sandy matrix and clasts ≤0.4 m. Stratification in this unit is subparallel to that of unit (1). The contact between the two units is unconformable and undulating, forming an overturned fold in the right-hand part of the exposure (Fig. 11). In the fold core, the sand units are deformed in a brittle fashion. The contact with unit (3), a compact, clast-supported massive fine to medium gravel, which extends into unit (1) as a gently folded wedge, is partly obscured by soil formation (Fig. 11).

Interpretation

Sedimentary units (1) and (2) can be interpreted as proglacial outwash deposits indicating different flow velocities, both capped by (3) a supraglacial debris flow deposit probably deposited shortly after deposition of the outwash. The fault sets and their uniform strike and dip indicate that deformation was induced by compression and possibly ice-proximal shearing from

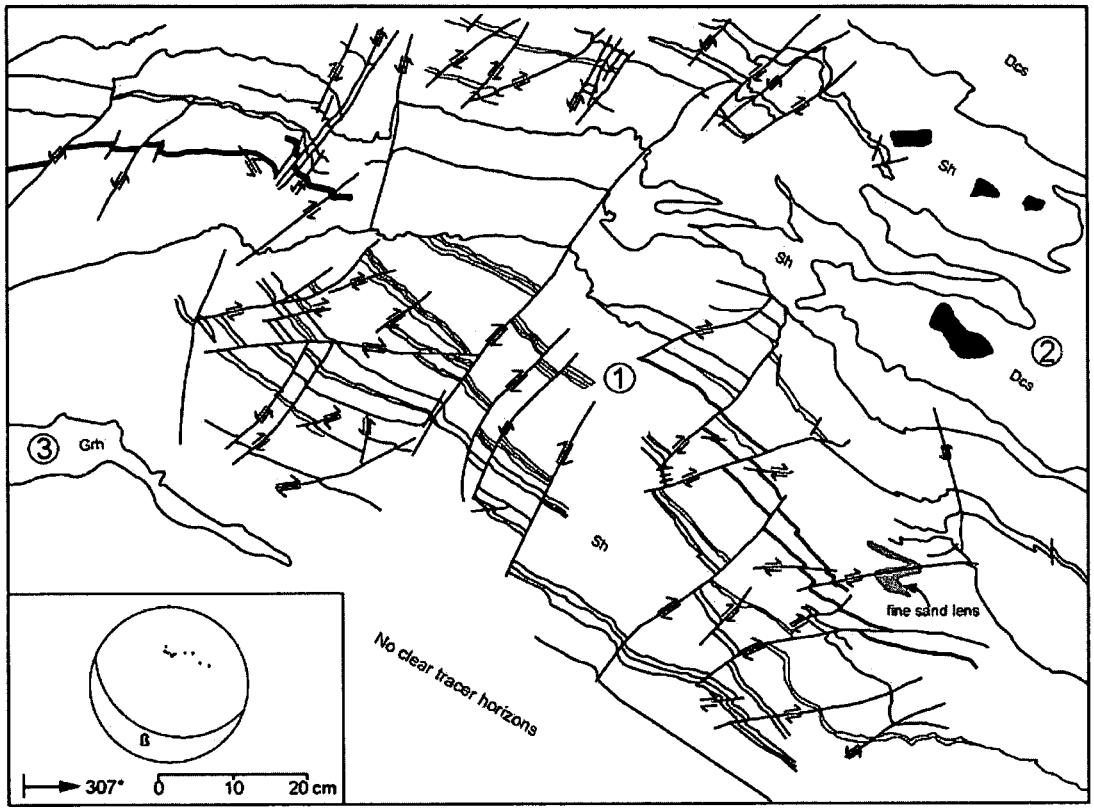


Fig. 11. Sedimentary log of section SLK 5; for key, see Fig. 2. The polar diagram represents a lower hemisphere Schmidt net (for explanation of structural data see text). Circled numbers refer to units mentioned in the text.

the SW (cf. case studies 3 and 4). The deformational evidence is compatible with proglacial deformation caused by lateral compression in which the more competent sand and fine sediment units (Sh and Fl) responded by brittle deformation, while the diamict units (Dcs) were folded (cf. Aber *et al.* 1989; Hart & Boulton 1991; Benn & Clapperton 2000; Lukas & Merritt 2004). This is interpreted as evidence for a readvance after deposition of, presumably, a valley train, the presence of which is indicated by outwash terraces in the surroundings.

Synthesis of sedimentary processes

Geomorphologically, moraine mounds and ridges are organized in discrete chains in all areas reported here. They are thus similar to those observed throughout large areas of the Scottish Highlands (Benn 1992; Bennett & Boulton 1993; Lukas 2003).

Sedimentologically, six of the seven examples reported here represent terrestrial ice-contact fans (Table 1). Numerous test diggings throughout the study area ($n = 49$), which have all yielded similar lithofacies associations, confirm that this is characteristic of the moraines in the NW Highlands. The evidence presented in case study 2 demonstrates that formation in succession is possible, which might explain the large number of asymmetrical ice-marginal moraines. In addition, the most detailed studies on Scottish 'hummocky moraine' to date (Benn 1990, 1992; Benn *et al.* 1992) recognize ice-contact fan deposits in the vast majority of exposures in moraines; similar evidence elsewhere (Lukas 2003, 2004) further strengthens the argument of almost ubiquitous terrestrial ice-contact fan formation. The deformed proglacial outwash sediments described above, however, indicate that there are exceptions, and local conditions such as the existence of ice-dammed lakes might add further variants yet to be discovered.

The frequent occurrence of deformation structures of varying intensities within the fans shows that their formation involves at least a two-staged process (Fig. 12): (A) the formation of the fan at a temporarily stationary ice margin followed by retreat and formation of a proximal rectilinear ice-contact slope (B). During a readvance (C), the fan units can be proglacially deformed and, when followed by a phase of retreat, a proximal rectilinear ice-contact slope is formed (D). During a sustained readvance, partial (E) or complete

Table 1. Classification of Younger Dryas 'hummocky' moraines in the NW Highlands and their implications for palaeo-glacier dynamics based on sedimentological composition, deformation structures and style.

Sedimentological interpretation	Deformation structures	Style of deformation	Implications for palaeo-glacier dynamics	Example[1]
Terrestrial ice-contact fan	Absent	None	Formation at temporarily stationary ice margin during overall retreat	SLK 1 (12A, B)
Deformed terrestrial ice-contact fan	Small-scale folds, low-angle reverse faults	Proglacial lateral compression	Short-lived readvance after formation	SLK 13 (12A–D)
Heavily deformed terrestrial ice-contact fan	Overturned folds, low- and high-angle reverse faults, décollement	Proglacial lateral compression	Prolonged readvance after formation	SLK 12 (12A–D)
Dislocated terrestrial ice-contact fan	Oversteepened units (dip 50–70°), high-angle reverse and normal faults, small-scale folds	Proglacial lateral compression	Prolonged readvance after formation	SLK 3 (12A–D)
Overridden terrestrial ice-contact fans	Long-wavelength, low-amplitude folding, boudins, symmetric drape around boulders, tectonic laminae, streaked-out lenses of sediment	Proglacial compression, subglacial simple shear, non-penetrative glaciotectonism	Prolonged readvance possibly long after formation, overriding of several moraines in one event	SLK 17 (12A–F)
Deformed and partly overridden terrestrial ice-contact fan	Hybrid of the above, deformation styles spatially separated	Proglacial lateral compression and subglacial simple shear	Partly overridden, partly pushed	SLK 4 (12A–E)
Deformed outwash sediments	Low- and high-angle reverse and normal faults, small-scale folds	Proglacial lateral compression	Prolonged readvance	SLK 5

[1]BGS geologist's code referred to in the text (steps of evolution shown in Fig. 12).

overriding (F) can take place, resulting in subglacial shearing of units. Complete overriding was also found to lead to the reduction of relief and glaciotectonization. The formation of Scottish ice-marginal 'hummocky' moraines can thus best be viewed as a continuum of undeformed to completely overridden and glaciotectonized terrestrial ice-contact fans with a variety of scales in between, allowing grouping into seven distinct classes at present (Table 1).

Terrestrial ice-contact fans showing deformation structures indicative of proglacial push are the most common in the study area, while glaciotectonized fans and those produced during stillstands without a successive readvance must be regarded as rarer cases. In how far these smaller-scale readvances could reflect annual mass balance fluctuations is an exciting possibility that might explain the large number of densely spaced recessional moraine arcs over small areas, but remains to be tested.

Implications for glacier dynamics

The frequent occurrence of deformation structures indicates that readvances were the norm during incremental, oscillatory glacier retreat in the Younger Dryas. Glaciers were thus close to equilibrium throughout retreat during the second half of the Younger Dryas, thus sedimentologically proving Ballantyne's (2002) suggestion based on morphological observations on the Isle of Mull.

Such oscillatory retreat is commonly found at temperate glaciers in maritime areas that react dynamically and with short response times to climate change, e.g. most outlet glaciers of the Jostedalsbre ice cap in SW Norway (Winkler *et al.* 1997) or those in Iceland (e.g. Boulton 1986; Krüger 1994; Evans & Twigg 2002).

Additionally, the absence of any disturbance caused by melt-out of potential dead ice cores within the moraines in the study area (cf. Benn 1992; Kjær & Krüger 2001) is interpreted as a strong indication for a temperate glacier environment that responded quickly enough to climate change to not leave stagnant ice bodies cut off from supply upglacier. The ubiquitous occurrence of roches moutonèes, ice-moulded bedrock, subglacial flutes, large number of moraines with large numbers of subglacially transported clasts (Lukas unpublished) and unaltered from dead ice melt-out strongly favours a temperate over a polythermal glacier regime. Hence, glaciers in Norway or Iceland are probably better analoga than those on Svalbard.

Thus, even if we cannot resolve the question whether the moraines reflect annual mass balance fluctuations, it is reasonable to assume that the high number of moraine ridges produced during the Younger Dryas in many areas of Scotland reflect moraine formation at the margins of highly active temperate glaciers with short response times.

Implications for the applicability of the thrusting model

Hambrey *et al.* (1997) and Bennett *et al.* (1998) attempt to explain the occurrence of proximal rectilinear slopes with angles near 30° in one site in Glen Torridon by

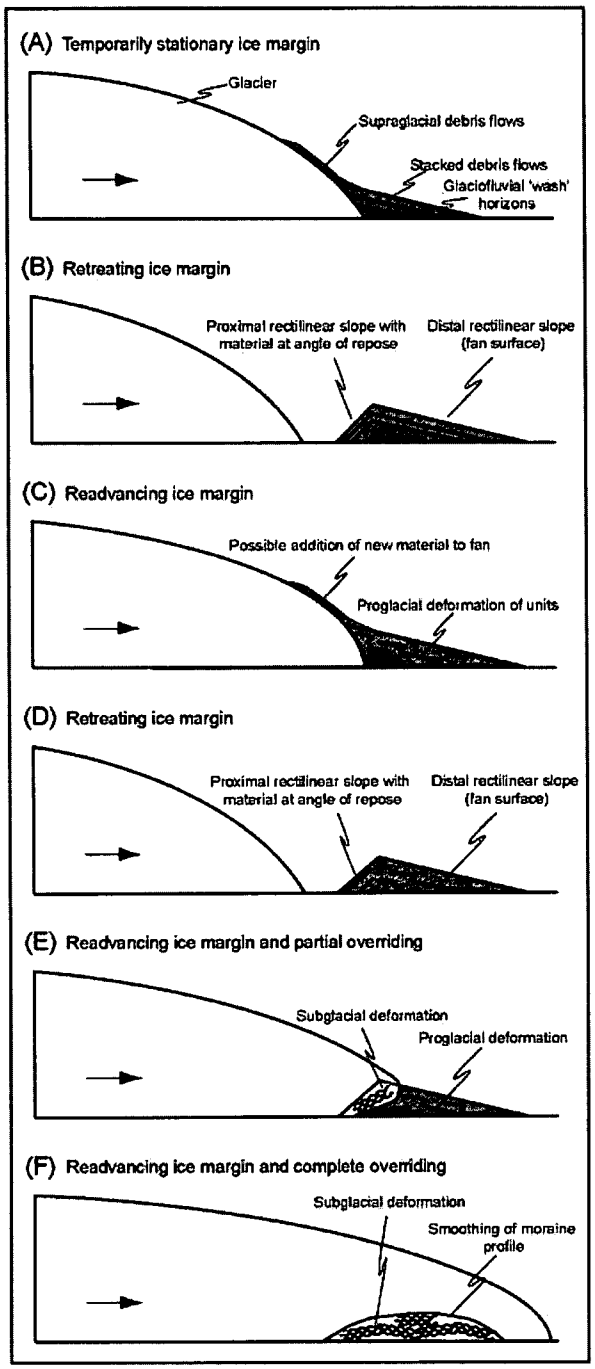


Fig. 12. Schematic representation of sequence of events involved in the formation of terrestrial ice-contact fans as described in this article (cf. Table 1). A. Fan formation along a temporarily stationary ice margin by stacking of supraglacial debris flows and glaciofluvial sediments. B. Formation of the rectilinear ice-contact face where material is at the angle of repose as a result of partial collapse following withdrawal of ice support. C. Short-lived readvance of the ice margin (with a possible annual signature in some areas) causing widespread deformation within the fan and occasionally the addition of new material. D. Formation of a new ice-contact face and abandonment of the fan. E. Partial overriding of the proximal part of a moraine leading to partial glaciotectonization. F. Larger-scale overriding leading to smoothing and alteration of the original moraine asymmetry and complete glaciotectonization.

invoking englacial thrusting leaving a large number of debris septa stacked on top of each other after deglaciation is complete. As shown above (Fig. 4), most moraines characteristically have rectilinear slopes with an angle in the same classes as used by Bennett *et al.* (1998), Graham & Midgley (2000) and Graham (2002) to invoke Svalbard glaciers as modern analogues. However, sedimentological evidence unequivocally demonstrates that these proximal rectilinear slopes correspond to the former ice-contact faces of small terrestrial ice-contact fans where material is at the angle of repose. Likewise, individual hummocks do not simply display 'single facies or facies associations', as proposed by Hambrey *et al.* (1997: p. 624) and Bennett *et al.* (1998: p. 19), but complex sedimentary structures in a variety of sedimentary facies and facies associations when good exposure conditions exist or are created.

Additional problems of their approach are inherent in the interpretation of the thrusts as such (e.g. Evans & Rea 1999; Woodward *et al.* 2002, 2003) and the relatively little evidence that is used to support their model in Britain.

To conclude, transfer of the thrusting model to Scottish Younger Dryas glacier systems is based on a number of assumptions that do not fit the overall picture of the available evidence in large parts of the Scottish Highlands and Islands. There is no sedimentological proof for englacial thrusting being responsible for any of the moraines summarized under 'hummocky moraine'.

Conclusions

The vast majority of 'hummocky' ice-marginal moraines formed during the Younger Dryas in the NW Scottish Highlands represent terrestrial ice-contact fans. These consist of stacked supraglacial debris flow deposits and glaciofluvial sediments and usually form at a temporarily stationary ice margin. During continued retreat, a rectilinear ice-contact face at angles of repose around 30° is formed by collapse of the proximal part of the fan. This steeper slope contrasts with a much gentler distal slope – the fan surface – creating a marked morphological asymmetry (Fig. 12B, D).

The evidence presented in this article demonstrates that such terrestrial ice-contact fans (a) can form in succession, (b) reflect individual palaeo-ice-marginal positions that can be reconstructed by linking moraine crestlines, and (c) explain the marked moraine ridge

asymmetry that is frequently observed in the Scottish Highlands.

Deformation structures in the vast majority of fans demonstrate that glaciers readvanced into these fans after formation. Varying degrees of proglacial and/or subglacial deformation appear to form a continuum of deformation structures, dependent on the duration and intensity of the glacier readvance (Fig. 12). This evidence is only compatible with highly active, temperate glaciers that readvanced into previously formed ice-contact fans in an oscillatory retreat fashion and compatible with evidence found in other parts of Scotland. The geomorphological and sedimentological evidence indicates that glaciers had short response times, being closely coupled with climate; it has not been resolved yet whether individual suites of these moraines were formed on an annual basis.

The most important implication of this work is that englacial thrusting is not a mechanism that can explain the formation of Scottish ice-marginal 'hummocky' moraines as it is not supported by sedimentological evidence. Rectilinear slopes, used as a definite criterion for the surface continuation of former thrust planes, instead represent former ice-contact faces where material is at the angle of repose.

The claim that 'British Loch Lomond Stadial landforms [have] a closer affinity with Arctic glaciers than those in the Alps or Iceland' (Hambrey *et al.* 2001: p. 25) thus cannot be maintained. Rather than an arctic analogue (e.g. Hambrey *et al.* 1997, 2001), this study demonstrates that a Norwegian or Icelandic analogue is much closer to the Scottish Younger Dryas setting in terms of the geomorphology and sedimentology, but also latitudinal proximity. The findings imply that the temperature depressions in the upland areas were not as severe as implied by the thrusting model. This study highlights the need for a more geographically diverse framework of studies into a Younger Dryas Britain that might have been climatically and glaciologically more complex than hitherto assumed.

Acknowledgements. – This article has greatly benefited from comments and discussions with my principal supervisor, Dr. Doug Benn, for which I would like to thank him. I am grateful for helpful and stimulating discussions with Professor Colin Ballantyne, Dr. Tom Bradwell, Dr. Jeremy Everest, Mr. Nick Golledge and Professor Ole Humlum which significantly brought this work forward. I thank my wife Tina for invaluable help with digging and company in the field. Likewise, the assistance of Benedict Reinardy in substantially enlarging and creating several exposures is much appreciated. The work was undertaken as part of the British Geological Survey Moine Thrust Mapping Programme while I was in receipt of a UCAC studentship (2K01E22). I thank Dr. Kenneth F. Rijsdijk and Professor Johannes Krüger for detailed reviews that helped to improve the clarity of the article.

References

Aber, J. S., Croot, D. G. & Fenton, M. M. 1989: *Glaciotectonic Structures and Landforms*. 200 pp. Kluwer, Amsterdam.

Ballantyne, C. K. 1989: The Loch Lomond Readvance on the Isle of Skye, Scotland: glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science 4*, 95–108.

Ballantyne, C. K. 2002: The Loch Lomond Readvance on the Isle of Mull, Scotland: glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science 17*, 759–771.

Benn, D. I. 1990: *Scottish Lateglacial Moraines: Debris Supply, Genesis and Significance*. Ph.D. dissertation, University of St Andrews, 458 pp.

Benn, D. I. 1992: The genesis and significance of 'hummocky moraine': evidence from the Isle of Skye, Scotland. *Quaternary Science Reviews 11*, 781–799.

Benn, D. I. & Clapperton, C. M. 2000: Pleistocene glacitectonic landforms and sediments around central Magellan Strait, southernmost Chile: evidence for fast outlet glaciers with cold-based margins. *Quaternary Science Reviews 19*, 591–612.

Benn, D. I. & Evans, D. J. A. 1996: The interpretation and classification of subglacially deformed materials. *Quaternary Science Reviews 15*, 23–52.

Benn, D. I. & Evans, D. J. A. 1998: *Glaciers and Glaciation*. 734 pp. Arnold, London.

Benn, D. I. & Owen, L. A. 2002: Himalayan glacial sedimentary environments: a framework for reconstructing and dating the former extent of glaciers in high mountains. *Quaternary International 97–98*, 3–25.

Benn, D. I., Lowe, J. J. & Walker, M. J. C. 1992: Glacier response to climatic change during the Loch Lomond Stadial and early Flandrian: geomorphological and palynological evidence from the Isle of Skye, Scotland. *Journal of Quaternary Science 7*, 125–144.

Bennett, M. R. 1995: The morphology of glacially fluted terrain: examples from the Northwest Highlands of Scotland. *Proceedings of the Geologists' Association 106*, 27–38.

Bennett, M. R. 2001: The morphology, structural evolution and significance of push moraines. *Earth-Science Reviews 53*, 197–236.

Bennett, M. R. & Boulton, G. S. 1993: A reinterpretation of Scottish 'hummocky moraine' and its significance for the deglaciation of the Scottish Highlands during the Younger Dryas or Loch Lomond Stadial. *Geological Magazine 130*, 301–318.

Bennett, M. R. & Glasser, N. F. 1991: The glacial landforms in Glen Geusachan, Cairngorms: a reinterpretation. *Scottish Geographical Magazine 107*, 116–123.

Bennett, M. R., Hambrey, M. J., Huddart, D. & Glasser, N. F. 1998: Glacial thrusting and moraine-mound formation in Svalbard and Britain: the example of Coire a' Cheud-chnoic (Valley of hundred hills), Torridon, Scotland. *Quaternary Proceedings 6*, 17–34.

Bickerton, R. W. & Matthews, J. A. 1993: 'Little Ice Age' variations of outlet glaciers from the Jostedalsbreen ice-cap, southern Norway: a regional lichenometric-dating study of ice-marginal moraine sequences and their climatic significance. *Journal of Quaternary Science 8*, 45–66.

Boulton, G. S. 1986: Push-moraines and glacier-contact fans in marine and terrestrial environments. *Sedimentology 33*, 677–698.

Bradwell, T. 2005: Bedrock mega-grooves in Assynt, NW Scotland. *Geomorphology 65*, 195–204.

Brooks, S. J. & Birks, H. J. B. 2000: Chironomid-inferred late-glacial air temperatures at Whitrig Bog, south-east Scotland. *Journal of Quaternary Science 15*, 759–764.

Evans, D. J. A. & Benn, D. I. (eds.) 2004: *A Practical Guide to the Study of Glacial Sediments*. 266 pp. Arnold, London.

Evans, D. J. A. & Rea, B. R. 1999: Geomorphology and sedimentology of surging glaciers: a land-systems approach. *Annals of Glaciology 28*, 75–82.

Evans, D. J. A. & Twigg, D. R. 2002: The active temperate glacial landsystem: a model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews 21*, 2143–2177.

Eybergen, F. A. 1987: Glacier snout dynamics and contemporary push moraine formation at the Turtmann-glacier, Wallis,

Switzerland. *In* van der Meer, J. J. M. (ed.): *Tills and Glaciotectonics*, 217–231. Balkema, Rotterdam.

Eyles, N. 1983: Modern Icelandic glaciers as depositional models for 'hummocky moraine' in the Scottish Highlands. *In* Evenson, E. B., Schlüchter, C. & Rabassa, J. (eds.): *Tills and Related Deposits*, 47–59. Balkema, Rotterdam.

Eyles, N., Eyles, C. H. & Miall, A. D. 1983: Lithofacies types and vertical profile models. An alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences. *Sedimentology 30*, 393–410.

Eyles, N., Boyce, J. I. & Barendregt, R. W. 1999: Hummocky moraine: sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology 123*, 163–174.

Golledge, N. R. 2002: Glaci-tectonic deformation of proglacial lake sediments in the Cairngorm Mountains. *Scottish Journal of Geology 38*, 127–136.

Graham, D. J. 2002: *Moraine Mound Formation During the Younger Dryas in Britain and the Neoglacial in Svalbard*. Ph.D. dissertation, University of Wales, 268 pp.

Graham, D. J. & Midgley, N. G. 2000: Moraine-mound formation by englacial thrusting: the Younger Dryas moraines of Cwm Idwal, North Wales. *In* Maltman, A. J., Hubbard, B. & Hambrey, M. J. (eds.): Deformation of glacial materials. *Geological Society of London Special Publication 176*, 321–336.

Gripp, K. 1975: Hochsander – Satzmoräne – Endmoränenvertreter. *Zeitschrift für Geomorphologie 19*, 490–496.

Hambrey, M. J., Huddart, D., Bennett, M. R. & Glasser, N. F. 1997: Genesis of 'hummocky moraines' by thrusting of glacier ice: evidence from Svalbard and Britain. *Journal of the Geological Society of London 154*, 623–632.

Hambrey, M. J., Bennett, M. R., Glasser, N. F. & Huddart, D. 2001: High-arctic glaciers as analogues for glacial landform development in Britain. *Geology Today 17*, 24–28.

Hart, J. K. & Boulton, G. S. 1991: The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. *Quaternary Science Reviews 10*, 335–350.

Hodgson, D. M. 1986: A study of fluted moraines in the Torridon area, NW Scotland. *Journal of Quaternary Science 1*, 109–118.

Hoppe, G. 1952: Hummocky moraine regions, with special reference to the interior of Norrbotten. *Geografiska Annaler 34*, 1–72.

Humlum, O. 1985: Genesis of an imbricate push moraine, Höfdabrekkujökull, Iceland. *Journal of Geology 93*, 185–195.

Johnstone, G. S. & Mykura, W. 1989: *British Regional Geology – The Northern Highlands of Scotland*. 219 pp. HMSO, London.

Kjær, K. H. & Krüger, J. 2001: The final phase of dead-ice development: processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology 48*, 935–952.

Krüger, J. 1994: Glacial processes, sediments, landforms, and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica 21*, 1–233.

Krüger, J. 1996: Moraine ridges formed from subglacial frozen-on sediment slabs and their differentiation from push moraines. *Boreas 25*, 57–63.

Krüger, J. 1997: Development of minor outwash fans at Kötlujökull, Iceland. *Quaternary Science Reviews 16*, 649–659.

Krzyszkowski, D. 2002: Sedimentary successions in ice-marginal fans of the Late Saalian glaciation, southwestern Poland. *Sedimentary Geology 149*, 93–109.

Krzyszkowski, D. & Zielinski, T. 2002: The Pleistocene end moraine fans: controls on their sedimentation and location. *Sedimentary Geology 149*, 73–92.

Lawson, D. E. 1981: Distinguishing characteristics of diamictons at the margin of the Matanuska glacier, Alaska. *Annals of Glaciology 2*, 78–84.

Lawson, D. E. 1982: Mobilization, movement and deposition of active subaerial sediment flows, Matanuska Glacier, Alaska. *Journal of Geology 90*, 279–300.

Lawson, D. E. 1988: Glacigenic resedimentation: classification concepts and application to mass-movement processes and deposits. *In* Goldthwait, R. P. & Matsch, C. L. (eds.): *Genetic Classification of Glacigenic Deposits*, 147–169. Balkema, Rotterdam.

Liedtke, H. 1981: Die nordischen Vereisungen in Mitteleuropa. 2. Auflage. *Forschungen zur deutschen Landeskunde 204*, 1–307.

Lukas, S. 2003: Scottish Landform Example No. 31. The moraines around the Pass of Drumochter. *Scottish Geographical Journal 119*, 383–393.

Lukas, S. 2004: The pattern of deglaciation around the Pass of Drumochter. *In* Lukas, S., Merritt, J. W. & Mitchell, W. A. (eds.): *The Quaternary of the Central Grampian Highlands: Field Guide*, 68–77. Quaternary Research Association, London.

Lukas, S. & Merritt, J. W. 2004: Evidence for a former ice-dammed lake in Coire Mhic-sith. *In* Lukas, S., Merritt, J. W. & Mitchell, W. A. (eds.): *The Quaternary of the Central Grampian Highlands: Field Guide*, 149–158. Quaternary Research Association, London.

Matthews, J. A., McCarroll, D. & Shakesby, R. A. 1995: Contemporary terminal-moraine ridge formation at a temperate glacier: Styggedalsbreen, Jotunheimen, southern Norway. *Boreas 24*, 129–139.

McCarroll, D. & Rijsdijk, K. F. 2003: Deformation styles as a key for interpreting glacial depositional environments. *Journal of Quaternary Science 18*, 473–489.

McDougall, D. A. 2001: The geomorphological impact of Loch Lomond (Younger Dryas) Stadial plateau icefields in the central Lake District, northwest England. *Journal of Quaternary Science 16*, 531–544.

Midgley, N. G. 2001: *Moraine-Mound Development in Britain and Svalbard – The Development of 'Hummocky Moraine'*. Ph.D. dissertation, Liverpool John Moores University, 351 pp.

Mitchell, W. A. & Lukas, S. 2004: Drumochter Mast. *In* Lukas, S., Merritt, J. W. & Mitchell, W. A. (eds.): *The Quaternary of the Central Grampian Highlands: Field Guide*, 163–170. Quaternary Research Association, London.

Nesje, A. & Dahl, S. O. 2000: *Glaciers and Environmental Change*. 203 pp. Arnold, London.

Penck, A. & Brückner, E. 1901/1909: *Die Alpen im Eiszeitalter*. 1157 pp. Tauchnitz, Leipzig.

Rijsdijk, K. F. 2001: Density-drive deformation structures in glacigenic consolidated diamicts: examples from Traeth y Mwnt, Cardiganshire, Wales U.K. *Journal of Sedimentary Research 71*, 122–135.

Sharp, M. 1984: Annual moraine ridges at Skálafellsjökull, southeast Iceland. *Journal of Glaciology 30*, 82–93.

Sissons, J. B. 1967: *The Evolution of Scotland's Scenery*. 259 pp. Oliver & Boyd, Edinburgh.

Sissons, J. B. 1974: A lateglacial ice cap in the central Grampians, Scotland. *Transactions of the Institute of British Geographers 62*, 95–114.

Sissons, J. B. 1979: The Loch Lomond Stadial in the British Isles. *Nature 280*, 199–203.

van der Meer, J. J. M., Kjær, K. H. & Krüger, J. 1999: Subglacial water-escape structures and till structures, Slettjökull, Iceland. *Journal of Quaternary Science 14*, 191–205.

van der Wateren, F. M. 1995: Structural geology and sedimentology of push moraines – processes of soft sediment deformation in a glacial environment and the distribution of glaciotectonic styles. *Mededelingen Rijks Geologische Dienst 54*, 1–168.

van der Wateren, F. M. 1999: Structural geology and sedimentology of the Heiligenhafen till section, Northern Germany. *Quaternary Science Reviews 18*, 1625–1639.

van der Wateren, F. M., Kluiving, S. J. & Bartek, L. R. 2000: Kinematic indicators of subglacial shearing. *In* Maltman, A. J., Hubbard, B. & Hambrey, M. J. (eds.): Deformation of glacial materials. *Geological Society of London Special Publication 176*, 259–278.

Wilson, P. 2002: Morphology and significance of some Loch Lomond Stadial moraines in the south-central Lake District, England. *Proceedings of the Geologists' Association 113*, 9–21.

Wilson, S. B. & Evans, D. J. A. 2000: Scottish Landform Example No. 24: Coire a' Cheud-chnoic, the 'hummocky moraine' of Glen Torridon. *Scottish Geographical Journal 116*, 149–158.

Winkler, S. & Nesje, A. 1999: Moraine formation at an advancing temperate glacier: Brigsdalsbreen, western Norway. *Geografiska Annaler 81A*, 17–30.

Winkler, S., Haakensen, N., Nesje, A. & Rye, N. 1997: Glaziale Dynamik in Westnorwegen – Ablauf und Ursachen des aktuellen Gletschervorstoßes am Jostedalsbreen. *Petermanns Geographische Mitteilungen 141*, 43–63.

Woodward, J., Murray, T. & McCaig, A. 2002: Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard. *Journal of Quaternary Science 17*, 201–209.

Woodward, J., Murray, T. & McCaig, A. 2003: Reply: Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard. *Journal of Quaternary Science 18*, 99–100.

Zielinski, T. & van Loon, A. J. 2000: Subaerial terminoglacial fans III: overview of sedimentary characteristics and depositional model. *Geologie en Mijnbouw 79*, 93–107.

FORMATION, MELTOUT PROCESSES AND LANDSCAPE ALTERATION OF HIGH-ARCTIC ICE-CORED MORAINES—EXAMPLES FROM NORDENSKIÖLD LAND, CENTRAL SPITSBERGEN[1]

Sven Lukas,[2] Lindsey I. Nicholson, and Fionna H. Ross
School of Geography and Geosciences, University of St Andrews, Irvine Building, North Street, St Andrews KY16 9AL, Scotland

Ole Humlum
Department of Physical Geography, Institute of Geoscience, University of Oslo, P.O. Box 1047 Blindern, N-0316 Oslo, Norway

Abstract: The debris-covered ice-margins of three largely cold-based glaciers in central Spitsbergen were investigated to reconstruct their formation and degradation. Clast shapes indicate dominant englacial and supraglacial transport with a smaller subglacial component. Emplacement of material is inferred to have been through meltout along flowlines due to the relatively uniform and continuous debris cover along the glacier margins; no evidence of thrusting has been found. Degradation of all three belts is rapid and involves debris flows at unstable places—e.g., the margins of meltwater channels. Resultant exposure of underlying ice initiates or accelerates melting, thereby leading to further debris flows. Hence, once degradation starts, a self-reinforcing cycle that removes material from the glacier commences. Landform preservation potential on millennial time scales in a high-arctic, continuous permafrost environment is thus limited. This work has implications for the interpretation of Pleistocene landform associations that use modern analogues from Svalbard.

INTRODUCTION

Extensive complexes that have been termed "ice-cored moraines" characterize the margins and mark the neoglacial maximum positions of central Spitsbergen glaciers. These "ice-cored moraines" in fact consist of a zone of marginal supraglacial debris between 0.1 and 4 m thick, which retards the melting of underlying glacier ice (cf. Etzelmüller, 2000; Lyså and Lønne, 2001; Sletten et al., 2001; Sørbel et al., 2001). Because these features consist of supraglacial material covering the frontal part of the

[1]The authors would like to thank Dr. Douglas I. Benn for his invaluable input and help in designing and organizing these separate projects. Discussions with him that ultimately improved the quality of the manuscript are greatly appreciated. Dr. Hanne H. Christiansen and the staff at UNIS, Longyearbyen are thanked for local support in Svalbard and numerous enthusiastic discussions. We are grateful to Dr. Arthur S. Dyke and an anonymous reviewer who provided helpful reviews that improved the readability of the manuscript. SL and FHR acknowledge financial support for the fieldwork from the Bill Bishop Memorial Trust, the British Geomorphological Research Group, the Dudley Stamp Memorial Fund, and the School of Geography and Geosciences, University of St Andrews. LIN acknowledges the Royal Scottish Geographical Society for supporting her fieldwork through a Small Grant.

[2]Corresponding author; email: SL33@st-andrews.ac.uk

Polar Geography, 2005, **29**, No. 2, pp. 79-109.

glacier, as opposed to isolated pockets of buried ice that occur within larger bodies of sediment, we refer to them as *debris-covered ice-margins* or *zones* in this paper.

Investigations on debris-covered ice-margins on Spitsbergen have aimed at understanding the geomorphic significance of small, high-arctic valley glaciers (e.g., Etzelmüller, 2000; Etzelmüller et al., 2000), while the few process investigations focus on moraine formation rather than longer-term landscape evolution and largely deal with glaciers that have undergone marginal retreat (e.g., Lyså and Lønne, 2001; Sletten et al., 2001). Such studies have been undertaken on ice-cored moraines in the North American arctic (e.g., Johnson, 1971; McKenzie and Goodwin, 1987; Mattson and Gardner, 1991), and the fundamental processes appear to be well understood in temperate, non-permafrost environments (Østrem, 1959; Krüger and Kjær, 2000; Kjær and Krüger, 2001; Everest and Bradwell, 2003).

We present results from geomorphological and sedimentological investigations carried out during the summers of 2002 and 2003 with the following objectives: (1) to determine the processes leading to the formation of the supraglacial debris cover and geomorphological features developed within it; (2) to improve understanding of sedimentary processes associated with the decay of stagnant glacier ice in a high-arctic environment; and (3) to develop a conceptual model of landform evolution.

STUDY AREA AND GLACIER CHARACTERISTICS

The Svalbard archipelago is underlain by continuous permafrost up to 500 m thick (Landvik et al., 1988; Humlum et al., 2003), and the mean annual air temperature at Longyearbyen airport is –6°C (Hagen et al., 1993). With about 25% glacier cover, the study area of Nordenskiöld Land in central Spitsbergen is one of the least glaciated areas in Svalbard due to its aridity. The margins of three glaciers were investigated (Fig. 1): Larsbreen, Longyearbreen, and an unnamed glacier on the northern side of Nordenskiöldtoppen, hereafter termed Nordenskiöldtoppenbreen. All three glaciers are part of the Longyeardalen catchment. Mean annual precipitation at the equilibrium-line altitudes of these glaciers is between 500 and 700 mm water equivalent (Humlum, 2002).

Radio-echo soundings and glaciological investigations at Longyearbreen and Larsbreen demonstrate that these are largely cold-based and that subglacial topography is V- rather than U-shaped, indicating that subglacial erosion has been minimal (Tonning, 1996; Etzelmüller et al., 2000). In contrast, Nordenskiöldtoppenbreen occupies a broad, flat, cirque-like depression that leads into a plateau (Platåberget, ca. 450m asl). Nordenskiöldtoppenbreen is smaller and at a higher altitude, so it is assumed to also be predominantly cold-based due to the penetration of permafrost underneath thin glacier margins (cf. Björnsson et al., 1996). There is no known surging history for these three glaciers (Liestøl, 1969, 1993; Hagen et al., 1993; Etzelmüller et al., 2000).

Glaciers on Svalbard reached their Little Ice Age (LIA) maximum around 1900 AD (Svendsen and Mangerud, 1997), and the surface of Longyearbreen has lowered by up to 50 m since ca. 1936 (Justad, 1997), with similar figures having been obtained for nearby Rieperbreen (Lyså and Lønne, 2001) and other glaciers in the area (Ziaja, 2001). Signs of frontal retreat cannot be found at any of these three glaciers, which is characteristic of many non-surging central Spitsbergen glaciers (Etzelmüller, 2000;

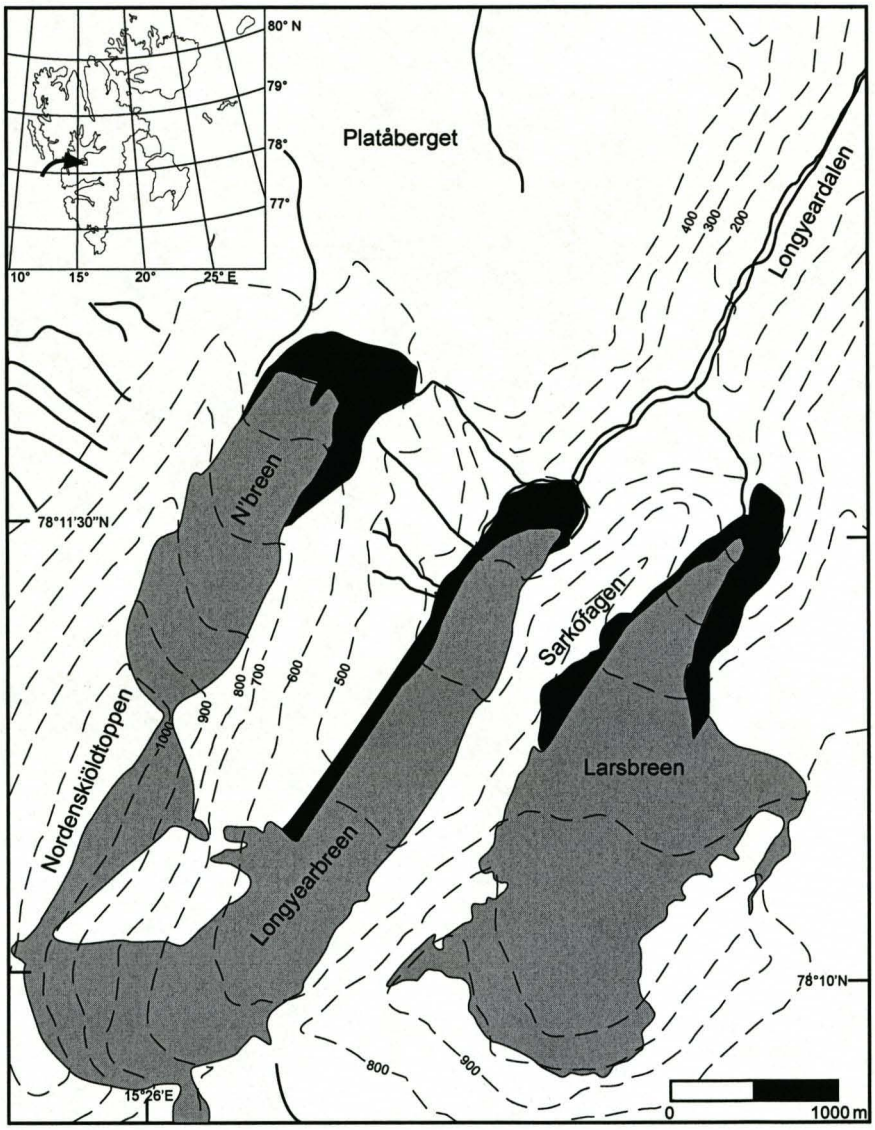


Fig. 1. Overview map of the study area in Central Spitsbergen, showing the three glaciers studied. Black shading indicates the location of the debris-covered ice-margins. Note the frequent occurrence of meltwater channels radiating out from the glaciers.

Etzelmüller et al., 2000; Ziaja, 2001). Hence, the marginal positions of these glaciers are out of equilibrium with climate by >100 years in this part of Svalbard, as they have been stagnating and downwasting since reaching their LIA maximum extent around 1900 AD.

At all three glaciers the LIA maximum is marked by accumulations of supraglacial debris along the presently buried ice front. The width of these debris-covered ice-margins is up to 600 m on Nordenskiöldtoppenbreen, 400 m on Longyearbreen,

TABLE 1
Summary of Characteristics of the Three Glaciers Studied

Glacier	Unit of measure	Nordenskiöld-toppenbreen	Longyearbreen	Larsbreen
Size	km	1.70	3.00	3.20
Area of frontal belt	km^2	0.55	0.55	0.50
Area covered by debris	Percent	32.40	18.30	15.60
Average debris thickness	m	0.38	1.84	1.33
Standard deviation	m	0.34	1.32	0.72
Number of measurements	*N*	32	10	12
Minimum (maximum) thickness of debris	m	0.05 (1.2)	0.2 (4.0)	0.2 (2.5)

and ca. 200m on Larsbreen (Fig. 1), covering up to a third of the glacier surface. Debris thickness varies across individual zones (Table 1). Larsbreen and Longyearbreen share a similar surface form—concave in the accumulation zone and convex at lower altitudes, while Nordenskiöldtoppenbreen has a more linear slope. Larsbreen and Nordenskiöldtoppenbreen flatten out where the debris cover begins.

Tertiary sandstones, siltstones, shales, and localized coal seams of the Palaeocene and Eocene Van Mijenfjorden Group, which dip gently to the WSW, underlie the study area (Hjelle, 1993; Dallmann et al., 2001). One of the most prominent strata is the Grumantbyen Formation, which consists of sandstones and forms prominent cliffs and plateau surfaces such as those surrounding Longyeardalen and Platåberget (Dallmann et al., 2001). Fossils such as bivalves, calcareous foraminifera, worm tracks, and impressions of leaves and other plant remains are common in the Battfjellet Formation, which partly overlooks the slopes in the accumulation areas of all three glaciers (Major et al., 2000; Dallmann et al., 2001), providing ideal tracers for the source area of glacial debris.

METHODS

Geomorphological field mapping and aerial photograph interpretation at a scale of 1:6,000 was used to produce geomorphological maps of the debris-covered margins of the glaciers. Sedimentological logging of sections was carried out on square-millimeter paper and later corrected on overlays of enlarged photomosaics to ensure planimetric accuracy. Sedimentary units were identified on the basis of physical properties including grain size range, compaction, sedimentary structures, and clast shape following guidelines detailed by Evans and Benn (2004) and logged utilizing a modified version of the lithofacies code of Eyles et al. (1983). Clast shape was determined using the method of Benn and Ballantyne (1993, 1994). Fifty clasts were measured at each site and compared to control samples of known environments (glaciofluvial outwash, avalanche deposits, stratified slope deposits, scree). Control samples for subglacially transported clasts could not be obtained, and published subglacial control samples with similar lithological characteristics were used. Particle-size analysis of

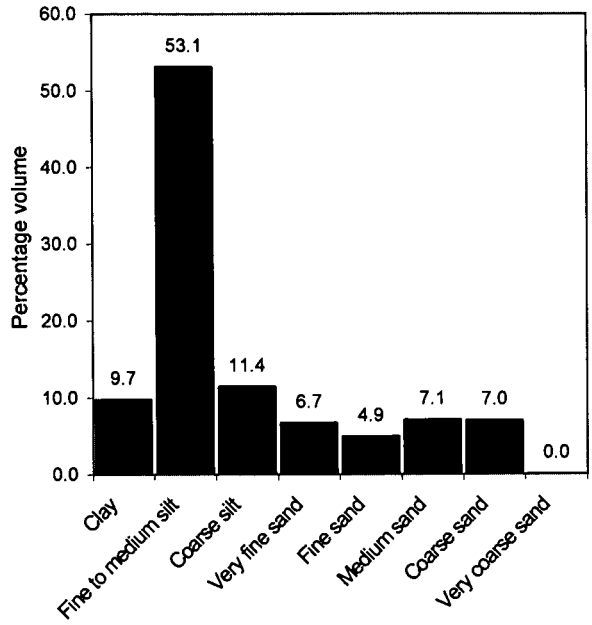


Fig. 2. Particle size fractions of matrix material in the debris-covered ice-margin of Larsbreen.

the matrix material of the Larsbreen debris cover was carried out by Coulter LS230. The retreat rate of four debris flow head-scarps was measured with reference to a stake over a period of 10 days in July 2002. Ablation beneath debris cover was measured on Larsbreen from 9–20 July 2002. Ten plots of ~0.3 m^2 were prepared by clearing the ice of debris, then leveling the ice surface, and drilling in an ablation stake before replacing the debris to the original depth. As far as possible, the debris was replaced in its original stratigraphic position. The debris was left to settle for at least 12 hours prior to measurement. Ablation was measured as the increase in length of the exposed stake, measured above a rule laid across the general ice surface. Debris thickness was measured at the start and end, as settling and minor migration caused the debris thickness to change. Representative debris thickness was taken to be the mean of the two measurements.

DEBRIS-COVERED ICE MARGINS

Physical Characteristics

All three margins are covered by a compact, structureless (massive) clast- to matrix-supported diamicton (Fig. 3A and Table 1). Its matrix is usually silt rich (Fig. 2), and the a-axes of embedded clasts vary from 0.2 to 4 m with all local lithologies present, although larger clasts are predominantly sandstone. On Nordenskiöldtoppenbreen and Larsbreen, blocks of siltstone and shale of up to 7 m in diameter stand out at the surface, surrounded by weathered aprons of scree (Figs. 3D, 4, and 5).

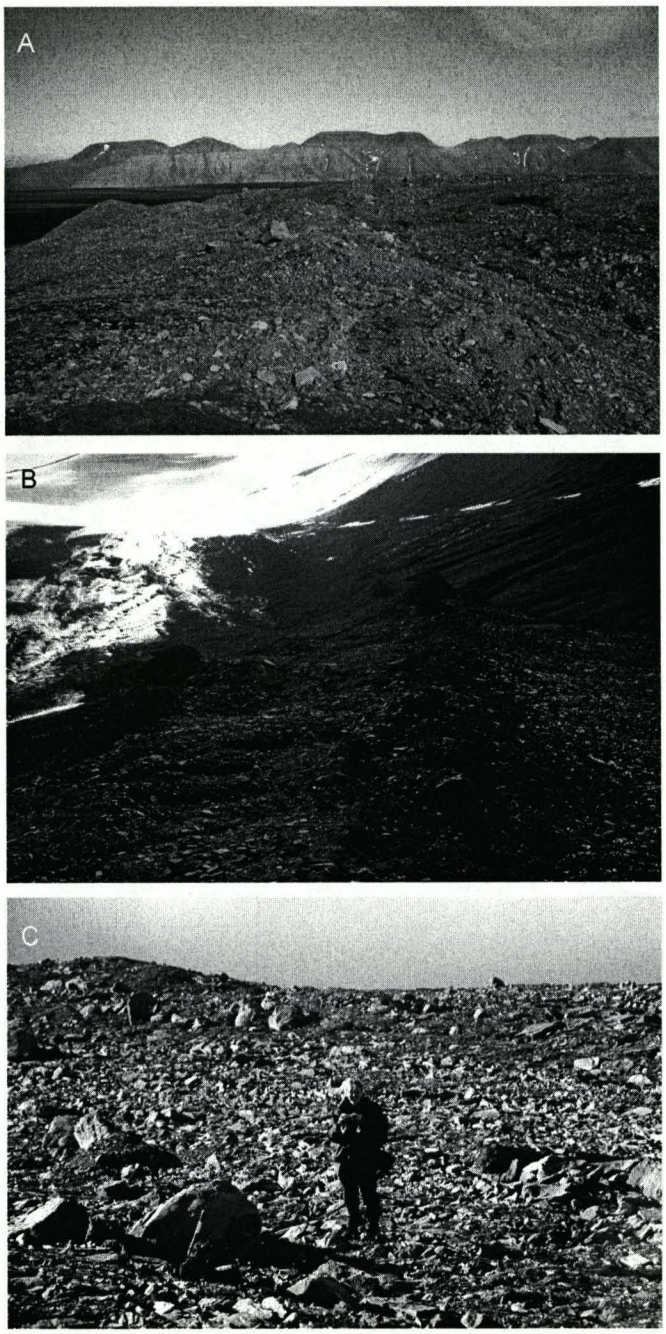


Fig. 3. A. Surficial "ridges" on the surface of the debris-covered zone of Nordenskiöldtoppenbreen (view to the east). B. Western lateral moraine on Nordenskiöldtoppenbreen (view to the south). C. Close-up of subangular, blocky, and partly striated clasts on the surface of Nordenskiöldtoppenbreen (FHR for scale). (*Figure 3 caption continues on following pages*)

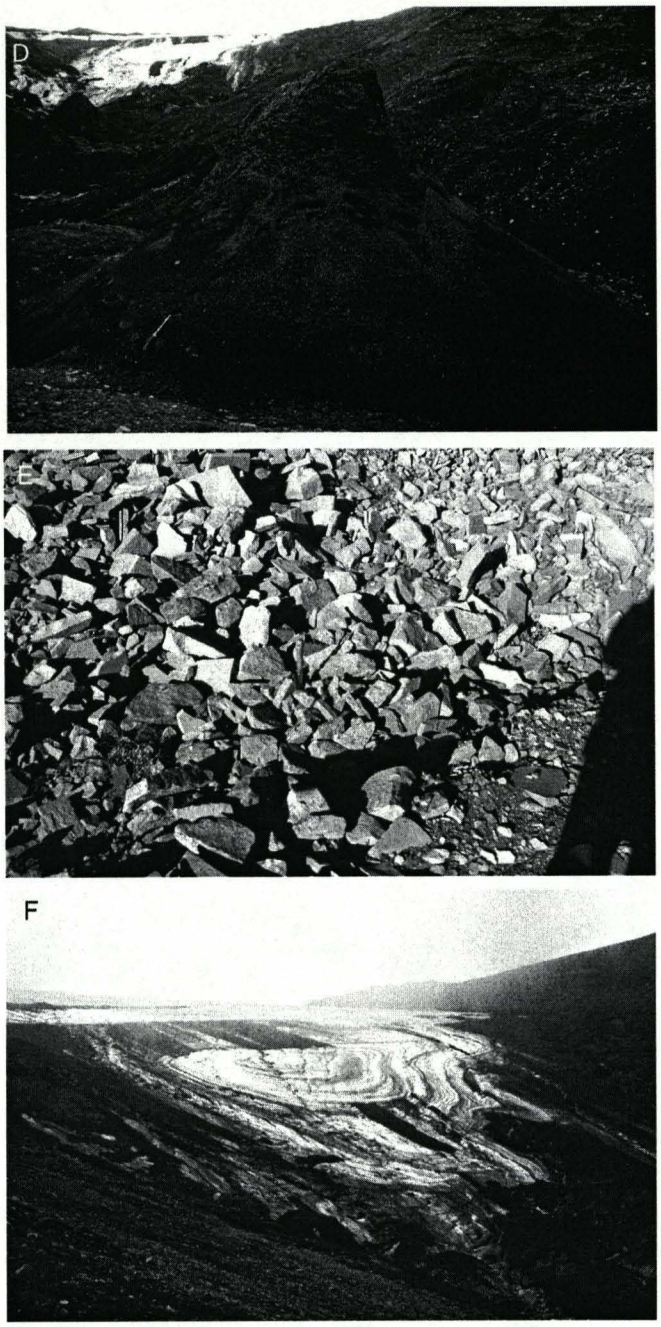


Fig. 3 (continued). D. Detail of a large shale bedrock fragment with surrounding scree slopes on the western side of Larsbreen. Rifle at the bottom left is 1.10 m long. E. Close-up of openwork angular, prolate, and oblate clasts in an area connected to a meltwater channel on Nordenskiöldtoppenbreen. Openwork structure is interpreted as evidence of winnowing of fine material. F. Englacial debris bands cropping out on the surface Larsbreen, suggesting horizontal stratification (view to the southwest). (*Figure 3 caption continues on following pages*).

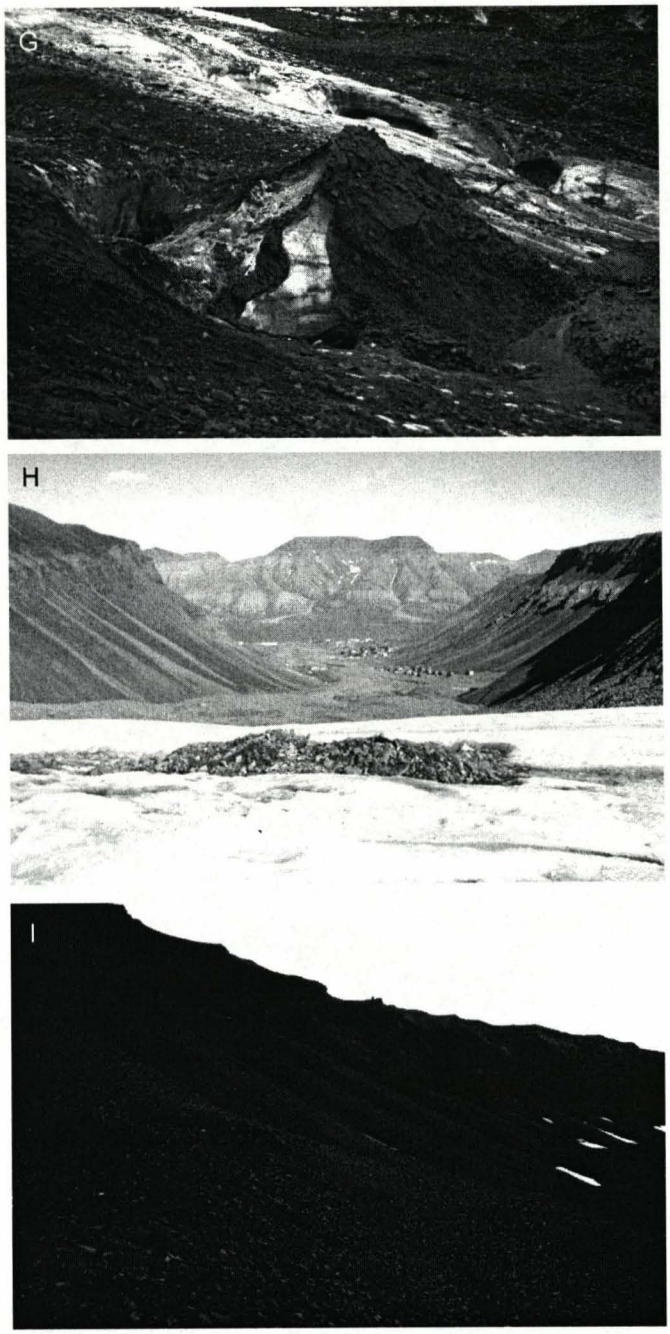


Figure 3 (continued). G. Cross-bedded sand filling a buried crevasse trace on Larsbreen. H. Discrete debris accumulation on the surface of Longyearbreen perpendicular to ice flow. The thickness of this debris accumulation is < 0.2 m (view to the northeast). I. Avalanche debris cones on the eastern side of Larsbreen linking the free faces (left) with the glacier surface (outside right margin). (*Figure 3 caption continues on following pages*).

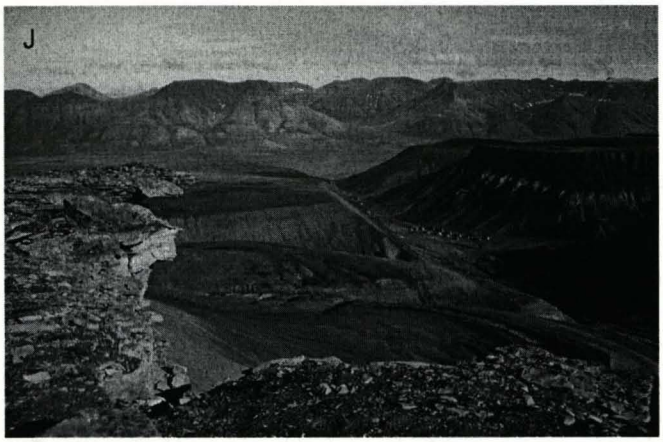


Figure 3 (continued). J. Overhanging rock towers overlooking the surface of Nordenskiöldtoppenbreen in its source area (view to the northeast).

All three glaciers show ice-cored lateral moraines of varying dimensions (e.g., Fig. 3B). Concentric ridges mimicking the buried ice margin can be traced throughout the debris-covered margin of Nordenskiöldtoppenbreen (Figs. 3A and 5). On both sides of Larsbreen, lateral moraines grade into an amorphous, debris-covered zone (cf. Fig. 1). On Longyearbreen, the entire western side is flanked by a prominent lateral moraine (Fig. 1); along the eastern margin, the ice-cored lateral moraine is less well developed. In all cases, test pits have shown that the thickness of debris does not exceed 0.15 m. Moreover, the debris surface closely mimics the surface of the underlying ice, so that "ridges" on the surface are the result of ridges in the glacier ice.

The northeastern margin of Larsbreen is bounded by three ridges, separated by steep-sided gullies (Fig. 4). Clast shape effectively links these ridges (Figs. 6B–6D), which are devoid of fine matrix material, to avalanche debris cones originating from the eastern valley side (Figs. 3I, 6A, and 6F); both consist of very angular to angular, prolate clasts of sandstone and siltstone with very few subangular clasts. These ridges are interpreted as a talus-derived rock glacier that was pushed up in front of the glacier during its neoglacial advance (Humlum, 2005). Avalanche material also overlaps the sides of Larsbreen and Longyearbreen, indicating a potential source area (Figs. 6E and 6G), although some mixing with rounded material is evident (see below). Steep sandstone cliffs (Fig. 3J) overlook the three glaciers in their source areas, and accumulations of openwork, very angular to angular rockfall material are frequently found at the foot of such cliffs on the glacier surface.

Localized sediment accumulations are found on all three glaciers. A discrete debris ridge unequivocally associated with an englacial wedge was found on the surface of Larsbreen (Figs. 3G and 4). This ridge, which is lithologically anomalous to the rest of the debris layer, consists of laminated and cross-bedded sands. Due to the preservation of bedding structures that would undoubtedly be destroyed during prolonged glacial transport, this ridge is interpreted as a crevasse-fill formed from supraglacial stream deposits. Moss fragments contained within the sands gave an uncalibrated ^{14}C age of 150 ± 39 yr (calibrated 1δ age: AD 1682–1947; AAR-7999).

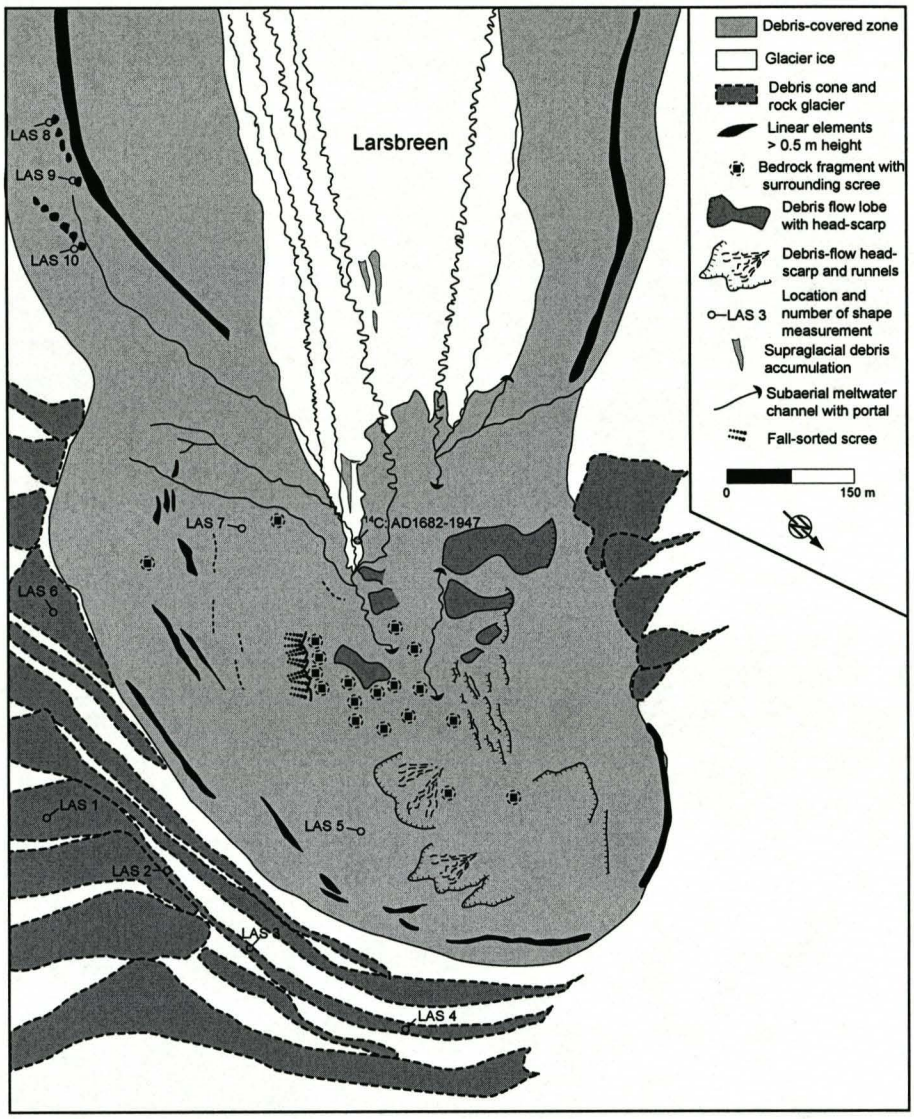


Fig. 4. Geomorphological map of Larsbreen showing the major elements that were identifiable in 2002 and 2003. Compare with Figure 15.

The source area for both sands and moss is inferred to be nivation niches just above the LIA trimline on the slopes north of Trollsteinen ca. 3 km southeast of the sampling site; the ^{14}C age is consistent with a formation during, or since, the Little Ice Age.

Six mounds with heights of up to 1.2 m, widths of up to 4 m, and lengths of up to 5.2 m occur at the eastern side of Larsbreen; they are aligned and regularly spaced. Artificial exposures revealed well-sorted, rounded to well-rounded sand and gravel (Figs. 6J and 6K) beneath a thin (<0.1 m) veneer of angular, supraglacially derived clasts. Towards the margins of these mounds, stratification is disturbed, and the deposits show normal faults, interpreted as collapse features. The sand and gravel

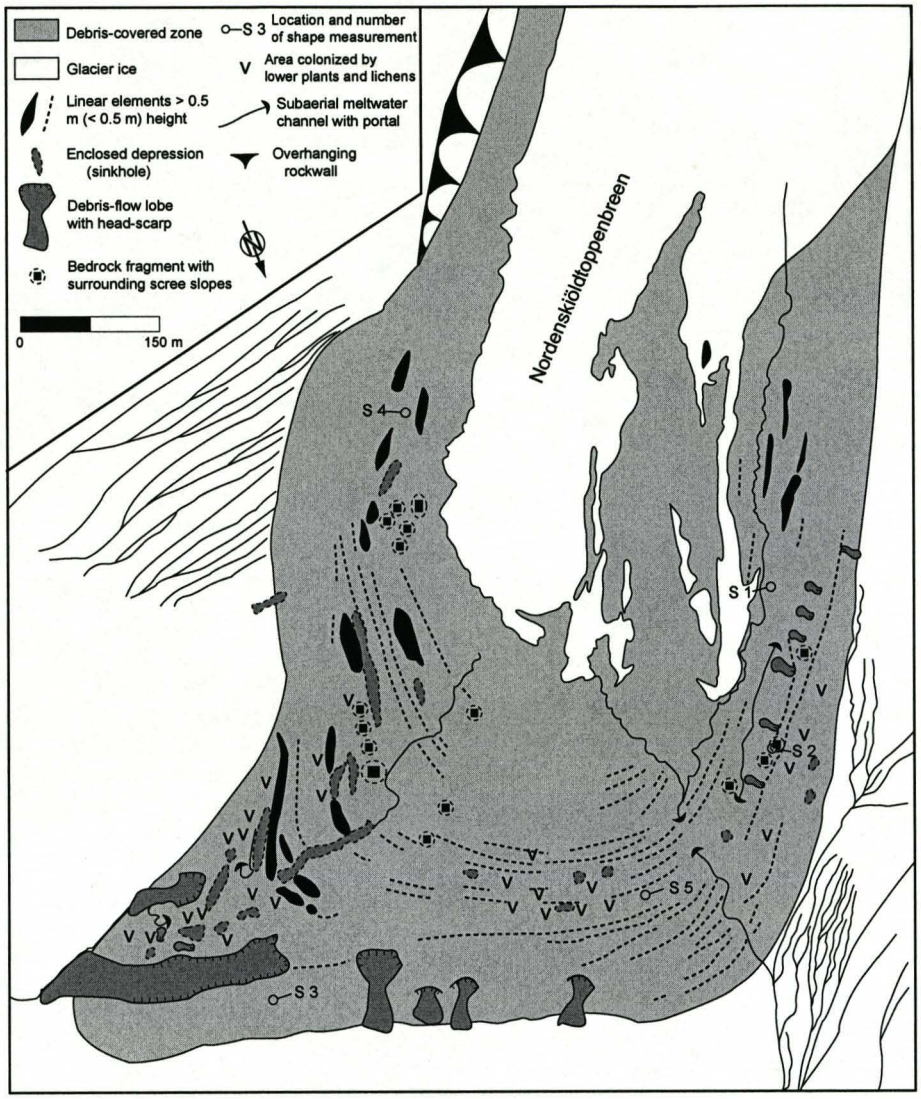


Fig. 5. Geomorphological map of Nordenskiöldtoppenbreen.

rests on diamicton. These isolated, yet aligned deposits are interpreted as pool fills of former supra- or englacial meltwater channels that were topographically inverted during downmelting. Clast shape analysis of neighboring deposits shows, however, that mixing of supraglacial and glaciofluvial material has occurred near the cones (Figs. 6H and 6I).

Upglacier of the main debris-covered zone on Longyearbreen, a ridge of matrix-supported diamicton trends perpendicular to ice flow (Fig. 3H). Similar features can be found in the accumulation area of Nordenskiöldtoppenbreen, where they trend both parallel and perpendicular to ice flow. In all cases the clasts are angular, the

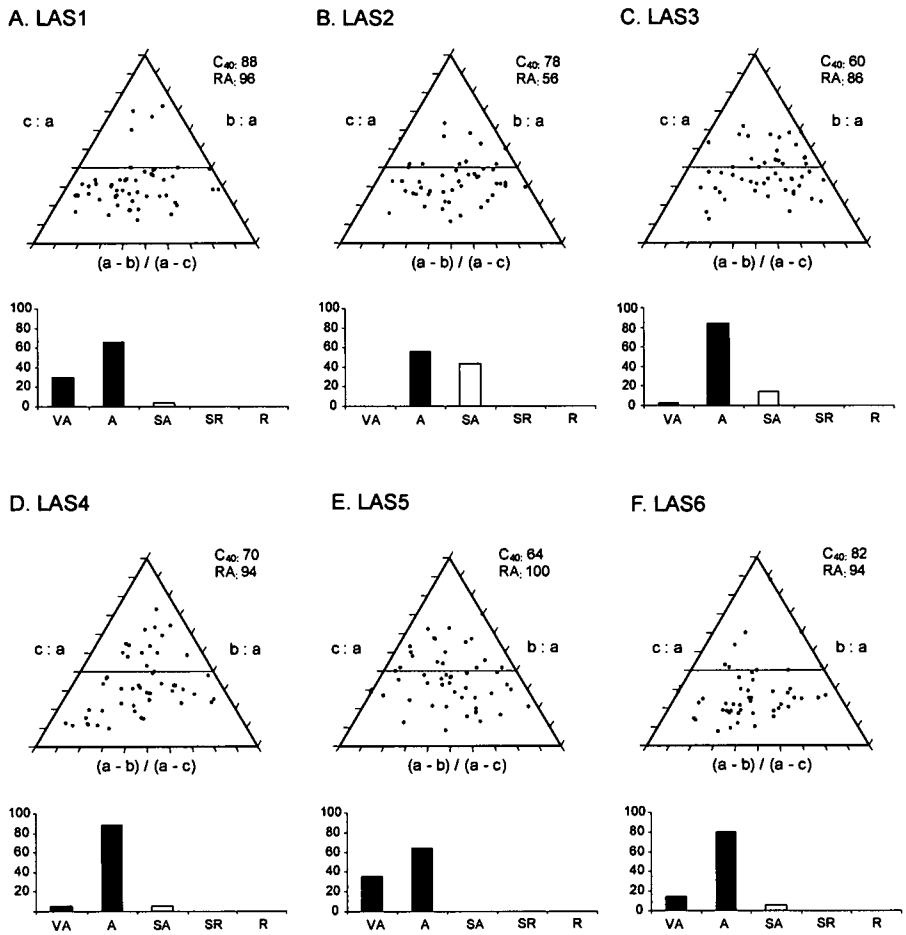


Fig. 6. Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Larsbreen (cf. Fig. 4).

debris cover is < 0.2 m thick, and excavation shows no evident connection to englacial debris bands. This material might represent rockfall onto the ice surface, which was then transferred englacially until it melted out at the surface (see Kirkbride, 1995; Sletten et al., 2001).

CLAST PROVENANCE

Based on the V-shaped subglacial topography of Larsbreen and Longyearbreen and absence of substantial temperate ice patches, Tonning (1996) and Etzelmüller et al. (2000) suggested that basal erosion is minimal. This implies that the primary debris source is supraglacial. Etzelmüller et al. (2000) noted that clasts in the supraglacial debris belts of Larsbreen and Longyearbreen appear to be dominantly angular, which is also suggestive of subaerial debris sourcing. These assumptions are likely to apply at Nordenskiöldtoppenbreen as well, although the overall topography is less

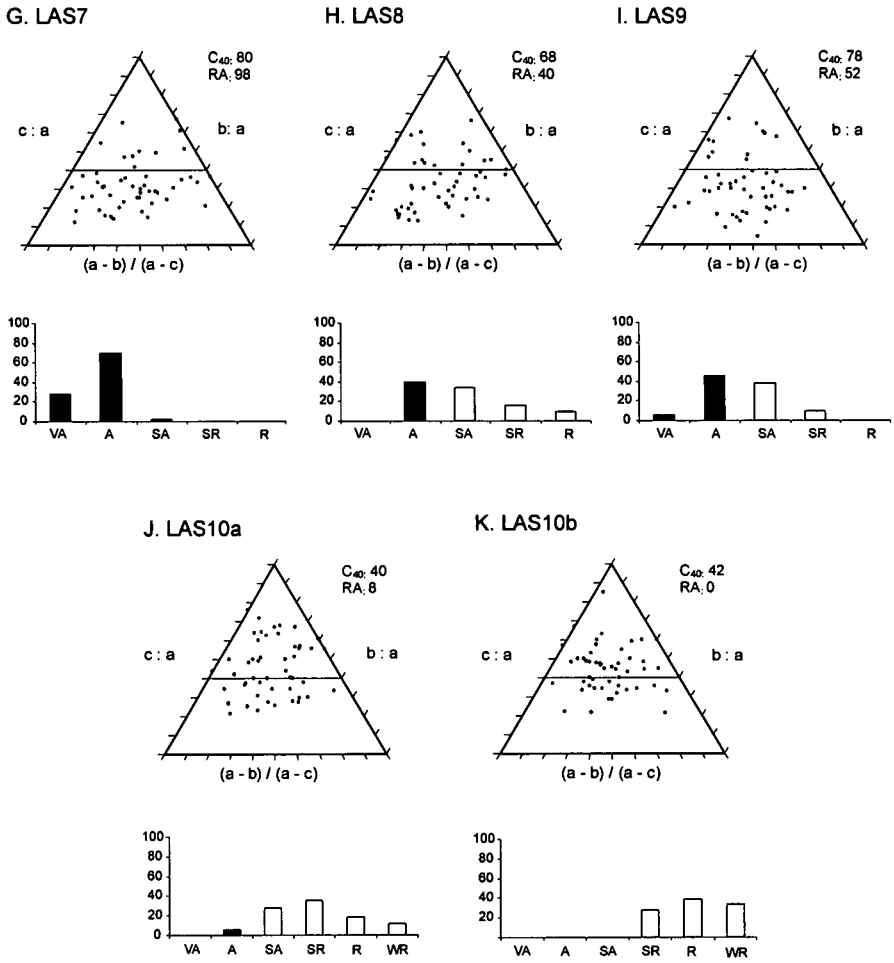


Fig. 6 (continued).

confined than for the former two glaciers. Clast shape analyses were employed to determine the origin and modification of clasts and to identify sedimentary processes.

Some clasts can be traced to source areas due to their containing distinctive fossil assemblages; on Nordenskiöldtoppenbreen, for example, a distinct unit of the Battfjellet Formation crops out in a narrow zone in the glacier source area above the ice, and several blocky and rod-shaped clasts of this unit were found in the debris-covered zone, allowing a reconstruction of the respective transport paths. Embedded clasts of up to 1 m are dominantly rod-shaped and very angular (VA) to angular (A) (Figs. 7A, 7B, and 7D), indicating little alteration during transport and, thus, supraglacial or englacial transport paths (see Boulton, 1978; Benn and Ballantyne, 1993, 1994). However, some larger clasts in the same samples are blocky (having a high c:a ratio) and/or subangular (Figs. 7A–7C). Together with striae on such clasts, this indicates that they have been transported subglacially. Frost shattering of the finer-grained lithologies is evident in the production of very angular prolate and oblate

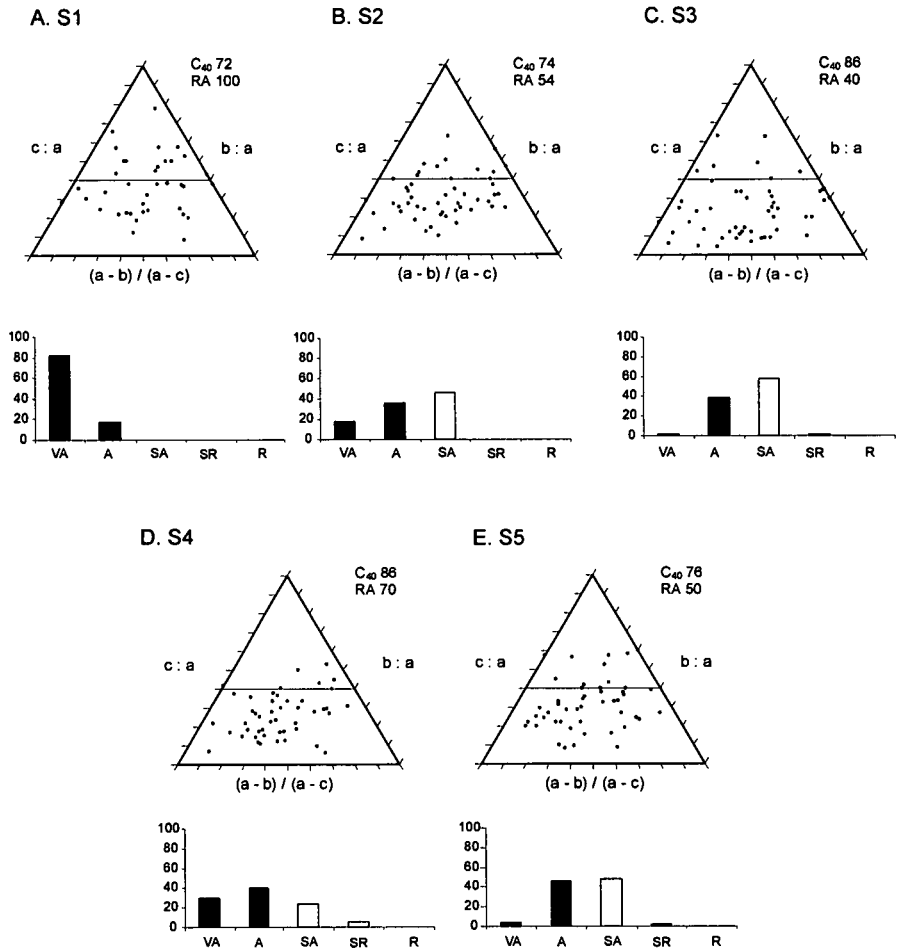


Fig. 7. Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Nordenskiöldtoppenbreen (cf. Fig. 5).

shards, whereas sandstones tend to produce blockier shapes. Coal is rare in the debris-covered zones due to its low strength and, where present, tends to be blocky.

Clasts in all three ice-marginal debris-covered zones (Figs. 6, 7, and 9) have high C_{40} indices, indicating predominantly oblate and prolate clasts with few blocky clasts. Larger proportions of subrounded, or even rounded, clasts only occur under special circumstances as described for Larsbreen above. On Longyearbreen, the admixture of subangular and subrounded material (Figs. 8 and 9C–9D) appears to indicate more efficient rounding during subglacial and/or fluvial transport, which might be linked to this being the longest glacier. Admixture of a small portion of fluvially rounded clasts in sample LS5 (Fig. 9E) might relate to localized fluvial reworking of subglacial material.

The difference between purely subaerially sourced and transported material and the mixed composition of the ice-marginal debris-covered zones can best be seen at

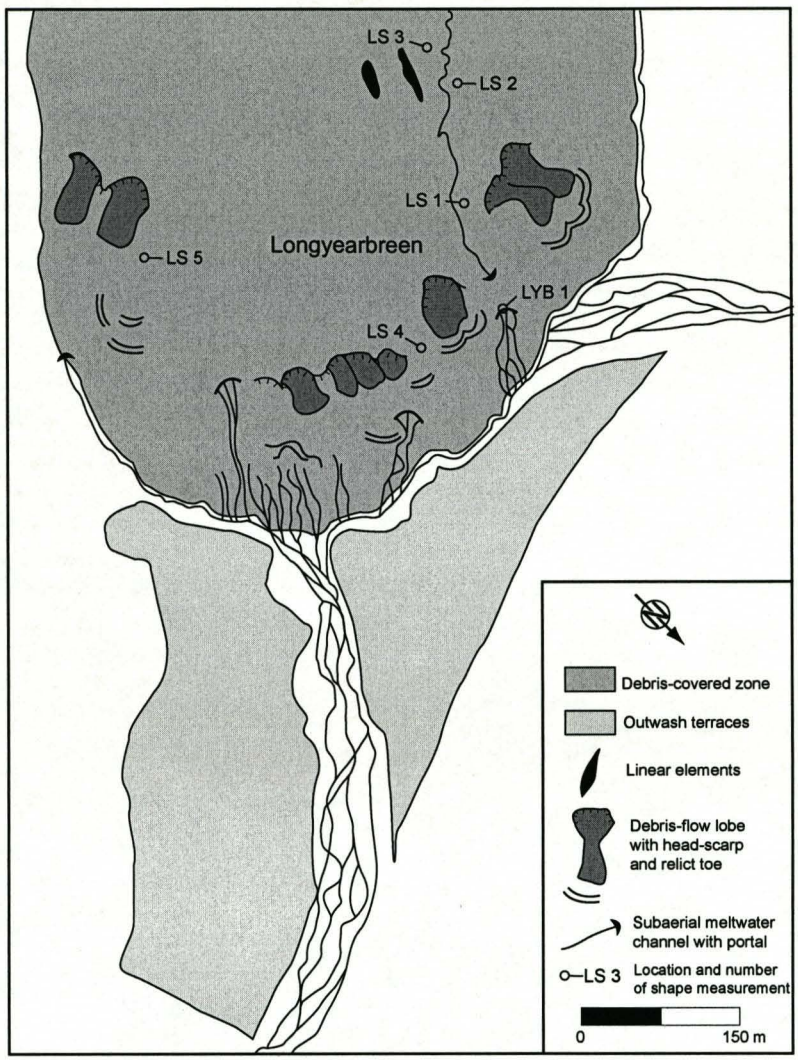


Fig. 8. Geomorphological map of the frontal part of the ice-marginal debris-covered zone of Longyearbreen.

the margin of Larsbreen. Here, control samples taken from avalanche cones (Figs. 6A and 6F) have a high C_{40} index and a negligible subangular component. The latter is attributed to edge-rounding as a result of clast impacting during avalanches. Clast shape largely falls in the continuum between slabs and blocks. This compares well with samples taken from the rock glacier ridges in front of the debris-covered zone (Figs. 6B–6D), although the C_{40} index is somewhat lower, probably due to weathering. A good example of the influence of weathering is given by sample LAS2 (Fig. 6B), which contains an unusually large proportion of subangular clasts. This was found to be due to the presence of voids underneath larger clasts at the surface that had been infilled with smaller clasts that had undergone weathering and rounding. In

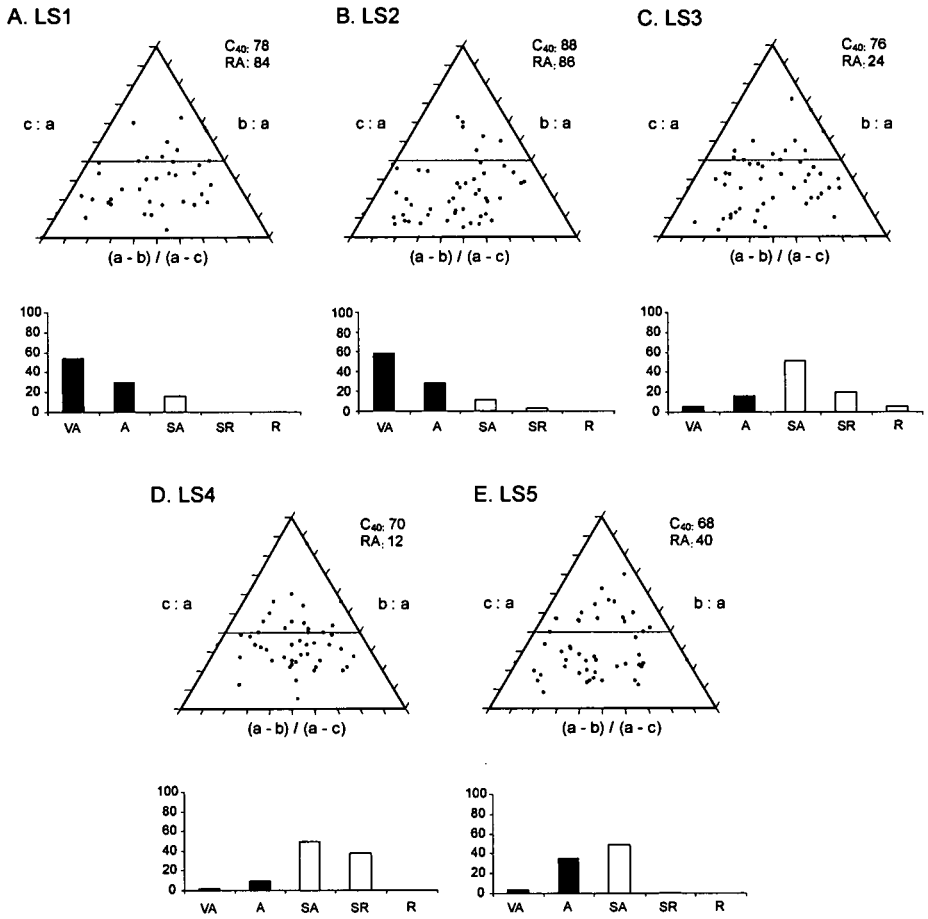


Fig. 9. Shape diagrams (ternary diagrams and clast form frequency plots) for clast shape samples taken at Longyearbreen (cf. Fig. 8).

contrast, samples taken from within the debris-covered zone show a much wider spread of the C_{40} and RA indices (Figs. 6E, 6G–6I), which is attributed to mixing of supraglacial and englacial with subglacial material rather than purely subaerial weathering.

MODE OF FORMATION

Comparison of clast shapes that dominate the debris-covered zones (Figs. 6, 7, and 9) with control samples (Figs. 6A and 6F) indicates supra- or englacial transport. The fact that many clasts on Nordenskiöldtoppenbreen and Longyearbreen can be traced to distinct strata in the headwall indicates that rockfall is the predominant source of material. This material will then be buried by snow and incorporated into the glacier, most likely being transported englacially (e.g., Kirkbride, 1995; Spedding and Evans, 2002). Along the frontal margin of Larsbreen, cliffs are separated from the

glacier surface by debris cones over a horizontal distance of up to 100m. Thus, in this case it is not the immediate effect of larger blocks fallen onto the glacier surface, but the weathered and smaller avalanche material that largely contributes to the formation of the finer-grained debris-covered zone. In all three cases, however, supraglacial and englacial transport of supraglacially derived (rockfall and avalanche) deposits appears to dominate in the present glacial transport systems.

Subglacial transport has played a minor role in the formation of all three debris-covered zones, as evident from the presence of striae and clearly subglacially transported (blocky) clasts. Also, although there is a large component of very angular and angular oblate and prolate clasts indicative of supraglacial and englacial transport in the debris-covered zones, a large matrix component is present. Supraglacial transport does not generate large amounts of fine material due to the absence of crushing and grinding (Boulton, 1976; Benn and Evans, 1998); additional and potentially significant nival (see above) and aeolian dust sources are suspected. The former is evident from the presence of nivation hollows surrounding the glacier and the notion that nivation is an important geomorphic factor in high-arctic areas (e.g., Christiansen, 1998). The latter can be deduced from the surrounding landscape being largely unvegetated and local wind velocity being generally high, with a mean of 3.8 ms^{-1} at Gruvefjellet station (477 m) for the period of 18.08.2001 to 06.09.2004 (Ole Humlum, unpublished data). Hence, subglacial transport, as evident from the presence of striated boulders, contributed to the formation of the debris-covered zones; subaerial weathering, nivation, and aeolian processes alone can probably not account for the large amount of matrix material. It is likely that all three glaciers had larger temperate areas and were more active than today close to reaching their Little Ice Age maximum due to an increased thickness (cf. Sletten et al., 2001; Lyså and Lønne, 2001), explaining the widespread occurrence of subglacially transported clasts.

On Larsbreen and Nordenskiöldtoppenbreen the outcrop pattern of englacial, debris-rich layers on the glacier surface (Fig. 3F) is concentric and characterized by an absence of folds or shears within the ice; we interpret this as evidence of subhorizontal, parallel stratification. Outcrop patterns of isolated debris patches and the continuity of all three debris covers suggests dominant transport along glacier flowlines. The occurrence of a bedded sand wedge on the surface of Larsbreen cannot be used to infer en bloc elevation of older material within a thrust as the material is: (1) likely to have formed at the time of, or since, the LIA glacier advance; and (2) is unlikely to have retained clear and undisturbed bedding structures during thrusting or squeezing. The same complications hold true for isolated sediment bodies on the other two glaciers, where a connection to englacial debris septa could not be found. Such discrete debris lenses could resemble "fossil crevasse fills" that have retained their original coherence during transport and downglacier rotation (Small and Gomez, 1981); transport along flowlines and subsequent meltout appears to be the most reasonable explanation in these cases (cf. Type C Ridges, Sletten et al., 2001). Where relatively thin (≤0.3m), isolated debris patches melt out, cones tend to form, as ice beneath the thickest part of the debris patch is protected from ablation. As the surrounding ice surface melts down, the slope angles of the cone increase until debris becomes unstable, resulting in radial redistribution (cf. Figs. 3G and 3H), explaining the relatively even debris cover parallel to the underlying ice surface along "medial moraines" (see above). Evidence of thrusting, as inferred at the margins of some Svalbard glaciers

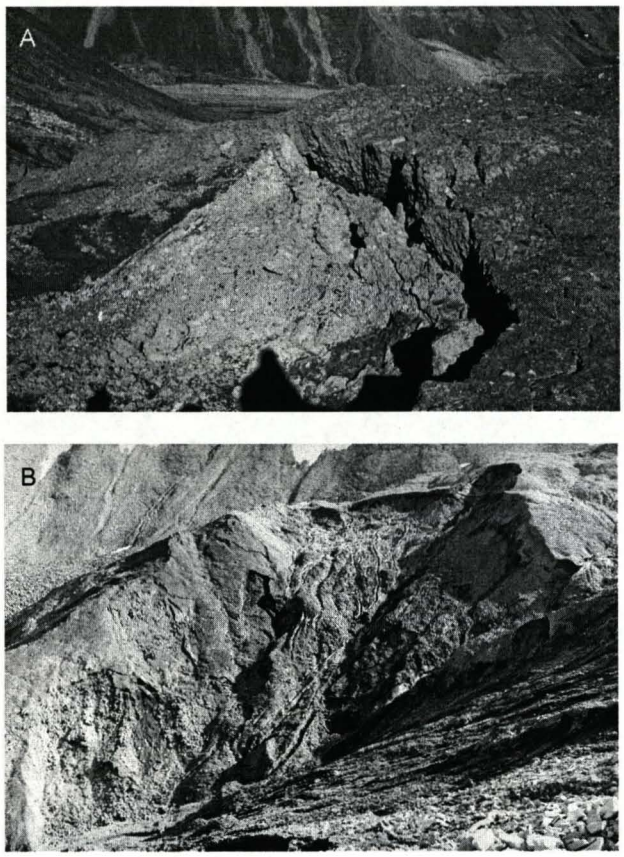


Fig. 10. A. Close-up of debris-flow scarp on eastern side of Larsbreen showing numerous tension cracks and dried material of previous slump. B. Overview of northeastern funnel-shaped debris flow complex showing several inset flow lobes, runnels and skinflows. (*Figure 10 caption continues on facing page*).

(e.g., Hambrey et al., 1997, 1999; Bennett et al., 1998), was not found; this generally agrees with findings from glaciers nearby (Lyså and Lønne, 2001; Sletten et al., 2001).

Melting out of the bedload of former drainage channels (Pelto, 2000; Spedding, 2000), individual rockfalls and crevasse-fills form localized concentrations of debris that are likely to be reworked by meltwater and gravitational sliding to form a more continuous debris cover. Emerging debris bands or crevasse fills containing fine debris would probably become liquefied to form extensive debris covers (cf. "flow tills", Boulton, 1968). In addition, compressional longitudinal flow towards the margins tends to cause medial moraines and surface debris to spread laterally across the glacier terminus (Anderson, 2000).

DEGRADATION OF ICE-MARGINAL ZONES

Wet, cohesive debris flows are ubiquitous on sloping surfaces in the marginal zones (Figs. 4, 5, 8, 10A, 10B, 10E, and 10F). Active debris flows commonly incise

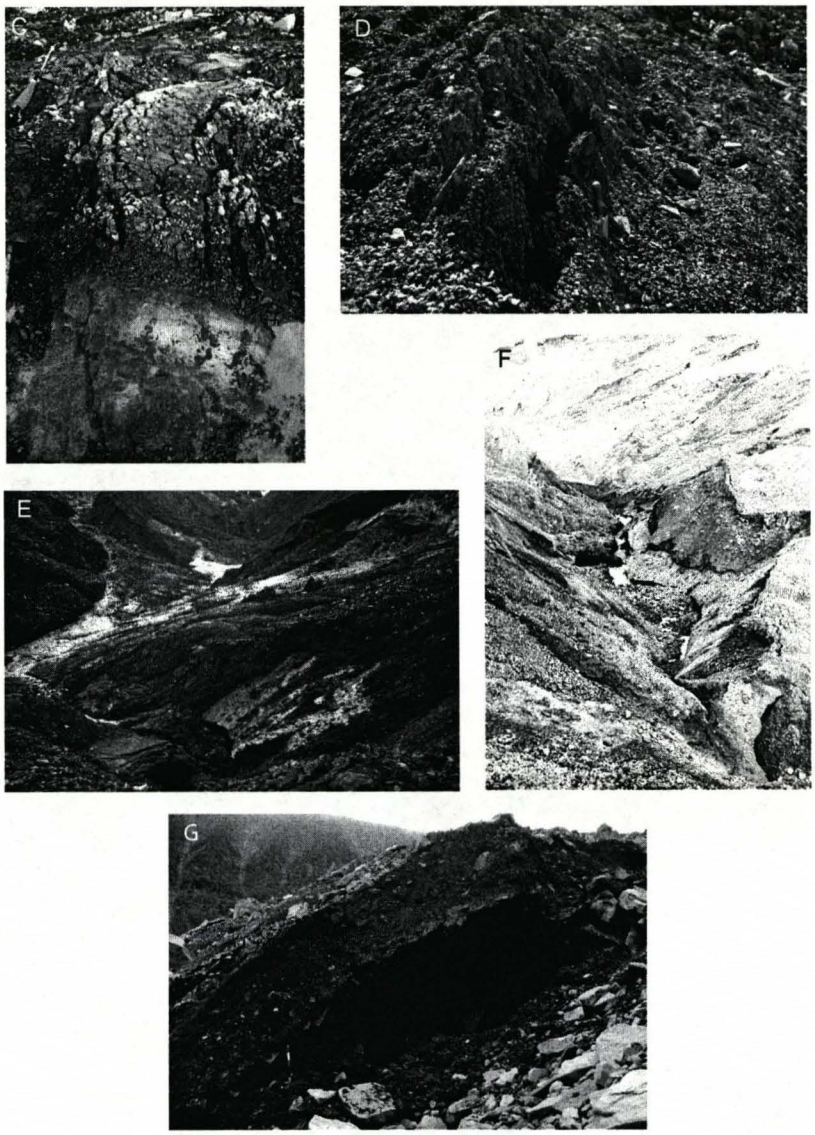


Figure 10 (continued). C. Tension cracks in thin (0.15 m) debris cover on Nordenskiöldtoppenbreen. The ice surface in the foreground has been cleaned to show the conditions at the sediment-ice interface where fine material has been winnowed by topmelting of the ice to form an openwork fine gravel lag on which the debris cover rests. D. Wider (0.2 m) tension cracks on ca. 1.0 m–thick debris cover on western lateral moraine of Nordenskiöldtoppenbreen.E. Exposed ice along the main meltwater channel on Larsbreen. Debris flows along this channel propagated upslope and led to increased melting of the formerly buried ice surface during the observation period in 2003. F. Smaller-scale erosion of a former englacial meltwater channel flanked by debris flows and unstable material upslope in the northeastern part of Nordenskiöldtoppenbreen (view to west). G. Alternating units of clast-supported, stratified diamicton interpreted as stacked debris flows in the steep "frontal wedge" of Longyearbreen. Compare with Figure 13.

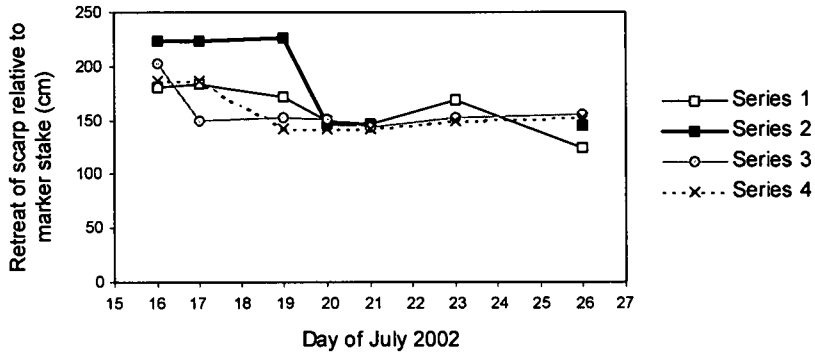


Fig. 11. Scarp retreat rates of four debris flows on the eastern side of Larsbreen during July 2002.

relict flow lobes, in places giving the hillside a staircase-like appearance (Figs. 4, 5, and 8). No relationship between debris-flow activity and slope angle was found, with both active and inactive flows on Larsbreen existing on slopes ranging from 9° to 43°, suggesting that the trigger for debris flows is likely to be a result of variation in moisture conditions rather than a critical slope angle. Mean slope angle of active debris flows was 21°, while that of inactive debris flow was 20°, suggesting that drying out rather than slope angle may be responsible for debris stabilization. Active flow scarps are located around the crestlines of ridges, such as along the lateral moraines (Figs. 4 and 5). The largest debris flows are bounded by head-scarps up to 2 m high (Figs. 10A and 10B) that usually show signs of inherent collapse by opening tension cracks (Figs. 10A and 10C). The rate of retreat of the head-scarps was measured above four active debris flows on Larsbreen in July 2002 (Fig. 11). The mean retreat rate ranged from 3.5–7.8 cm day^{-1}, suggesting that, in summer conditions, up to 0.5 m width of debris around head-scarps of up to 2 m in thickness and several tens of meters in length can be removed in a week. Failure occurs in discrete events with surface cracks expanding behind the scarp in the days preceding the toppling failure of the scarp face (Fig. 10A). Failed material is then liquefied by meltwater and flows downslope, forming flow lobes and runnels (Fig. 10B).

Debris flows are mainly found on unstable slopes. The largest flows are on the distal terminal slopes of Longyearbreen and Nordenskiöldtoppenbreen and flanking the central supraglacial drainage channel on Larsbreen (Figs. 4, 5, and 8). Smaller debris flows form near meltwater channels where the roof of a formerly englacial channel has collapsed (Figs. 10E and 10F). These propagate upslope, creating further instability, and the remobilized material is rapidly evacuated from the debris-covered zone via the fluvial system (Figs. 10E and 10F; cf. Barsch et al., 1994; Alley et al., 1997; Etzelmüller, 2000). Where debris flows occur, they expose the underlying buried glacier ice, allowing rapid melting. The sub-debris ablation rate measured over 10 summer days at Larsbreen showed rates decreasing non-linearly with increasing debris thickness (Fig. 12), as has been found in previous studies (e.g. Østrem, 1959; Loomis, 1970; Nakawo and Young, 1981; Mattson et al., 1993). These measurements suggest that, beneath debris of >0.40 m in thickness, less than 0.005 m of surface lowering occurs per day and that very little melt is likely if debris thickness is >0.60 m.

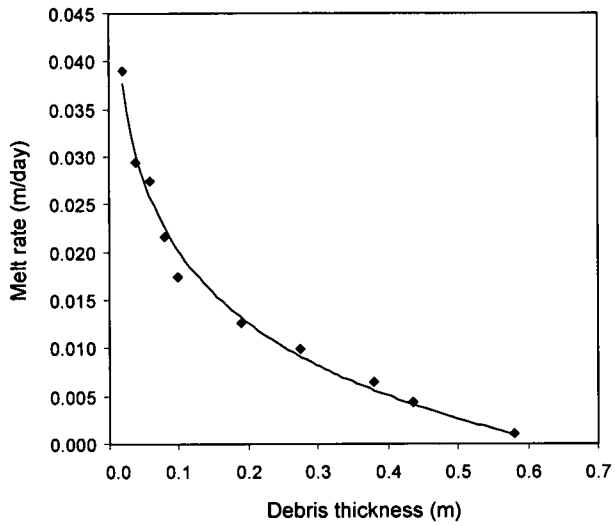


Fig. 12. Relationship between debris thickness and ablation rate measured at Larsbreen from July 9 to 20, 2002, with a logarithmic line of best fit.

Maximum measured melt, of just below 0.04 m day^{-1}, occurred beneath the thinnest debris cover, which was 0.02 m thick. Where the debris cover is completely removed, surface melt increases tenfold, so that areas of exposed ice or very thin debris cover are responsible for a disproportionate amount of ablation in relation to their area (Johnson, 1971; Sakai et al., 2000; Iwata et al., 2000). Experiments have shown that light dustings, or thin surface washes of debris, produce ablation rates in excess of that of clean ice (e.g., Adhikary et al., 2000). Variation in the distribution of debris thickness results in differential ablation beneath the debris cover and the creation of local relief at the ice surface. Meltwater at the ice-sediment interface facilitates translational failure of the whole layer as a cohesive unit, with incipient failures indicated by tension cracks (Figs. 10C and 10D). Removal of material in a narrow zone subparallel to the channel by thin flows results in further instability, which in turn is compensated for by additional debris flows upslope. The effectiveness of the ensuing chain reaction can best be seen at Larsbreen, where comparison of photomosaics suggests that the glacier surface has lowered by up to 5 m around the central former englacial tunnel between August 2002 and August 2003 following its partial roof collapse (see Fig. 15).

Meltwater saturation of debris can also result in large skinflows (Fig. 10B). Increased mobility of thinner debris, which is more likely to be saturated by meltwater, creates a positive feedback for mass movement processes, in which areas of more rapid melt maintain local slope angles, leading to rapid widening of the channel and downslope movement, destabilizing the debris cover upslope and thus perpetuating the erosive undercutting of the surrounding debris cover. Therefore, enhanced melting leads to further debris flows, and the system enters a self-reinforcing cycle.

The observed increase in debris thickness toward the glacier margin can be accounted for by a combination of supra-, en- and subglacial debris concentrations

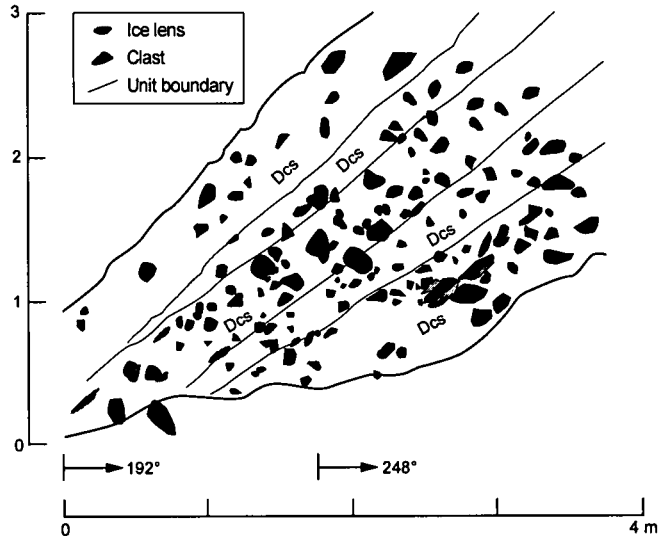


Fig. 13. Sedimentary log of section LYB 1 exposing alternating units of clast-supported, stratified diamicton (Dcs) interpreted as stacked debris flows.

being greatest at the margins and ice flow further concentrating debris toward the terminus (Kirkbride, 2000). Our study shows that, in the case of central Spitsbergen glaciers, these effects are exacerbated by ubiquitous debris flows that shift debris centrifugally thereby constantly increasing the "frontal wedge" by stacking (Fig. 10G). Indeed, this may be the dominant process in this environment. Only one section exposing the sedimentary endproduct of this process was found (LYB1 in Fig. 8). Figure 13 shows compact, crudely stratified, clast-supported diamict units that steeply dip toward the NNE by ca. 20° and are interpreted as stacked debris flows. The matrix throughout the exposure is composed of silty to fine sand with numerous aligned clasts in the fine to coarse gravel fractions and boulders, with a-axes of up to 2 m; the alignment of clasts causes the apparent stratification. At a unit scale, however, stratification is rapidly lost. Numerous ice lenses were observed in cracks parallel to the dip of the units and around larger boulders. Undercutting by an emerging englacial stream led to collapse of a large part of the exposure two days after it had been logged.

Slope steepening and undercutting due to incision of supraglacial meltwater is a frequent process that leads to surface lowering, inducing instabilities and leading to debris flows. The association of the largest flows around the frontal distal slopes of Nordenskiöldtoppenbreen with supraglacial and englacial channels is best explained this way, as indeed some of the largest debris flows appear to have deposited very little sediment at the foot of the flow. Meltwater activity also tends to winnow out fine material from the diamicton, leaving localized areas of coarse, openwork deposits (Fig. 3E).

Subhorizontal marginal areas support a thin cover of lichens, mosses, and small vesicular plants (Figs. 4, 5, and 8). Enters (2000) has described such small plant colonies on Rieperbreen in neighboring Endalen and was able to link the occurrence of vegetation to relatively stable areas that were not disrupted by debris flow activity.

This evidence of marginal stability emphasizes the importance of supraglacial meltwater in destabilising the debris cover.

SYNTHESIS OF DEGRADATIONAL PROCESSES

The observations and data presented above suggest complex links between individual processes. Figure 14 shows a conceptual model that synthesizes the observed evidence into a degradational process–response system. Conditions of the central Spitsbergen glacier system are initially characterized by a steep frontal slope that presents the initial instability along which material is mobilized and transferred away from the glacier front. Collapse of englacial meltwater channels forms steep gradients within a formerly continuous debris cover. Debris flows into these channels result in removal of material and probably enhance fluvial downcutting into the ice. Removal of material in a zone close to the channel induces further debris flows upslope. In all steps, debris flows result in the thinning or complete removal of supraglacial debris, hence enhancing melting. Enhanced melting results in further debris flows. Supraglacial material—once mobilized by debris flows—thus enters a self-maintaining and self-reinforcing cycle of degradation (Fig. 14).

The strong link between the degradation of ice-marginal debris-covered zones and the occurrence of roof collapse triggered by fluvial undercutting is particularly evident on the three glaciers studied. The main drainage routes on Larsbreen and Nordenskiöldtoppenbreen cross the ice-marginal debris-covered zones, as opposed to those of Longyearbreen that run largely parallel to the debris-covered zones in a marginal position (Etzelmüller et al., 2000). Our data, therefore, support the notion that: (1) most surface-debris reworking is due to erosion and destabilization by surface meltwater, rather than gravitational processes; and (2) once fluvial erosion has commenced, rapid degradation of the debris-covered zones ensues (King and Volk, 1994; Etzelmüller, 2000). These observations confirm theoretical reasoning that, where a glacier links sediments and meltwater, most of the sediment will be removed (Alley et al., 1997). This effect can best be seen at Larsbreen, where large-scale debris removal along the central channel initiated enhanced melting and degradation (Figs. 15A and 15B).

PRESERVATION POTENTIAL AND LANDFORM GENESIS

Debris flows redistribute material where the supraglacial debris cover is thinner than the active layer. On some glaciers, mass movements redistribute debris in talus fans to topographic lows (Iwata et al., 1980), within which ablation is then retarded. Debris redistribution can cause topographic inversion, where hollows become highs and vice versa (Clayton, 1964; Hands, 2004). Such inversion may happen several times during a period of glacial retreat and is an important process by which debris is distributed more uniformly across the glacier (Clayton, 1964; Drewry, 1972; Watson, 1980; Anderson, 2000). However, in the case of the glaciers studied here, most of this material is evacuated along meltwater channels. Consequently, it is unclear what features would identify the sedimentary endproduct of this style of marginal deposition.

The observations made here have implications for long-term landscape evolution. The larger proportion of mobilized material is removed from the system to be stored

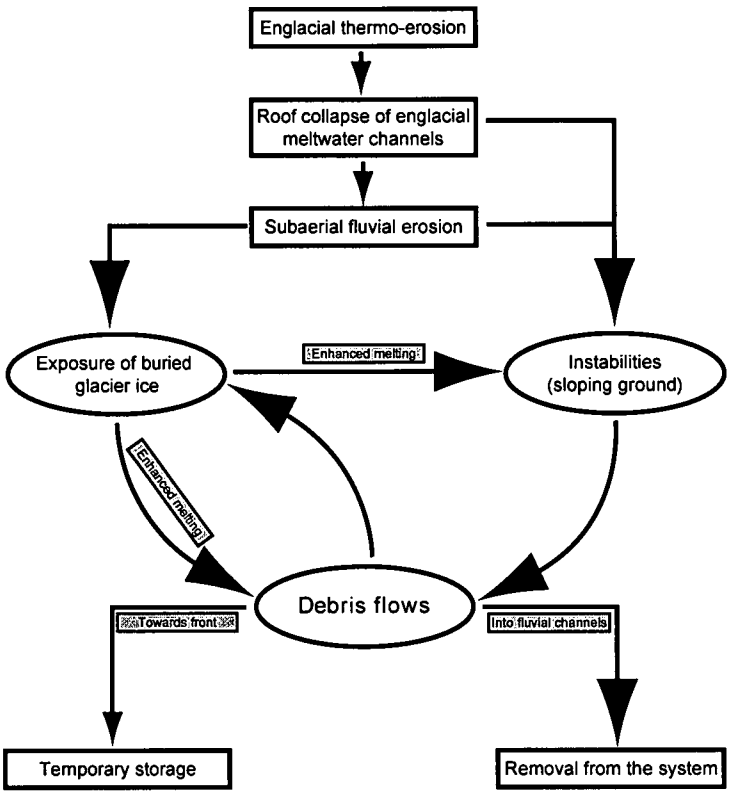


Fig. 14. Conceptual model of the process-response system of the degradation of ice-marginal debris-covered zones of cold-based to polythermal Central Spitsbergen glaciers. Temporarily stored material will be subject to reworking due to gradual degradation of the underlying ice.

in outwash plains or the sea, and only a small proportion might survive in what we have termed the "frontal wedge" (Fig. 14). Even this thickening "frontal wedge" will degrade with time due to solifluction under continuous permafrost conditions and little constructional evidence of a formerly glaciated terrain will be preserved on time scales ranging from centuries to millennia. These observations add to a growing body of evidence from Svalbard that the preservation potential of glacigenic sediments in a continuous permafrost environment is quite limited due to a very efficient coupling of the slopes with the fluvial system via glaciers (e.g., Barsch et al., 1994; Blümel et al., 1994; Etzelmüller, 2000; Etzelmüller et al., 2000; Lyså and Lønne, 2001). This also explains the problems of constructive landform preservation encountered in attempts to reconstruct the history of high-arctic glaciers on Svalbard (e.g., Blümel et al., 1994; Landvik et al., 1998; Lyså and Lønne, 2001; Sletten et al., 2001; Sørbel et al., 2001; Eitel et al., 2002). These findings contrast with clear glacial geomorphological evidence from high-arctic North America where classic glacial landsystems have been preserved over several millennia (e.g., Dyke and Evans, 2003). The steeper topography in Svalbard and the smaller size of the glaciers compared to the large lowland

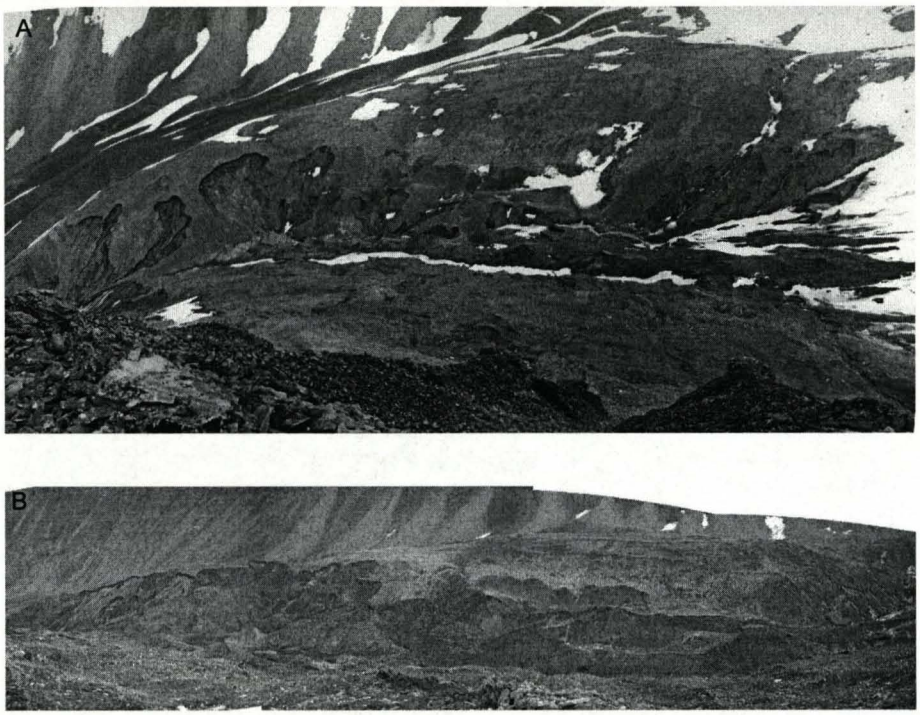


Fig. 15. Panorama picture of the eastern side of Larsbreen. A. August 2002. B. August 2003. Note the enlargement of the failure complex on the left that has grown together in 2003 and the relatively unaltered part farther away from the central meltwater channel on the right.

ice sheet lobes in North America might explain the different response of these two systems to reworking.

In addition, the short-lived nature of the LIA advance in central Spitsbergen might have affected the preservation potential of these moraines. The longer a terminus position is sustained, the more debris can be delivered to it by meltout along englacial flowlines. Consequently a longer LIA could lead to a thicker marginal debris cover. This may inhibit the onset of the decay cycle (Fig. 14), and in some cases this might aid the survival of ice-cored moraines. A thicker debris cover masking the glacier margin could mean that even a decomposed marginal ice-cored moraine may be more likely to leave an observable marginal landform following de-icing. However, two counterarguments can be made. Firstly, a thicker debris layer that accumulated over a longer time span inhibits melt more effectively than a thinner layer. This reduces the amount of ablation that can occur and, assuming constant climatic conditions, would necessitate further glacier advance in order to maintain a mass balance equilibrium. During an advance, the sediment accumulated at the glacier margins would be reworked and redeposited by meltwater and glacial processes. Thus, if an equilibrium situation is re-established, the process of debris accumulation that may lead to ice-cored moraine formation must begin again. Secondly, although a thicker debris cover may inhibit the onset of the positive feedback cycle (Fig. 14), our field observations

suggest that thick debris accumulations are as vulnerable to debris remobilization as thinner ones once the cycle is initiated. This second point also applies to the addition of material by gradual meltout of debris contained in buried ice bodies, which will cause the debris cover to become thicker over time. The preservation potential of remnant features could be increased this way. However, as the decay feedback cycle described in Figure 14 is a positive one, regardless of the thickness of debris, a thicker accumulation of debris on an older feature still does not guarantee its survival, as any meltwater or slope failure event could trigger the onset of decay and the inevitable wasting of the buried ice.

Perhaps the most likely geomorphological signature of small high-arctic valley glaciers is meltwater channels and outwash fans radiating out from areas formerly covered by glaciers (cf. Lyså and Lønne, 2001; Sletten et al., 2001), which also corresponds to findings from high-arctic Canada (e.g., Dyke and Evans, 2003). Primary glacial landforms and sediments such as moraines and till will most likely be modified, obliterated, or completely destroyed. Because of this limited preservation potential of constructional glacial landforms, polythermal, high-arctic glaciers can only serve as modern analogues for areas glaciated during the Pleistocene, where glacial landforms are obliterated and perhaps only the outermost moraine ridge remains clearly identifiable (e.g., Sollid & Sørbel, 1988). Where clear moraines and glacial sediments are preserved and indicate highly active, oscillatory retreat—for example, in the Scottish Highlands—invoking such a high-arctic analogue is unrealistic (Lukas, 2005).

CONCLUSIONS

The debris-covered ice-margins of three cold based to polythermal glaciers were dominantly formed by the incorporation of rockfall and avalanche material sourced from free faces overlooking the glaciers in their source areas; a smaller proportion of material indicates subglacial transport. One implication of these findings is that subglacial temperate areas must have been more extensive when the glacier was thicker during the LIA. Fine-grained matrix material was produced by glacial crushing aided by advection of fine-grained sediment through nivation and aeolian processes. Transfer of material was along flowlines as evident from widespread discrete sediment accumulations on the glaciers interpreted as crevasse fills; evidence for englacial thrusting was not found. Meltout at the front and redistribution of debris led to the emplacement of a continuous debris cover.

Debris flows are the dominant means of material transfer on the three glaciers. These are initiated where instabilities are formed along meltwater channels. Debris flows propagate upslope along such channels, leading to rapid enlargement of a debris-free area. Our results support previous findings that regard glaciers as links between slopes surrounding them and the fluvial system, forming an effective denudation system.

Although stacked debris flows have been observed in one exposure, their preservation potential over many centuries is limited due to glaciofluvial reworking and removal. These results confirm the notion that glacier and ice sheet reconstruction in high arctic Svalbard is extremely difficult due to this effective coupling of the slopes and fluvial system. Our observations permit the conclusion that at best a subdued

outer dump moraine might be the resulting landform after complete de-icing of the landscape. Instead of constructional glacial landforms, meltwater channels and outwash fans are much more likely to indicate a formerly glaciated terrain. Our findings imply that cold-based to polythermal, high-arctic glaciers cannot be used as modern analogues for areas that show clear and well-preserved glacial landforms and sediments such as those in the Scottish Highlands.

LITERATURE

Adhikary, S., M. Nakawo, K. Seko and B. Shakya. "Dust influence on the melting process of glacier ice: experimental results from Lirung Glacier, Nepal Himalayas," in: M. Nakawo, M., C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers*. Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 43-52.

Alley, R. B., K. M. Cuffey, E. B. Evenson, J. C. Strasser, D. E. Lawson, and G. J. Larson. "How glaciers entrain and transport basal sediment: physical constraints," *Quaternary Science Reviews*, Vol. 16, 1997, pp. 1017-1038.

Anderson, R. S. "A model of ablation-dominated medial moraines and the generation of debris-mantled glacier snouts," *Journal of Glaciology*, Vol. 46, 2000, pp. 459-469.

Barsch, D., M. Gude, R. Mäusbacher, G. Schukraft, and A. Schulte. "Recent fluvial sediment budgets in glacial and periglacial environments, NW Spitsbergen," *Zeitschrift für Geomorphologie Supplement Band*, Vol. 97, 1994, pp. 111-122.

Benn, D. I. and C. K. Ballantyne. "The description and representation of particle shape," *Earth Surface Processes and Landforms*, Vol. 18, 1993, pp. 665-672.

Benn, D. I. and C. K. Ballantyne. "Reconstructing the transport history of glacigenic sediments: A new approach based on the co-variance of clast shape indices," *Sedimentary Geology*, Vol. 91, 1994, pp. 215-227.

Benn, D. I. and D. J. A. Evans. *Glaciers and glaciation*. London, UK: Arnold, 1998.

Bennett, M. R., M. J. Hambrey, D. Huddart, and N. F. Glasser. "Glacial thrusting and moraine-mound formation in Svalbard and Britain: The example of Coire a' Cheud-chnoic (Valley of Hundred Hills), Torridon, Scotland," *Quaternary Proceedings*, Vol. 6, 1998, pp. 17-34.

Björnsson, H., Y. Gjessing, S. E. Hamram, J. O. Hagen, O. Liestøl, F. Pålsson, and B. Erlingsson. "The thermal regime of subpolar glaciers mapped by multi-frequency radio-echo soundings," *Journal of Glaciology*, Vol. 12, 1996, pp. 23-31.

Blümel, W. D., J. Eberle, and B. Eitel. "Zur jungquartären Vereisungsgeschichte und Landschaftsentwicklung in NW-Spitzbergen (Liefde-, Bock- und Woodfjord) (On the late Quaternary history of glaciation and landscape evolution in NW Spitsbergen)," *Zeitschrift für Geomorphologie Supplement Band*, Vol. 97, 1994, pp. 31-42.

Boulton, G. S. "Flow tills and related deposits on some Vestspitsbergen glaciers," *Journal of Glaciology*, Vol. 7, 1968, pp. 391-412.

Boulton, G. S. "A genetic classification of tills and criteria for distinguishing tills of different origin," *Geografia*, Vol. 12, 1976, 65-80.

Boulton, G. S. "Boulder shapes and grain-size distributions of debris as indicators of transport paths through a glacier and till genesis," *Sedimentology*, Vol. 25, 1978, 773-799.

Christiansen, H. H. "Nivation forms and processes in unconsolidated sediments, NE Greenland," *Earth Surface Processes and Landforms*, Vol. 23, 1998, 751-760.

Clayton, L. "Karst topography on stagnant glaciers," *Journal of Glaciology*, Vol. 5, 1964, pp. 107-112.

Dallmann, W. K., T. Kjærnet, and A. Nøttvedt. *Geological map of Svalbard 1:100,000—Sheet C9Q Adventdalen. Explanatory text.* Tromsø, Norway: Norsk Polarinstitutt, 2001.

Drewry, D. J. "A quantitative assessment of dirt-cone dynamics," *Journal of Glaciology*, Vol. 11, 1972, pp. 431-446.

Dyke, A. S. and D. J. A. Evans. "Ice-marginal terrestrial landsystems: Northern Laurentide and Innuitian ice sheet margins," in: D. J. A. Evans, ed., *Glacial Landsystems*. London, UK: Arnold, 2003, pp. 143-165.

Eitel, B., K. van der Borg, J. Eberle, and H. Megies. "Late Pleistocene/Early Holocene glacial history of northern Andreeland (northern Spitsbergen/Svalbard Archipelago): Evidence from glacial and fluvio-glacial deposits," *Zeitschrift für Geomorphologie*, Vol. 46, 2002, pp. 337-364.

Enters, D. Primäre Sukzession auf dem Gletschervorfeld des Rieperbreen, Spitzbergen (Primary Plant Succession in the Glacier Foreland of Rieperbreen, Spitsbergen). M.Sc. Thesis, University of Karlsruhe, 2000.

Etzelmüller, B. "Quantification of thermo-erosion in pro-glacial areas—examples from Svalbard," *Zeitschrift für Geomorphologie*, Vol. 44, 2000, pp. 343-361.

Etzelmüller, B., R. S. Ødegard, G. Vatne, R. S. Mysterud, T. Tonning, and J. L. Sollid. "Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen," *Norsk Geografisk Tidsskrift*, Vol. 154, 2000, pp. 157-168.

Evans, D. J. A. and D. I. Benn, eds. *A practical guide to the study of glacial sediments*. London, UK: Arnold, 2004.

Everest, J. D. and T. Bradwell. "Buried glacier ice in southern Iceland and its wider significance," *Geomorphology*, Vol. 52, 2003, pp. 347-358.

Eyles, N., C. H. Eyles, and A. D. Miall. "Lithofacies types and vertical profile models. An alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences," *Sedimentology*, Vol. 30, 1983, pp. 393-410.

Hagen, J. O., O. Liestøl, E. Roland, and T. Jørgensen. "Glacier Atlas of Svalbard and Jan Mayen," *Norsk Polarinstitutt Meddelelser*, Vol. 129, 1993, pp. 1-141.

Hambrey, M. J., D., M. R. Bennett, J. A. Dowdeswell, N. F. Glasser, and D. Huddart. "Debris entrainment and transfer in polythermal valley glaciers," *Journal of Glaciology*, Vol. 45, 1999, pp. 69-86.

Hambrey, M. J., D. Huddart, M. R. Bennett, and N. F. Glasser. "Genesis of 'hummocky moraines' by thrusting of glacier ice: Evidence from Svalbard and Britain," *Journal of the Geological Society of London*, Vol. 154, 1997, pp. 623-632.

Hands, K. A. Downwasting and Supraglacial Lake Eolution on the Debris-Covered Ngozumpa Glacier, Khumbu Himal, Nepal. Ph.D. thesis, University of St. Andrews, 2004.

Hjelle, A. *Geology of Svalbard.* Oslo, Norway: Norsk Polarinstitutt, 1993.

Humlum, O. "Modelling late 20th-century precipitation in Nordenskiöld Land, Svalbard, by geomorphic means," *Norsk Geografisk Tidsskrift*, Vol. 56, 2002, pp. 96-103.

Humlum, O. "Holocene permafrost aggradation in Svalbard," *Geological Society of London Special Publications*, Vol. 242, 2005, pp. 119-130.

Humlum, O., A. Instanes, and J. L. Sollid. "Permafrost in Svalbard: A review of research history, climatic background, and engineering challenges," *Polar Research*, Vol. 22, 2003, pp. 191-215.

Iwata, S., O. Watanabe, and H. Fushimi. "Surface morphology and in the ablation area of the Khumbu Glacier," *Seppyo (Journal of the Japanese Society of Snow and Ice)*, Vol. 41, 1980, pp. 9-17.

Iwata, S., S. Yamaguchi, T. Aoki, T. Kadota, and K. Seko. "Morphological evolution of the debris cover on Khumbu Glacier, Nepal, between 1978 and 1995," in: M. Nakawo, C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers*. Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 3-11.

Johnson, P. G. "Ice-cored moraine formation and degradation, Donjek Glacier, Yukon Territory, Canada," *Geografiska Annaler*, Vol. 53A, 1971, pp. 198-202.

Justad, G. H. Morfologiske og sedimentologiske studier av et iskontakt system ved Longyearbreen, Svalbard (Morphological and Sedimentological Investigations of the Ice-Contact System of Longyearbreen, Svalbard). Cand. Scient. thesis, University of Tromsø, 1997.

King, L. and M. Volk. "Glaziologie und Glazialmorphologie des Liefde- und Bockfjordgebietes, NW-Spitzbergen (Glaciology and glacial geomorphology of the Liefde- and Bockfjord area, NW Spitsbergen)," *Zeitschrift für Geomorphologie Supplement Band*, Vol. 97, 1994, pp. 145-159.

Kirkbride, M. P. "Ice flow vectors on the debris-mantled Tasman glacier, 1957–1986," *Geografiska Annaler*, Vol. 77A, 1995, pp. 147-157.

Kirkbride, M. P. "Ice-marginal geomorphology and Holocene expansion of debris-covered Tasman Glacier, New Zealand," in: M. Nakawo, C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers*. Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 211-218.

Kjær, K. H. and J. Krüger. "The final phase of dead-ice development: Processes and sediment architecture, Kötlujökull, Iceland," *Sedimentology*, Vol. 48, 2001, pp. 935-952.

Krüger, J. and K. H. Kjær. "De-icing procession of ice-cored moraines in a humid, subpolar climate, Kötlujökull, Iceland," *The Holocene*, Vol. 10, 2000, pp. 737-747.

Landvik, J. Y., J. Mangerud, and O. Salvigsen. "Glacial history and permafrost in the Svalbard area," *Permafrost. Fifth International Conference*. Trondheim, Norway: Tapir, 1988, pp. 194-198.

Landvik, J. Y., S. Bondevik, A. Elyerhøei, W. Fjeldskaar, J. Mangerud, O. Salvigsen, M. J. Siegert, J.-I. Svendsen, and T. O. Vorren. "The Last Glacial Maximum of Svalbard and the Barents Sea area: Ice sheet extent and configuration," *Quaternary Science Reviews*, Vol. 17, 1998, pp. 43-75.

Liestøl, O. "Glacier surges in West Spitsbergen," *Canadian Journal of Earth Sciences*, Vol. 6, 1969, pp. 895-897.

Liestøl, O. "Glaciers of Europe—glaciers of Svalbard, Norway," in: R. S. Williams and J. G. Ferriggo, eds., *Satellite image atlas of glaciers of the world.* Reston, VA: USGS Professional Paper, 1993, pp. E127-E151.

Loomis, S. R. "Morphology and ablation processes on glacier ice," *Proceedings of the Association of American Geographers*, Vol. 12, 1970, pp. 88-92.

Lukas, S. "A test of the englacial thrusting hypothesis of 'hummocky' moraines formation: case studies from the northwest Highlands, Scotland," *Boreas*, Vol. 34, 2005 (in press).

Lyså, A. and I. Lønne. "Moraine development at a small High-Arctic valley glacier: Rieperbreen, Svalbard," *Journal of Quaternary Science*, Vol. 16, 2001, pp. 519-529.

Major, H., P. Haremo, W. K. Dallmann, and A. Andresen. *Geological map Svalbard 1:100,000 Spitsbergen—sheet C9Q Adventdalen.* Tromsø, Norway: Norsk Polarinstitutt, 2000.

Mattson, L. E. and J. S. Gardner. "Mass wasting on valley-side ice-cored moraines, Boundary Glacier, Alberta, Canada," *Geografiska Annaler*, Vol. 73A, 1991, pp. 123-128.

Mattson, L. E., J. S. Gardner, and G. J. Young. "Ablation on debris covered glaciers: An example from the Rakhiot Glacier, Punjab, Himalaya," in: G. J. Young, ed., *Snow and glacier hydrology.* Wallingford, UK: IAHS-IASH Publication 218, 1993, pp. 289-296.

McKenzie, G. D. and R. G. Goodwin. "Development of collapsed glacial topography in the Adams Inlet area, Alaska, USA," *Journal of Glaciology*, Vol. 33, 1987, pp. 55-59.

Nakawo, M. and G. J. Young. "Field experiments to determine the effect of a debris layer on ablation of glacier ice," *Annals of Glaciology*, Vol. 2, 1981, pp. 85-91.

Østrem, G. "Ice melting under a thin layer of moraine, and the existence of ice cores in moraine ridges," *Geografiska Annaler*, Vol. 51, 1959, pp. 228-230.

Pelto, M. S. "Mass balance of adjacent debris-covered and clean glacier ice in the North Cascades, Washington," in: M. Nakawo, C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers.* Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 35-42.

Sakai, A., N. Takeuchi, K. Fujita, and M. Nakawo. "Role of supraglacial ponds in the ablation process of a debris-covered glacier in the Nepal Himalayas," in: M. Nakawo, C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers.* Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 119-130.

Sletten, K., A. Lyså, and I. Lønne. "Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard," *Boreas*, Vol. 30, 2001, pp. 272-284.

Small, R. J. and B. Gomez. "The nature and origin of debris layers within Glacier de Tsidjiore Nouve, Valais, Switzerland," *Annals of Glaciology*, Vol. 2, 1981, pp. 109-113.

Sollid, J. L. and L. Sørbel. "Influence of temperature conditions in formation of end moraines in Fennoscandia and Svalbard," *Boreas*, Vol. 17, 1988, pp. 553-558.

Sørbel, L., J. Tolgensbakk, J. O. Hagen, and K. Høgvard. *Geomorphological and Quaternary geological map of Svalbard 1:100,000—sheet C9Q Adventdalen.* Tromsø, Norway: Norsk Polarinstitutt, 2001.

Spedding, N. "Hydrological controls on sediment transport pathways: Implications for debris-covered glaciers," in: M. Nakawo, C. F. Raymond, and A. Fountain, eds., *Debris-Covered Glaciers.* Wallingford, UK: IAHS-AISH Publication 264, 2000, pp. 133-142.

Spedding, N. and D. J. A. Evans. "Sediments and landforms at Kvíárjökull, southeast Iceland: A reappraisal of the glaciated valley landsystem," *Sedimentary Geology,* Vol. 149, 2002, pp. 21-42.

Svendsen, J.-I. and J. Mangerud. "Holocene glacial and climatic variations on Spitsbergen, Svalbard," *The Holocene,* Vol. 7, 1997, pp. 45-57.

Tonning, T. Bruk av eit geografisk informasjonssystem på glasiologiske problemstilingar: døme fra Longyearbreen og Larsbreen på Svalbard (Use of a Geographical Information System for Glaciological Problems: Examples from Longyearbreen and Larsbreen, Svalbard). Cand. Scient. thesis, University of Olso, 1996.

Watson, R. A. "Landform development on moraines of the Klutlan Glacier, Yukon Territory, Canada," *Quaternary Research,* Vol. 14, 1980, pp. 293-300.

Ziaja, W. "Glacial recession in Sørkappland and Central Nordenskiöldland, Spitsbergen, Svalbard, during the 20th century," *Arctic, Antarctic and Alpine Research,* Vol. 33, 2001, pp. 36-41.

Problems of luminescence dating of Late Quaternary glacial sediments in the NW Scottish Highlands

Sven Lukas[a], Joel Q.G. Spencer[*,a,b], Ruth A.J. Robinson[a], Adrian A. Finch[a]

[a] School of Geography and Geosciences, University of St Andrews, St Andrews, Fife KY16 9AL, Scotland, U.K.

[b] Institut für Geologie und Paläontologie, Universität Innsbruck, A-6020 Innsbruck, Austria

* Corresponding author: Tel.: +43 512 507 5683. Fax: +43 512 507 2914.
E-mail address: joel.spencer@uibk.ac.at (J.Q.G. Spencer)

Abstract

Constraining the age of Late Quaternary glaciations is important for palaeoclimatic studies. In the NW Scottish Highlands a large number of closely spaced ('hummocky') moraines formed at glacier margins during overall retreat. Independent age control on one palaeo-ice margin is consistent with the timing of Younger Dryas (YD) glaciation in the area, but further glacier limits have remained undated due to the lack of sites and material for ^{14}C dating. Direct dating of ice-marginal moraines using OSL techniques has never been attempted before in Scotland, but is a useful alternative in such cases. Coarse-grained quartz and K-feldspar from supraglacial sheet flow deposits and glaciolacustrine sediments in ice-marginal moraines were analysed using the SAR protocol. All samples should yield YD or post-LGM ages. Quartz OSL showed significant medium-to-slow components and poor sensitivity to laboratory dose; LM-OSL measurements confirmed that the fast component was weak or absent. Although feldspar-IRSL was very bright with excellent SAR characteristics, aliquots 3 mm in diameter overestimate expected D_e by an order of magnitude. IRSL of single or small clusters (max. ~8) of feldspar grains considerably broaden the apparent D_e distribution but even the lowest value overestimates expected D_e by about 2-3 times, indicating that only a very low concentration of grains can be adequately bleached before

final deposition. CL of the feldspar grains shows equal division between red-brown and blue emission with the latter likely to be responsible for the IRSL signal. Rather than implying mineral or geochemical variability this variation is more likely reflecting different structural states with different luminescence emission. Our study implies that supraglacial sediments, at least from this region in NW Scotland, are very challenging to date with OSL techniques and that the rates of deposition have an important control on their suitability for dating.

Keywords: Optically stimulated luminescence dating; Younger Dryas; Scotland; Poor bleaching; Quartz-luminescence; K-feldspar-luminescence

1. Introduction

Constraining the age of any period of glaciation is the key to understanding glacier-climate interactions during the Quaternary. In the maritime Scottish Highlands, numerous glaciers and ice caps built up during the Younger Dryas (YD) and produced a large number of closely-spaced moraine ridges and mounds, commonly referred to as 'hummocky moraines' (e.g., Sissons, 1967). Elsewhere in Scotland, these landforms have been indirectly dated to the YD on the basis of geomorphological criteria (e.g. Benn, 1992; Benn and Ballantyne, 2004) and different lithostratigraphy observed within kettle holes inside and outside the glacier limits, often supplemented by radiocarbon dating of organic remnants and pollen analysis, but in general the understanding of the timing of the YD is limited to a few isolated sites (Benn et al., 1992; Gordon and Sutherland, 1993). Although luminescence dating studies have been carried out on distal glaciofluvial and glaciolacustrine deposits (e.g., Duller et al., 1995; Duller, 2005; Gemmell et al., 2005), no direct dating of proximal sediments in moraines has ever been attempted in Scotland.

In our study area in the NW Highlands of Scotland (Fig. 1), sites suitable for radiocarbon dating, such as kettle holes, are rare. Initial radiocarbon dates of sedge fragments recovered from an open delta system close to a distinct glacier limit yielded a calibrated 2-sigma age of 10680 to 11230 cal BP (Beta-189451), which is interpreted as marking the end of YD glaciation in the area. Older stratigraphic layers (such as the lateglacial Interstadial) have not been recovered.

Sampling for optically-stimulated luminescence (OSL) dating was conducted to further constrain the extent of YD glaciation. We expected all samples to yield either YD or post-Last Glacial Maximum (LGM) ages. This paper (a) describes the problems we have encountered in our attempts to obtain luminescence dates from these glaciogenic sediments from NW Scotland and (b) discusses implications for the dating of glaciogenic sediments.

2. Study area and sample characteristics

The study area comprises c. 1000 km^2 of the NW Highlands of Scotland in the county of Sutherland (58°5'-58°29'N; 4°58'-4°34'W; Fig. 1). The eastern and larger part of the area is underlain by Moine psammitic and pelitic schists, which contain frequent quartz veins (Johnstone and Mykura, 1989). To the north and west, these rocks have been transported over unmoved Lewisian Gneiss ('basement rocks') to form a series of nappes bounded by thrust planes grouped together in the Moine Thrust Zone (Johnstone and Mykura, 1989). Bedrock is at or near the surface in c. 70-80% of the area and its lithology and structure exert a large influence on the distribution of glacial erosional and depositional landforms (e.g. Bradwell, 2005). All samples described below have been taken from moraines that are formed on and from meta-sedimentary Moine rocks.

Four samples (locations on Fig.1) will be discussed here. The units from which samples BM 2-2 and BM 3-1 were obtained represent thin glaciofluvial wash horizons

produced during supraglacial runoff events near the ice margins. While the sedimentological details of the parent units are discussed elsewhere (Lukas, 2005), it is important to note that these wash horizons form as the saturation capacity of supraglacial diamicton is exceeded causing overland flow during which material is transported in shallow sheet flows and rills (Lawson, 1988; Krüger, 1997; Lukas, 2005). Theoretically, as the sediments were deposited from a supraglacial position, the likelihood of sufficient bleaching is higher due to transport and reworking on the glacier surface (cf. Benn and Owen, 2002). Likewise, sediment transport in shallow water depths prior to deposition has not been observed to represent a barrier to sunlight; the attenuation of the solar spectrum – particularly in the UV range – and shielding effects caused by suspended sediment in turbid flow appear to have a marked effect at greater water depths only (e.g., Wallinga, 2002). The expected age for these samples is between 11.5 and 12.3 ka.

Sample LSh 1 was taken from a similar glaciofluvial wash horizon in an overridden suite of moraines near the expected YD limit and would give a maximum age for the deformation event, probably also between 11.5 and 12.3 ka.

Sample SH 1 is taken from a lacustrine sand unit near the crestline of a moraine ca. 10 km outside of the conspicuous YD limit. This moraine dammed up a small lake in front of the ice before it was breached by meltwater during deglaciation. The expected age for this sample is likely to be between 15-18 ka.

3. Measurements of quartz

Quartz grains were prepared using standard laboratory methods (e.g., Spencer and Owen, 2004) and initially we examined a 125-180 μm grain size fraction. A Risø TL/OSL-DA-15 reader equipped for infrared stimulation (880Δ80 nm) from LEDs and blue-green (420-550 nm) stimulation from a 150 W filtered halogen lamp with liquid light guide and a

UV detection window characterised by 7.5 mm of Hoya U-340 filter and a EMI 9235QA PMT was used for OSL measurements. A single-aliquot regenerative-dose (SAR) procedure (Murray and Wintle, 2000, 2003) was used with post-infrared-blue-green stimulated luminescence (post-IR-BGSL).

Initial signals showed significant medium-to-slow components and a poor fast component in natural, regenerative and test dose data, resulting in scattered dose points and poor SAR characteristics due to poor signal-to-noise levels. Specific luminescence sensitivity was also poor, even though large aliquots (~8-9 mm diameter) were used throughout. Although poor, this initial data indicated very high D_e values ranging from ~200-300 Gy suggesting ages much older than expected (*in-situ* gamma data indicates typical dose-rates to quartz of ~2-3 $mGya^{-1}$; Table 1).

An initial attempt to investigate these problems was to examine a larger 180-212 μm quartz grain size fraction (prepared in parallel with K-rich feldspars – see next section). This was carried out both to test the initial findings but, conversely, also to determine if we could isolate brighter quartz luminescence and whether we could better resolve fast from medium-to-slow components. Additionally, if the initial indication of very high doses is due to poor bleaching, we hoped to see lower doses from the larger 180-212 μm grain size fraction (cf. Olley et al. 1998). However, this fraction yielded similar results (Fig. 2) to the data from 125-180 μm quartz. Linearly modulated-OSL (LM-OSL) measurements (Fig. 3) carried out at the University of Liverpool using a Risø system with a blue diode array (470±30 nm) and UV detection window (as described above) confirmed that the fast component was absent or weak in the natural signal of these samples. In the hope of obtaining useable signals, we changed the focus of our investigations to luminescence of K-rich feldspar, which we describe in the next section.

4 Measurement of 180-212 μm feldspar

K-rich feldspars were separated by floatation in 2.58 gcm^{-3} sodium heteropolytungstate (LST), then etched in 10% HF for 40 minutes and finally treated with 10% HCl. SAR infrared stimulated luminescence (IRSL) measurements were carried out in the Risø system using the infrared LED array described above. Detection was in the blue-violet using Schott BG39 and Corning 7-59 filters. Preheats and cutheats were set at 290°C for 10 s, and IR stimulation was carried out for 300 s at a sample temperature of 50°C (cf. Preusser, 2003). Dose-rate data (~3.0-3.6 $mGya^{-1}$; Table 1) for K-feldspar is based on *in situ* gamma spectrometry assuming internal K of 12.5% (after Huntley and Baril, 1997), although without detailed measurements we prefer to assume a large uncertainty in internal K of ±5%. The following luminescence measurements were carried out:

SAR IRSL of 3 mm diameter aliquots

Feldspar IRSL was very bright with excellent SAR characteristics (Fig. 4) but, similar to the indication from quartz, overestimated the expected D_e by an order of magnitude.

SAR IRSL of single or small clusters of grains

Grains were dispensed onto stainless steel discs using a dissection needle. The number of grains per disc varied from ~1-8 with an average of ~3. About a third of the discs measured had little or no IRSL signal, but those discs with signals had excellent SAR characteristics similar to the 3 mm multi-grain aliquots. Preliminary data for those discs with IRSL signals indicate there can only be a very low concentration of grains with expected D_e values (Fig. 5).

Cathodoluminescence (CL) measurements were also conducted on both the 3 mm aliquots and the single or small clusters of K-feldspar grains after completion of all SAR IRSL measurements. CL of the 3 mm aliquots indicates the grains are approximately equally divided between red-brown & blue emission (Fig. 6), with the latter likely to be responsible

for the IRSL signal. This was confirmed by CL of the single or small clusters of grains; those grains with no IRSL signal typically emit red-brown CL.

5 Discussion

Low specific luminescence sensitivity in quartz from glacially derived sediments has been reported in other studies (e.g., Rhodes and Pownall, 1994; Richards, 2000; Spencer and Owen, 2004) and has been previously attributed to either a young geological source, a lack of subaerial weathering (Rhodes and Pownall, 1994) or insufficient bleaching and dosing cycles in the sedimentary environment. In this work we also show that medium-to-slow components can dominate the signal from quartz from glaciogenic sediments. It may be possible that individual grains are present with a well-resolved fast component compared to medium-to-slow components, but our BGSL and LM-OSL data from a multi-grain approach indicates there are few, if any, such grains. Although highly scattered due to poor signal-to-noise ratio, initial SAR data indicates that ages are overestimated by an order of magnitude. Such a result suggests these sediments were very poorly bleached before deposition, although this is compounded by the fact that the medium-to-slow components that dominate these quartzes are more difficult to bleach than the fast component. Contrastingly, in their recent dating study of glaciogenic sediments from the Buchan area in NE Scotland, Gemmell et al. (2005) have generally observed apparently adequately bleached quartz with reasonable SAR characteristics and reliable age data, although the fast component, albeit present, was not always dominant (Murray, pers. comm., 2005). Studies of single grains of quartz from the Buchan area confirm the presence of a fast component and reasonably bright signals (Duller, pers. comm., 2005).

This difference in characteristic components could be either due to a lack of bleaching and dosing cycles so that the quartz, and more specifically the fast component, is

insufficiently sensitised, or alternatively it may be related to bedrock lithology. While in our study area metamorphic schistose rocks prevail, a greater lithological variety is found in Buchan, including sandstones, acid and basic igneous intrusive bodies and quartzose mica-schists (Stephenson and Gould, 1995). These findings might imply that sediments derived from geologically-old metamorphic lithologies might not produce the characteristics desirable for luminescence dating of quartz.

Although the luminescence from K-rich feldspar grains is very bright with excellent SAR characteristics, similar to the indication from quartz data the 3 mm sized aliquots also overestimate expected ages by about an order of magnitude. However, preliminary SAR-IRSL measurements of single or small clusters of K-feldspars significantly broaden the apparent dose-distribution, but even the lowest value overestimates the expected D_e by ~2-3 times. In line with the quartz data, the feldspar results suggest that the sediments are very poorly bleached, and in addition the single or small cluster K-feldspar data indicates only a very low concentration of grains must be adequately bleached.

Combined quartz and feldspar multi-grain aliquot luminescence studies on glaciogenic sediments from the Hunza and Chitral valleys in northern Pakistan suggested that, in most cases, coarse-grained K-rich feldspars are unsuitable for dating glaciogenic sediments or glaciogenic sediments from those regions (Owen et al., 2002; Spencer and Owen, 2004); the growth of luminescence for most feldspars studied indicated early onset of saturation implying large remnant geological doses – hence, poor bleaching at deposition – and produced large overestimated ages. In contrast, the quartz data was consistent from aliquot-to-aliquot and in general agreement with available independent chronology. The results from this study imply that neither quartz nor feldspar multi-grain aliquots can deliver expected dates, but aside from further complications such as anomalous fading initial data indicate that feldspar single grains show some promise in this glacial setting. It is difficult to assess for

certain whether a quartz single grain approach can give correct ages, but the BGSL and LM-OSL data would indicate this is unlikely.

The CL measurements clearly show the grains within the K-rich density fractions of all samples are approximately equally divided between blue and red-brown emission. The red-brown emission is likely to be a red/IR emission centred in the near IR but with a tail in the red. Although we cannot rule out the possibility these results indicate mineral variability and hence beta dosimetric variability, this is unlikely. In their CL spectroscopy studies of alkali feldspars, Finch and Klein (1999) showed that grains with nominally the same chemistry but different structural states had different luminescence. They assigned the blue emission to the presence of electron holes on bridging oxygens, particularly on the Al-O-AL bridge, and attributed the red/IR emission to the presence of Fe^{3+}. What is clear from our results is that those grains with visible red-brown CL emission either do not contribute to the IRSL signal or contribute very weakly to the IRSL signal. This has interesting implications as to whether such grains should or should not be included in an analysis. Perhaps also CL measurements of quartz grain assemblages would allow us to identify and discriminate not only contaminant grains but also zero dose grains.

6 Implications for OSL-dating of glaciogenic sediments

Our study carries implications for the application of OSL-techniques to the dating of sediments in glacial environments and thus adds to a growing body of data from different geographical settings. As discussed above, quartz from glaciogenic sediments from supraglacial settings has been shown to work well in debris-covered glacier systems such as the Himalayas where debris remains in a supraglacial position for some time (cf. Benn and Owen, 2002; Spencer and Owen, 2004). However, even though the facies sampled for dating here were deposited supraglacially, the constituent minerals are poorly behaved and do not

give the expected ages. The explanation for inadequate bleaching almost certainly lies in the different glacier response times in both systems and the *rates* of supraglacial sedimentation. In NW Scotland, a large number of recessional moraines, indicating a very active Younger Dryas glacial landsystem in which the glaciers retreated in an oscillatory fashion, occur throughout the valley bottoms (Lukas, 2005). The dense spacing and large number of moraines implies that the ice margins oscillated at subdecadal frequencies, producing a new moraine ridge every 3-5 years. Consequently, debris turnover would have been quick and removal from the debris-covered surface too fast to allow sufficient bleaching of grains. Thus, although the sediments chosen here share a common supraglacial origin with those from the Himalayas, the process *rates* are very different, suggesting that supraglacial sediments are not *per se* suitable for OSL dating.

Ultimately, the characterisation of material from different settings in a range of modern ice-marginal environments might help to identify such problems and guide sample selection (cf. Preusser, 1999).

7 Conclusions

Quartz and K-rich feldspar grains from four samples of glaciogenic sediments (supraglacial and proximal glaciolacustrine facies) obtained from ice-marginal moraines in NW Scotland have been measured using the SAR protocol. Independent age control and clear landsystem differences indicate a Younger Dryas or post-LGM age of the landforms we are attempting to date. OSL from large (~8-9 mm) quartz aliquots showed significant medium-to-slow components and poor sensitivity to laboratory dose; LM-OSL measurements confirmed that the fast component was weak or absent. Although feldspar-IRSL was very bright with excellent SAR characteristics, aliquots 3 mm in diameter overestimate expected D_e by an order of magnitude. IRSL of single or small clusters (max. ~8) of feldspar grains considerably

broaden the apparent D_e distribution but even the lowest value overestimates expected D_e by about 2-3 times, indicating only a very low concentration of grains can be adequately bleached before final deposition. CL of the feldspar grains shows equal division between red-brown and blue emission with the latter likely to be responsible for the IRSL signal. Rather than implying mineral or geochemical variability this variation is more likely reflecting different structural states with different luminescence emission.

Although the sediments discussed here were carefully chosen supraglacially-deposited sediments that have been shown to work elsewhere (e.g. in the Himalayas), this study has shown that supraglacial sediments *per* se are not necessarily suited for OSL dating and that sedimentary *rates* and duration of material exposed at the glacier surface are important. In addition to poor bleaching of quartz and feldspar minerals, which we attribute to the depositional environment characteristics, the low specific luminescence and weak or absent fast component in the quartz data makes these samples very challenging. We tentatively attribute the latter problems to either insufficient bleaching-dosing cycles or to the source lithology.

Acknowledgements

We wish to thank Barbara Mauz (University of Liverpool) for conducting the LM-OSL measurements. This study was conducted while SL was in receipt of a UCAC studentship (2K01E22). The financial support of the Russell Trust, University of St Andrews is also gratefully acknowledged.

References

Adamiec, G., Aitken, M., 1998. Dose-rate conversion factors: update. Ancient TL 16, 37-50.

Benn, D.I., 1992. The genesis and significance of 'hummocky moraine': evidence from the Isle of Skye, Scotland. Quaternary Science Reviews 11, 781-799.

Benn, D.I., Lowe, J.J., Walker, M.J.C., 1992. Glacier response to climatic change during the Loch Lomond Stadial and early Flandrian: geomorphological and palynological evidence from the Isle of Skye, Scotland. Journal of Quaternary Science 7, 125-144.

Benn, D.I., Ballantyne, C.K., 2004. Evidence for an independent West Drumochter icefield. In: Lukas, S., Merritt, J.W., Mitchell, W.A. (eds). The Quaternary of the Central Grampian Highlands: Field Guide. Quaternary Research Association, London, 200-209.

Benn, D.I., Owen, L.A., 2002. Himalayan glacial sedimentary environments: a framework for reconstructing and dating the former extent of glaciers in high mountains. Quaternary International 97-98, 3-25.

Bradwell, T., 2005. Bedrock mega-grooves in Assynt, NW Scotland. Geomorphology 65, 195-204.

Duller, G.A.T., Wintle, A.G., Hall, A.M., 1995. Luminescence dating and its application to key pre-Late Devensian sites in Scotland. Quaternary Science Reviews 14, 495-519.

Duller, G.A.T., 2005. Single grain optical dating of glacigenic deposits. LED2005 proceedings.

Finch, A.A., Klein, J., 1999. The causes and petrological significance of cathodoluminescence emissions from alkali feldspars. Contributions to Mineralogy and Petrology 135, 234-243.

Gemmell, A.M.D., Murray, A.S., Connell, E.R., 2005. A revised paradigm for Devensian (Weichselian) glacial events in Buchan (north-east Scotland): luminescence dating of glaciofluvial and glaciolacustrine sediments. Quaternary Geochronology, LED2005 proceedings.

Gordon, J.E., Sutherland, D.G., 1993. The Quaternary of Scotland. Chapman & Hall: London.

Huntley, D.J., Baril, M.R., 1997. The K content of the K-feldspars being measured in optical and thermoluminescence dating. Ancient TL 15, 11-13.

Johnstone, G.S., Mykura, W., 1989. British Regional Geology - The Northern Highlands of Scotland. HMSO, London.

Krüger, J., 1997. Development of minor outwash fans at Kötlujökull, Iceland. Quaternary Science Reviews 16, 649-659.

Lawson, D.E., 1988. Glacigenic resedimentation: Classification concepts and application to mass-movement processes and deposits. In: Goldthwait, R.P., Matsch, C.L. (eds). Genetic classification of glacigenic deposits. Balkema, Rotterdam, 147-169.

Lukas, S., 2005. A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: case studies from the northwest Highlands, Scotland. Boreas 34, 287-307.

Murray, A.S., Wintle, A.G., 2000. Luminescence dating of quartz using an improved single-aliquot regenerative-dose protocol. Radiation Measurements 32, 57-73.

Murray, A.S., Wintle, A.G., 2003. The single-aliquot regenerative-dose protocol: potential for improvements in reliability. Radiation Measurements 37, 377-381.

Olley, J.M. Caitcheon, G.G., Murray, A.S., 1998. The distribution of apparent dose as determined by optically stimulated luminescence in small aliquots of fluvial quartz: implications for dating young sediments. Quaternary Science Reviews 17, 1033-1040.

Owen, L.A., Kamp, U., Spencer, J.Q., Haserodt, K., 2002. Timing and style of Late Quaternary glaciation in the eastern Hindu Kush, Chitral, northern Pakistan: a review and revision of the glacial chronology based on new optically stimulated luminescence dating. Quaternary International 97-98, 41-55.

Prescott, J.R., Hutton, J.T., 1994. Cosmic ray contributions to dose rates for luminescence and ESR dating: large depths and long-term time variations. Radiation Measurements 23, 497-500.

Preusser, F., 1999. Lumineszenzdatierung fluviatiler Sedimente – Beispiele aus der Schweiz und Norddeutschland. Kölner Forum für Geologie und Paläontologie 3, 1-62.

Preusser, F., 2003. IRSL dating of K-rich feldspars using the SAR protocol: comparison with independent age control. Ancient TL 21(1): 17-23.

Rhodes, E.J., Pownall, L., 1994. Zeroing of the OSL signal in quartz from young glaciofluvial sediments. Radiation Measurements 23, 581-585.

Richards, B.W.M., 2000. Luminescence dating of Quaternary sediments in the Himalaya and High Asia: a practical guide to its use and limitations for constraining the timing of glaciation. Quaternary International 65/66, 49–61.

Sissons, J.B., 1967. The evolution of Scotland's scenery. Oliver & Boyd, Edinburgh, 259 pp.

Spencer, J.Q., Owen, L.A., 2004. Optically stimulated luminescence dating of Late Quaternary glaciogenic sediments in the upper Hunza valley: validating the timing of glaciation and assessing dating methods. Quaternary Science Reviews 23, 175-191.

Stephenson, D., Gould, D., 1995. British regional Geology: the Grampian Highlands. 4th ed., HMSO, London.

Wallinga, J., 2002. Optically stimulated luminescence dating of fluvial deposits: a review. Boreas 31, 303-322.

Figure captions

Fig. 1. Overview of study area, OSL sampling locations and location of initial ^{14}C date

Fig. 2. Sample BM 3-1 quartz OSL. (a) Natural; (b) 21.8 Gy test dose. Note significant medium-slow components.

Fig. 3. Natural LM-OSL measured from 0-95% power with no preheat. (a) BM 2-2; (b) BM 3-1. Fast component is absent or very weak.

Fig. 4. SAR IRSL growth curves from 3 mm aliquots of K-rich feldspar. (a) BM 3-1; (b) BM 2-2; (c) Lsh 1; (d) SH 1. Beta source dose-rate ~0.1 Gys^{-1}.

Fig. 5. K-feldspar single grain D_e data from sample BM 2-2. Lowest value overestimates expected D_e by ~2-3 times.

Fig. 6. CL image from 3 mm feldspar aliquot of BM 2-2. A few seconds exposure at 15 kV & 600 μA (Technocyn Mk 3 CL with 15 kV flood-gun electron source & Nikon Optiphot microscope with DVC high-sensitivity digital camera).

Table 1. Elemental concentrations from *in situ* gamma spectrometry[a], location information, cosmic dose rates and total dose-rates for quartz and feldspar minerals

Sample #	U	Th	K	Lat	Long	Alt	Depth	Cosmic[b]	Total dose-rate ($mGya^{-1}$)[c]	
	(ppm)	(ppm)	(%)	(°N)	(°E)	(m)	(cm)	($mGya^{-1}$)	Quartz	Feldspar[d]
BM 2-2	1.66±0.10	7.05±0.25	1.84±0.03	58.2666	-4.750	215	250	0.157	2.82±0.15	3.64±0.32
BM 3-1	1.38±0.09	4.29±0.19	1.57±0.02	58.3083	-4.666	230	70	0.201	2.36±0.15	3.17±0.32
LSh 1	1.50±0.09	5.12±0.21	1.39±0.02	58.1750	-4.675	100	120	0.183	2.25±0.15	3.07±0.32
SH 1	1.30±0.09	5.23±0.21	1.89±0.03	58.1166	-4.633	150	70	0.198	2.70±0.15	3.52±0.32

[a] Calculated using E. Rhodes' conversion software.

[b] Contribution to dose-rate from cosmic rays calculated according to Prescott & Hutton (1994) assuming constant burial depth using present day overburden measurements. Uncertainty taken as 10%.

[c] Total dose-rates calculated using conversion factors of Adamiec & Aitken (1998) and assuming external Rb of 100±10 ppm. External beta attenuation factors from R. Grün's revised 'Age' program. An additional error term has been included to allow for systematic uncertainty in present moisture content compared to average burial conditions.

[d] Assumed internal K of 12.5% (after Huntley and Baril, 1997) ±5% using absorbed dose fractions after Mejdahl (1979).

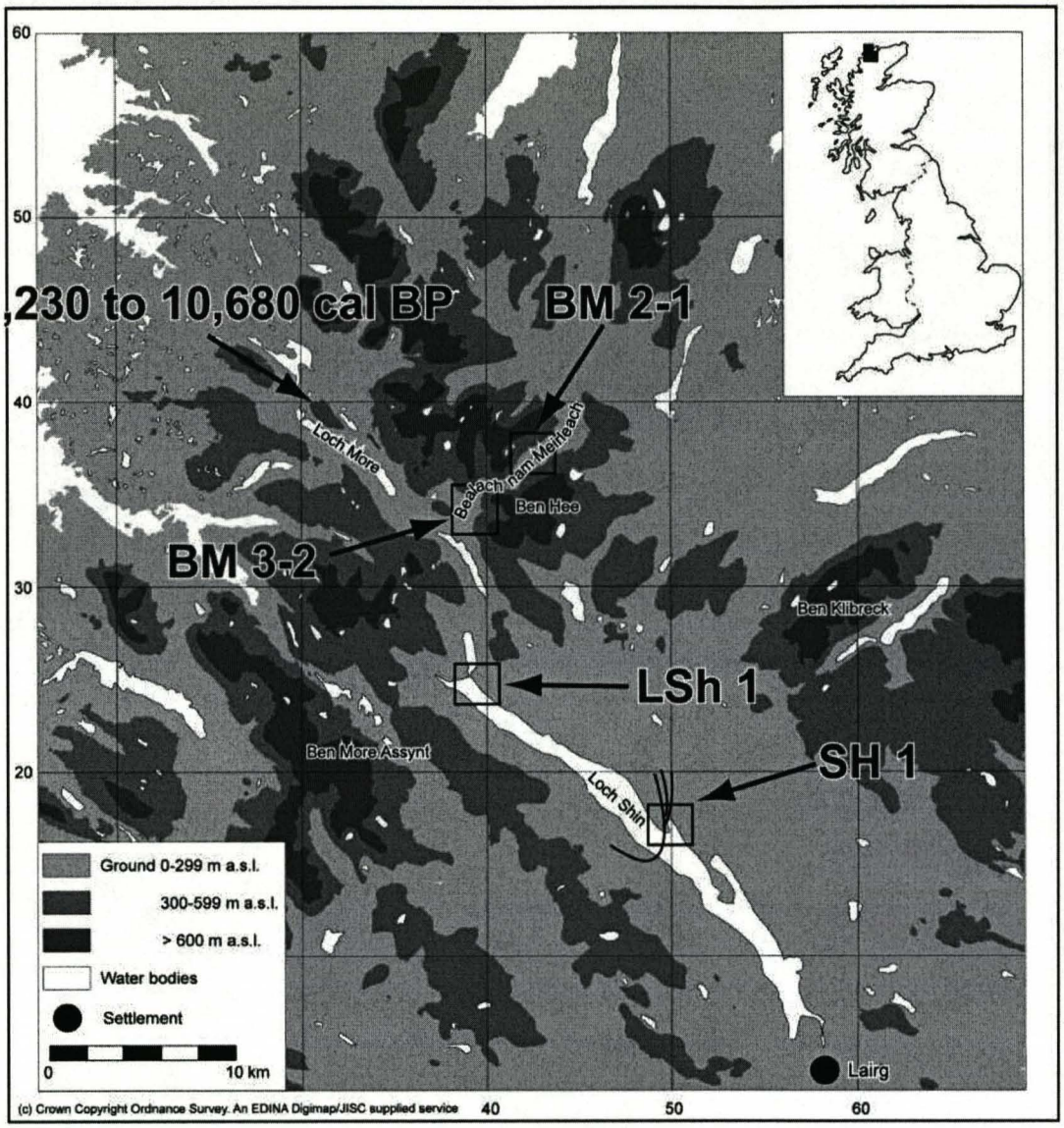

11,230 to 10,680 cal BP
BM 2-1
BM 3-2
LSh 1
SH 1
Loch More
Beatach nam Meineach
Ben Hee
Ben Klibreck
Ben More Assynt
Loch Shin
Lairg
Ground 0-299 m a.s.l.
300-599 m a.s.l.
> 600 m a.s.l.
Water bodies
Settlement
0
10 km
(c) Crown Copyright Ordnance Survey. An EDINA Digimap/JISC supplied service
60
50
40
30
20
40
50
60

Figure 1

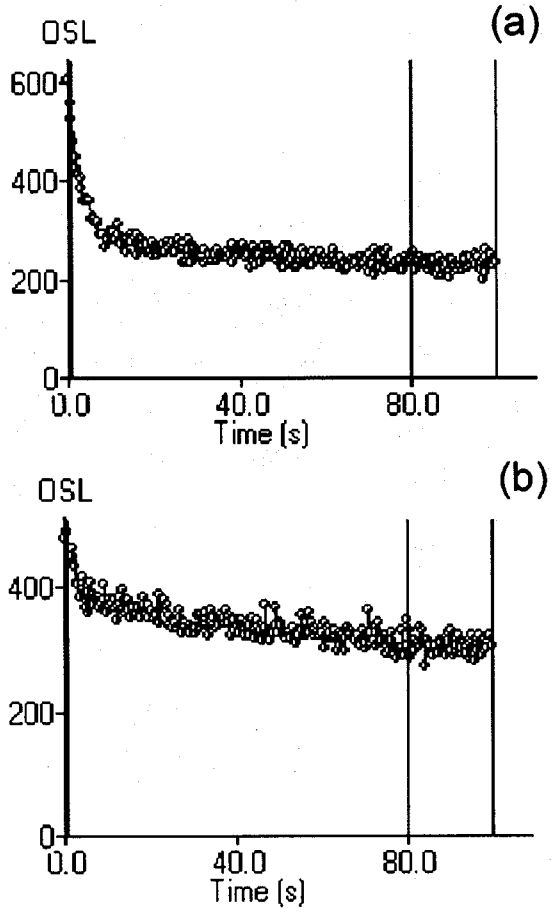

OSL
(a)
600
400
200
0
0.0
40.0
80.0
Time (s)
OSL
(b)
400
200
0
0.0
40.0
80.0
Time (s)

Figure 2

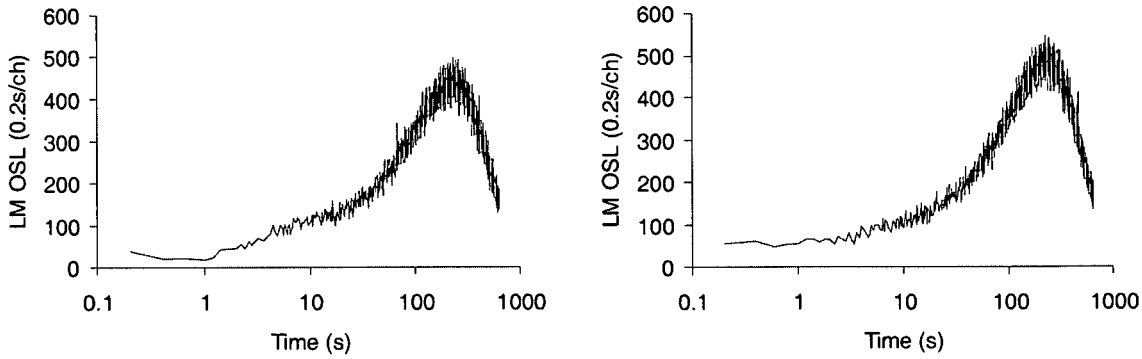

600
500
400
300
200
100
0
LM OSL (0.2s/ch)
0.1
1
10
100
1000
Time (s)
600
500
400
300
200
100
0
LM OSL (0.2s/ch)
0.1
1
10
100
1000
Time (s)

Figure 3

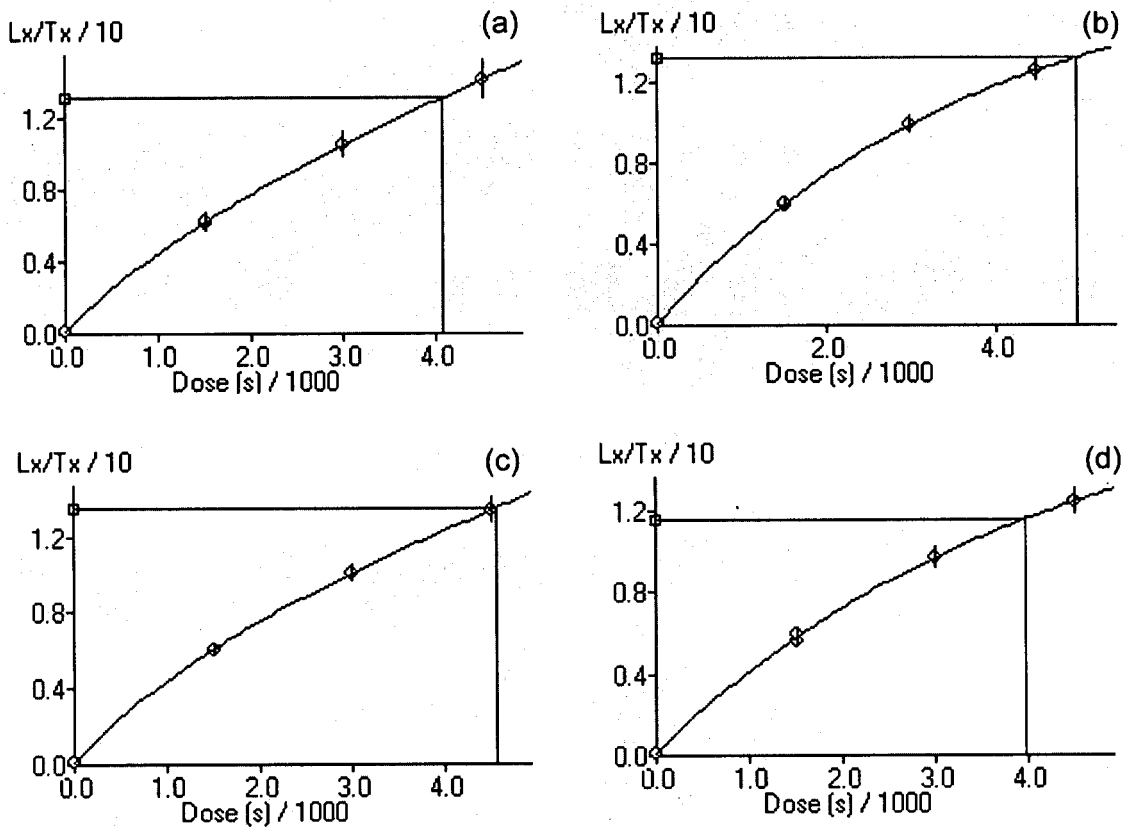

(a)
Lx/Tx / 10
1.2
0.8
0.4
0.0
0.0
1.0
2.0
3.0
4.0
Dose (s) / 1000
(b)
Lx/Tx / 10
1.2
0.8
0.4
0.0
0.0
2.0
4.0
Dose (s) / 1000
(c)
Lx/Tx / 10
1.2
0.8
0.4
0.0
0.0
1.0
2.0
3.0
4.0
Dose (s) / 1000
(d)
Lx/Tx / 10
1.2
0.8
0.4
0.0
0.0
1.0
2.0
3.0
4.0
Dose (s) / 1000

Figure 4

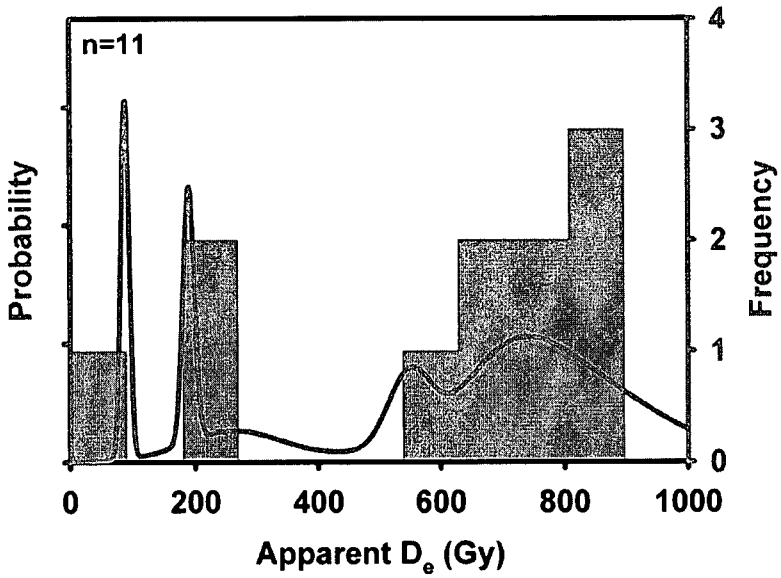

n=11
Probability
Frequency
0
1
2
3
4
0
200
400
600
800
1000
Apparent D_e (Gy)

Figure 5

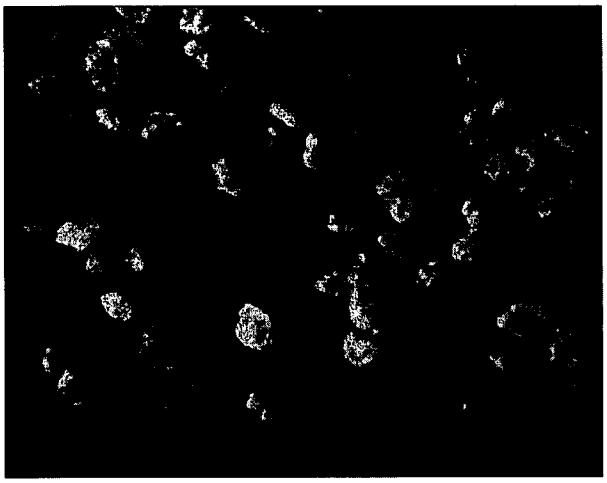
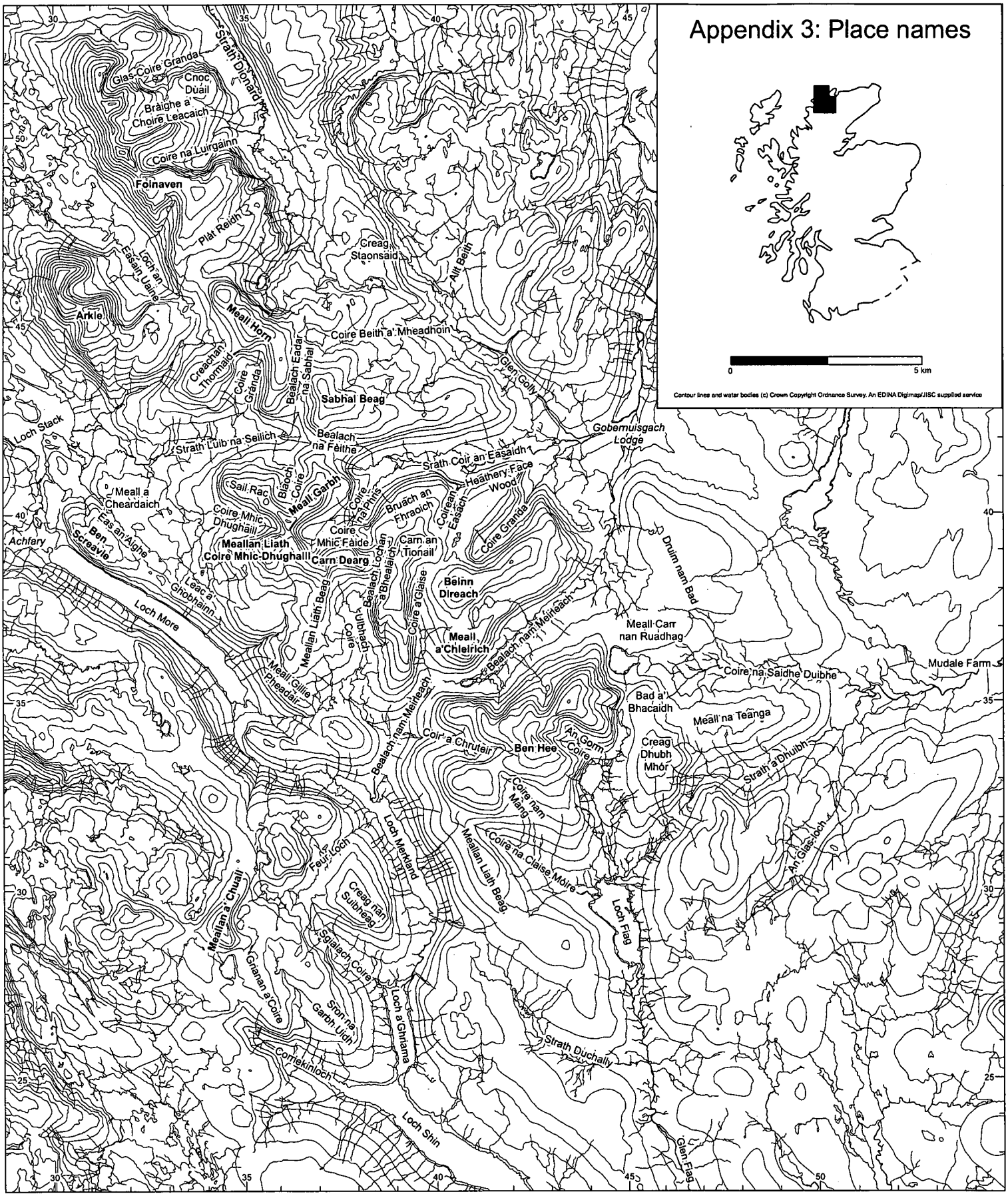


Figure 6

Appendix 3: Place names



Appendix 2
Geomorphological map of western and central Sutherland

- Moraine ridges and mounds
- Fluted moraines
- Till sheets or prisms
- Glacially-transported boulders
- Boulder line or limit
- Meltwater channel (with clear margin)
- Upslope termination of sediment blanket
- Dissected/gullied sediment cover
- Trimline
- Upslope termination of ice-moulded bedrock
- Roches moutonnées
- Striae
- Ice-moulded bedrock with clear indication of ice flow
- Ice-moulded bedrock without clear indication of ice flow
- Frost-weathered glacial erosional landforms
- Subglacial surface with lodged boulders
- Outwash terrace upper level (t1)
- Outwash terrace intermediate level (t2)
- Outwash terrace lower level (t3)
- Steep rock face
- Frost-weathered bedrock
- Bedrock at surface
- Mountaintop detritus
- Solifluction lobes
- Rockslope failure
- Thick talus accumulation
- Thick peat
- Alluvial fan
- Contour line at 50 m intervals
- Mountaintop (height in metres OD)

0 — 3 km

Loch Stack
Coire Loch
Loch More
Loch Merkland
Loch a' Ghriama
Loch Shin
Loch Fiag