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## The Glacial Geomorphology of the Loch Lomond (Younger Dryas) Stadial in Britain

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# The Glacial Geomorphology of the Loch Lomond (Younger Dryas) Stadial in Britain

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# The Glacial Geomorphology of the Loch Lomond (Younger Dryas) Stadial in Britain

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## Abstract

The Loch Lomond Stadial (LLS) refers to the abrupt return to severe cold conditions that occurred in Britain, between 12.9 and 11.7 ka, subsequent to the retreat of the last (Late Devensian) British-Irish Ice Sheet. This period has long been associated with the regrowth of glaciers in upland areas of Britain and left a wealth of geomorphological evidence in the landform record. However, previous research on these glaciers has largely comprised localised case studies, producing a fragmented and spatially inconsistent dataset. This thesis draws together the published geomorphological evidence for Loch Lomond Stadial glaciation to build a coherent picture of the extent, style and dynamics of glaciation during the stadial. Geomorphological mapping of glacial landforms associated with this period is compiled from the published literature to create a map and geographic information systems database of over 95,000 features. The evidence used to produce this map is critically assessed in the most comprehensive review of the Loch Lomond Stadial to date and is used to identify conceptual themes, common to the geomorphology in multiple sub-regions within Scotland, England and Wales. Persisting uncertainties, particularly regarding the extent and timing of Loch Lomond Stadial glaciation, are discussed and recommendations of future research to address these are made. Building on this review, the glacial geomorphological map is then used to construct five glacial landsystem models which reflect the style of Loch Lomond Stadial glaciation; the cirque/niche glacier landsystem, the alpine icefield landsystem, the lowland piedmont lobe landsystem, the plateau icefield landsystem and the ice cap landsystem. Use of these models to classify the Loch Lomond Stadial glacial geomorphology reveals the spatial distribution of each landsystem. Three styles of glacier retreat are represented by the glacial geomorphology. It is demonstrated that both landsystem and retreat style reflect the combined importance of pre-existing topography and palaeoclimate. Given the paucity of dating constraints on Loch Lomond Stadial landforms, the thesis pilots the use of a relative dating technique using soil chronosequences to differentiate between Loch Lomond Stadial and older moraines in the English Lake District. The results of this study highlight the potential of this technique to discriminate between Loch Lomond Stadial and pre-Loch Lomond Stadial moraines in Britain, although further work is required.

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## List of abbreviations

| | |
|---|---|
| AMOC | Atlantic Meridional Overturning Circulation |
| CDE | Colour Development Equivalent |
| DEM | Digital Elevation Model |
| DSM | Digital Surface Model |
| DTM | Digital Terrain Model |
| ELA | Equilibrium Line Altitude |
| GCP | Ground Control Point |
| GIS | Geographic Information System |
| GRIP | Greenland Ice Core Project |
| LGM | Last Glacial Maximum |
| LLS | Loch Lomond Stadial |
| LOI | Loss on ignition |
| m.a.s.l. | Metres above sea level |
| OD | Ordnance Datum |
| OS | Ordnance Survey |
| RMS | Root-mean-square |
| YD | Younger Dryas |

All ages quoted are in calendar years before present unless otherwise stated.

## Declaration and statement of copyright

I confirm that no part of the material presented in this thesis has previously been submitted by me or any other person for a degree in this or any other university. In all cases, where relevant, material from the work of others has been acknowledged.

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# Acknowledgements

# Chapter 1
# Introduction

## 1.1 Rationale

The Younger Dryas (YD) (12.9-11.7 ka) was characterised by an abrupt return to severe cold conditions, particularly in the regions surrounding the North Atlantic Ocean (Carlson, 2010). Such was the abruptness of this event that changes in climatic conditions are thought to have occurred over much of the Earth within thirty years of their signal appearing in Greenland ice core records (Alley, 2000). Thought to be the result of disruption to the Atlantic meridional overturning circulation (Carlson, 2010; Golledge, 2010), the period is most often associated with the regrowth or readvance of glaciers in North America (Lowell et al., 1999), Iceland (Ingólfsson et al., 2010), Scandinavia (Andersen et al., 1995; Mangerud et al., 2011), the Swiss Alps, Britain (Golledge, 2010) and Greenland, where temperatures dropped to 15°C below modern values (Alley, 2000).

However, the global signature of this climate event is complex, with much of the Southern Hemisphere warming, as indicated by an increase in $\delta^{18}O$ in Antarctic ice cores (Blunier and Brook, 2001; Carlson, 2010), the lack of evidence for a Younger Dryas readvance of glaciers on South Georgia (Bentley et al., 2007), and pollen records indicating general warming during this period from South America. Although a possible Younger Dryas readvance has been postulated for the Franz Joseph glacier in New Zealand (Denton and Hendy, 1994), this suggestion has been challenged (Torvar et al., 2008) and is at odds with speleothem evidence indicative of warming during the Stadial (Williams et al., 2005). This contrast between the climate cooling in the Northern Hemisphere and warming in the Southern Hemisphere contributed to a global net cooling of just 0.6°C, and has been attributed to the bipolar response of atmospheric and oceanic teleconnections (Carlson, 2010; Shakun and Carlson, 2010). Consequently, the Younger Dryas provides an excellent case study for abrupt climate change and thus an analogue for how elements of the atmospheric, oceanic and cryospheric systems might respond during possible future periods of rapid climate change.

Given that the driving force behind Younger Dryas cooling of the North Atlantic regions has been attributed to weakening of Atlantic meridional overturning circulation, it is unsurprising that the effects of this cooling were felt strongly in Great Britain, where the climate is currently greatly ameliorated by this mode of ocean circulation. Within the body of Younger Dryas research in Britain, particular focus has been given to identifying the evidence for glaciation during this period and reconstructing the extents of these ice

masses (e.g. Sissons, 1974; Thorp, 1984; Ballantyne, 2002a; Benn et al., 1992; Lukas and Bradwell, 2010; Boston et al., 2015). This is perhaps because the majority of Younger Dryas glaciers in Britain were relatively small in their extent and terminated on land, resulting in a wealth of geomorphological evidence that has made them popular for study. Since these glaciers reflect the cryospheric response to the abruptly changing climate during the Younger Dryas, developing a thorough understanding of their extent and behaviour forms an important component in the larger picture of post-Late Devensian climate change in the Northern Hemisphere. Furthermore, the possibility has been raised of similar Younger Dryas-like events following previous deglaciations (Carlson, 2008, 2010) and hence, improving our understanding of the Younger Dryas could provide an insight into the drivers of climate change over much longer timescales.

The glacial history of Great Britain, including during the Younger Dryas, has attracted the attention of researchers for well over 170 years, since Agassiz (1840) first identified evidence to suggest that Britain, along with much of northern Europe, had previously been occupied by glaciers. It was soon recognised that the glaciation of Britain by an ice sheet, had been succeeded by a phase of valley glaciation (Forbes, 1846; Jolly, 1868; Geikie, 1878). This period of glacial readvance later became known as the Loch Lomond Stadial (LLS), based on Simpson's (1933) identification of the limits of this phase of glaciation in the Lomond Valley, Scotland. Numerous studies have since shown that this readvance occurred between 12.9 and 11.7 ka (e.g. Alley, 2000; Brooks and Birks, 2000; Rasmussen et al., 2006; Lowe et al., 2008), subsequent to the retreat of the last (Late Devensian) British-Irish Ice Sheet, and it is accepted that the British LLS represents the regional signal of the global Younger Dryas Stadial (e.g. Benn et al., 1992; Bradwell, 2006; Lukas and Benn, 2006; Golledge, 2010; Ballantyne, 2012; Bendle and Glasser, 2012; McDougall, 2013; Boston et al., 2015) (Figure 1.1). Throughout this thesis, the term 'Younger Dryas' has been used when discussing this event on a global scale, whilst 'Loch Lomond Stadial' has been used when referring to the period specifically in Britain, as it is acknowledged that the precise synchroneity of these two events remains to be established.

Much of the research on Britain's glacial history comprises localised case studies (e.g. Hollingworth, 1931; Sissons, 1958, 1974; Stoker, 1988; Brown, 1993; McDougall, 2001, 2013; Ballantyne, 2002a, 2007a, b; Lukas and Lukas, 2006a, b), undertaken by a large number of independent researchers and using a variety of methodologies, which inhibits regional scale analysis of the extent and behaviour of glaciation across Britain. A significant step in overcoming this fragmented and spatially inconsistent dataset was taken by Clark et al. (2004) who compiled geomorphological evidence of the Late

Devensian Ice Sheet from the published literature (as detailed in Evans et al., 2005) into a geographic information system (GIS) database and glacial map. Since their project concentrated on features attributed to the ice sheet, geomorphological evidence associated with LLS glaciation was not included. Only the extent of the West Highland Glacier Complex, as compiled from summaries of LLS glacier extent (detailed in Clark et al. 2004), was included, in order to explain the lack of ice sheet geomorphology within that area.

As well as the non-systematic nature of much of the research associated with LLS glaciation in Britain, a common difficulty encountered in several localities has been the relative paucity of chronological controls on landforms associated with this period (e.g. Sissons, 1974; McDougall, 2013). Even in locations where dating has been attempted, this has frequently not been without problems and uncertainties (e.g. Ballantyne, 2012; Wilson et al., 2013; Standell, 2014). Given that the LLS glaciers occupied areas previously overridden by the last British-Irish Ice Sheet (Clark et al., 2012), landforms from both glacial phases may form a palimpsest landscape, making it difficult to differentiate between features of different ages. This has implications for identifying the maximum extents of LLS glaciers, particularly when these limits are used in glacial reconstructions from which palaeoclimatic conditions are inferred (e.g. Ballantyne, 2002a; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010).

This thesis focuses on drawing together the geomorphological evidence of the LLS glaciation in Britain, forming a more coherent picture of the extent and nature of these ice masses than has previously been possible. In doing so, the quality of the existing evidence is assessed and areas of continuing uncertainty are highlighted. In addition, the potential of an alternative approach to relative dating is tested.

It should be noted that the scope of this thesis is limited to the LLS glacial geomorphology of Scotland, England and Wales. This was done for several reasons. Firstly, whilst it seems probable that the Nahanagan Stadial, during which glaciers regrew in Ireland (Gray and Coxon, 1991), corresponds to the global Younger Dryas, assessing the synchroneity of this period with the British LLS is difficult given the paucity of absolute dates. On a practical note, by restricting study to Great Britain a consistent approach using Ordnance Survey maps in British National Grid and NEXTMap data could be adopted, whereas these resources do not cover Ireland.
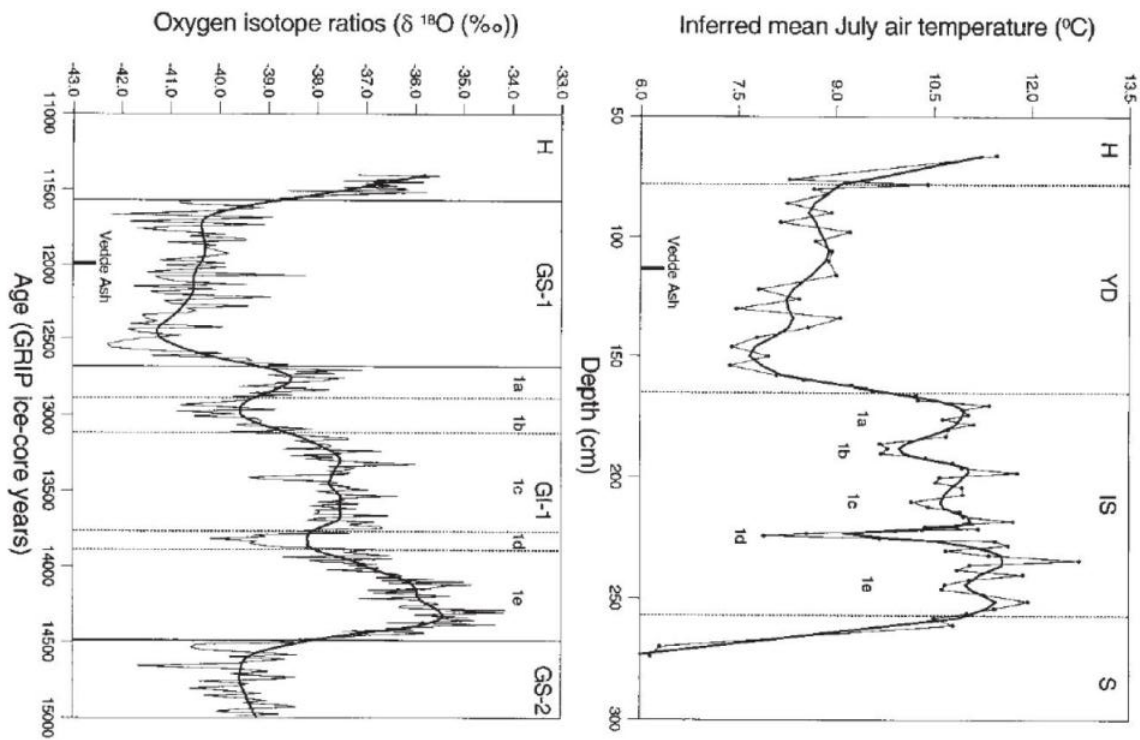
Figure 1.1 Oxygen isotope record from the Greenland Ice Core Project (GRIP) showing the global signal of the Younger Dryas (Greenland Stadial-1, GS-1), illustrating the marked different in δ18O ratios during this period, compared to a chironomid-inferred mean July air temperature record from Whitrig Bog, Southern Uplands, Scotland. Reprinted from Journal of Quaternary Science, Vol. 15(8), Brooks, S.J. and Birks, H.J.B. Chironomid-inferred Lateglacial air temperatures at Whitrig Bog, southeast Scotland, 759-764. Copyright (2000) with permission from John Wiley and Sons.

## 1.2 Research questions, aims and objectives

This thesis seeks to answer the following research questions:

1. What is the range of evidence that has been used to map the LLS glaciation in Britain?

2. How robust are the previously mapped limits of LLS glaciation, both spatially and temporally?

3. What was the style of LLS glaciation and what factors influenced this?

4. Can soil chronosequences be used to resolve some of the current chronological uncertainty associated with LLS ice limits?

The aims of the thesis are:

1. To draw together published evidence of the geomorphology of LLS glaciation in Britain in order to build a coherent picture of the extent, style and dynamics of glaciation during the stadial.

2. To test a relative dating approach using soil chronosequences in providing chronological control on glacial landforms.

These aims will be met through the following objectives:

1. To compile the glacial geomorphology of the LLS from the published literature into a GIS database and glacial map.

2. To systematically review this evidence, identifying its strengths and weaknesses, and to make recommendations for future research to address current deficiencies.

3. To identify the type of glacial landsystems and retreat styles represented by the LLS geomorphology and to map the distribution of landsystems at a regional scale to determine possible controls on the nature of LLS glaciation across Britain.

4. To undertake a pilot study into the potential use of soil chronosequences as a tool to provide a relative chronology on moraines and differentiate between LLS and older landforms in the English Lake District.


## 1.3 Thesis structure and results

Each of the objectives above are addressed in a series of research papers that constitute this thesis. Adapted versions of these papers have been published, submitted or prepared for peer-reviewed journals. Chapter 2 presents a new map and database of the LLS glacial geomorphology in Britain, compiled from previously published evidence (Objective 1) and Chapter 3 provides a review of the evidence which comprises it (Objective 2). Adopting a landsystems approach, Chapter 4 uses the database to examine the style of glaciation and different retreat patterns (Objective 3) and Chapter 5 focuses on dating these features using the soil chronosequence method (Objective 4). Chapter 6 draws together the main conclusions from each of the papers to answer the research questions proposed above.

### 1.3.1 Chapter 2

Bickerdike, H.L., Evans, D.J.A., Ó Cofaigh, C. and Stokes, C.R. (2016) 'The glacial geomorphology of the Loch Lomond Stadial in Britain: a map and geographic information system resource of published evidence', *Journal of Maps*, 12: 1178-1186. DOI: 10.1080/17445647.2016.1145149

This paper details the motivation for, and approach to, the creation of a new map and GIS database of LLS glacial geomorphology, which is presented. Evidence was compiled from the published literature and was georeferenced and digitised into the GIS. The results of the mapping show that recessional and hummocky moraines are particularly widespread within the extent of the LLS glaciers but that there are variations between the landform assemblages displayed at different locations. The map and GIS database overcome the non-systematic and fragmented nature of previous studies of LLS glacial geomorphology but highlight the variable quality and uneven quantity of existing mapping.

For this paper, I compiled and digitised data from the published literature into the GIS database. Data for specific locations (Snowdonia, Beinn Dearg and the eastern Lake District) were provided as shapefiles by the original authors of each study, Jacob Bendle, Andrew Finlayson and Derek McDougall. I wrote the manuscript and created the map and figures. All authors contributed ideas, advised in the creation of the map and commented on the text. The paper has been published in *Journal of Maps*. The map itself is available as a fold-out in the back of this thesis and is also available, with the shapefiles, from the Journal of Maps website.

### 1.3.2 Chapter 3

Bickerdike, H.L., Evans, D.J.A., Stokes, C.R. and Ó Cofaigh. C. (in review) 'The glacial geomorphology of the Loch Lomond Stadial in Britain: a review', *Journal of Quaternary Science.*

This paper is a systematic review of the geomorphological evidence of LLS glaciation in Britain. The paper critically assesses the evidence in Scotland, England and Wales, providing the most comprehensive review to date. The final section of the paper draws together conceptual themes common of multiple sub-regions. These include: the uneven quality and quantity of evidence; difficulties in assessing the timing of the LLS and identifying the age of landforms; the changing paradigm of the style of LLS glaciation; the results of numerical modelling experiments; and the approximate extent of LLS glaciation. In this final section, recommendations to directly address current uncertainties associated with LLS glaciation are made to set the agenda for future research in this field.

For this paper, I wrote the manuscript and drew the figures. All authors contributed ideas and commented on the text. Photographs B-F in Figure 3.6 were provided by D.J.A. Evans and the photograph in Figure 3.13 was provided by H. Kinley.

### 1.3.3 Chapter 4

Bickerdike, H.L., Ó Cofaigh, C., Evans, D.J.A. and Stokes, C.R. (in press) 'Glacial landsystems, retreat dynamics and regional controls on Loch Lomond Stadial glaciation in Britain', *Boreas.*

This paper uses the geomorphology in the GIS database to identify five glacial landsystem models: the cirque/niche glacier landsystem, the alpine icefield landsystem, the lowland piedmont lobe landsystem, the plateau icefield landsystem and the ice cap landsystem. Geomorphological features common to each landsystem are identified and process-form relationships are discussed. The landsystems are then used to build a first-order

classification of LLS ice masses, allowing spatial patterns in the distribution of these landsystems to be identified. Alongside the landsystem types, three retreat styles are inferred from the geomorphology: active retreat throughout deglaciation, two phase retreat and uninterrupted retreat. The importance of both pre-existing topography and palaeoclimate in controlling the distribution, extent and retreat dynamics of these glaciers is then discussed.

For this paper, I wrote the manuscript and drew the figures. All authors contributed ideas and commented on the text.

### 1.3.4 Chapter 5

Bickerdike, H.L., Evans, D.J.A., Ó Cofaigh, C. and Stokes, C.R. (in prep.) 'Relative age dating of moraines in the English Lake District using soil chronosequences: a pilot study', *Boreas.*

This paper presents the preliminary results of a pilot study to test the potential of soil chronosequences as a tool to assign relative ages to moraines in the English Lake District, in order to differentiate between LLS and older features. Pits were excavated on inset moraine sequences in nine valleys within the English Lake District. Soil depths and horizon thicknesses were measured in the field and samples were collected for analysis of colour, loss on ignition, and particle size. No sequential change in these latter three variables with relative age was observed, but the total soil depths and combined E and B horizon thicknesses generally increase with relative age. Existing chronological control at one site (Mosedale) is used to calibrate the chronosequence and suggests that moraines which predate the LLS are present in some valleys within the study area. Whilst it appears that this technique has some potential to differentiate between different aged moraines, the resolution of the technique is coarse and further independent chronological control is required to refine this.

For this paper, I wrote the manuscript and created the figures. The fieldwork was conducted by myself and the authors, particularly D.J.A. Evans, and field assistants acknowledged at the beginning of this thesis. I conducted laboratory tests on the samples.

# Chapter 2

# The glacial geomorphology of the Loch Lomond Stadial in Britain: a map and geographic information system resource of published evidence

## Abstract

The Loch Lomond Stadial (LLS) was an abrupt period of renewed cooling between 12.9 and 11.7 ka and has long been associated with the regrowth of glaciers in much of upland Britain. Mapping the glacial landforms associated with this period has been undertaken for over a century, but in a non-systematic nature and at specific locations. In this paper, glacial geomorphology associated with the LLS in Britain has been compiled from the published literature into a glacial map and accompanying geographical information system database that is available electronically as supplementary information. A variety of scales have been used to best represent the evidence in the database. Map A is at 1:310,000; B, C, D, E, F, J, L, M and O are at 1:175,000; K, N, P are at 1:100,000 and G, H and I are at 1:50,000. The database contains over 95,000 individual features, which are organised into thematic layers and each attributed to its original citation. The evidence includes moraines, drift and boulder limits, drift benches, periglacial trimlines, meltwater channels, eskers, striations and roches moutonneés, protalus ramparts and ice-dammed lakes. Creation of this database overcomes the drawbacks posed by the non-systematic nature of previous mapping output from studies of LLS glaciation. It is intended to be a catalyst for future research in this area, with especial significance for regional palaeoglaciological and palaeoclimatic reconstructions of the Younger Dryas and numerical modelling.

## 2.1 Introduction

The glacial history of Britain has attracted the attention of researchers for over a century, inspiring a large and evolving body of research and published output. However, much of this research comprises localised case studies which inhibit a more holistic understanding of the extent and dynamics of regional glaciation styles. A significant step in overcoming this problem was the BRITICE Project, which created a 'Glacial Map of Britain' and geographic information system (GIS) database of over 20,000 individual features dated to the last (Late Devensian) British Ice Sheet (Clark et al., 2004; Evans et al., 2005). This has facilitated evidence based reconstructions of the extent and dynamics of the ice sheet (Clark et al., 2012). Geomorphological evidence dating to the Loch Lomond Stadial (LLS), however, was excluded from BRITICE.

The LLS period was characterised by an abrupt return to severe cold conditions following ice sheet retreat when glaciers regrew, forming most significantly a large icefield (the West Highland Glacier Complex) running the length of the western Highlands (Thorp, 1986; Bennett and Boulton, 1993a). This icefield was the largest of numerous satellite icefields, ice caps, valley glaciers and cirque/niche glaciers on several of the Hebridean islands and at other upland sites around Britain and Ireland (Golledge, 2010). The BRITICE database included the limit of the LLS icefield in the western Highlands only. This was done primarily to indicate that the area was subjected to renewed mountain icefield glaciation after ice sheet recession and to explain the consequent absence of data from this area, and was compiled from relatively few sources (Clark et al., 2004).

The main map presented in this paper is the product of a comprehensive compilation of LLS features in a GIS database, aimed at improving our understanding of the local and regional signatures of glaciation style, extent and dynamics during this critical part of the Late Quaternary. This compilation aims to overcome the fragmented and spatially inconsistent nature of much of the published LLS research by synthesising it into a single body of evidence, while retaining the valuable high resolution mapping achieved in localised case studies. The database facilitates comparison of landform assemblages, and therefore palaeo-ice dynamics, at scales spanning individual mountain ranges to regional uplands and local valleys. Although several publications have shown the reconstructed extent of LLS ice over large areas or between drainage basins (Charlesworth, 1955; Golledge, 2010; McDougall, 2013) the map and database presented here are the first to collate evidence to inform reconstructions of the extent of LLS glaciation from the whole region of upland Britain.

## 2.2 Methods

### 2.2.1 Selection of evidence

The BRITICE map and database (Clark et al., 2004) was used as a template for informing decisions regarding what evidence should be included/excluded and how this should be organised. Evidence within the database originates from academic journal articles, field guides and theses. Cross-referencing between sources was used to identify relevant papers and Ph.D. theses, rather than searching through all journals or institutions. No work from below Ph.D. level has been included, in order to ensure that sources are accessible to other researchers. Although the number of publications covered is extensive (more than 200), it is expected that some sources will have been overlooked. The census date for data inclusion in this project is June 2014 but the intention is that the database can be updated as future research becomes available.

The geomorphological evidence has been organised into a series of layers, each containing a different landform type, similar to the BRITICE project (Clark et al., 2004). These include moraines (individual moraine mounds and ridges, fluted moraines and areas of hummocky moraine), drift and boulder limits, drift benches, meltwater channels, eskers, periglacial trimlines, striations and roches moutonneés, and protalus ramparts. Major ice-dammed lakes associated with LLS glaciers are also mapped. In some cases (e.g. in Glens Doe, Gloy, Spean and Roy), the authors report the extent of the lakes at multiple stages during the LLS, but in most publications the lakes have only been reported as dammed by LLS glaciers at their maximum extent (e.g. Benn and Ballantyne, 2005; MacLeod et al., 2011).

Some evidence could not be easily presented in map form, for example sedimentological characteristics, whilst some landform types either had not been mapped systematically throughout the literature (glacial outwash) or just replicated the information derived from more commonly mapped features. It was concluded that including the distribution of glacigenic deposits would largely replicate the Quaternary Map of the United Kingdom (IGS, 1977) and thus this has not been attempted. Although most LLS glaciers terminated onshore, some extended beyond the current coastline into particular sea lochs along the western coast of Scotland. Where moraines have been mapped from bathymetric surveys of these sites and have been attributed to the LLS, these have been included in the database (Boulton et al., 1981; Dix and Duck, 2000; McIntyre et al., 2011).

Some differences in approach between this study and the BRITICE database have been necessary. Striations were not included in the BRITICE mapping because it was argued

that these are not reliable indicators of ice flow and the volume of data made the task unrealistic (Clark et al., 2004). However, differential striation directions have been instrumental in reconstructing the extent of LLS glaciers in several locations, sometimes being amongst the most important landform evidence available, for example, at Loch Coruisk, Isle of Skye (Ballantyne, 1989). It was, therefore, deemed necessary to include striations from the source maps to show how such reconstructions have been derived. Similarly, hummocky moraine was excluded from the BRITICE mapping due to inconsistencies within the literature regarding the definition and genesis of this landform (Clark et al., 2004). Given that this landform type was once interpreted as diagnostic of a LLS age and has been central to several reconstructions of LLS glaciers (Sissons and Grant, 1972; Sissons, 1972, 1974), it was deemed important to include it here. Protalus or pronival ramparts are also included as some of these features have previously been linked to a glacial origin and, indeed, debate continues about the glacial versus periglacial origin of such landforms in many areas (Carr et al., 2007a; Shakesby, 2007).

### 2.2.2 Map formatting and georeferencing

The reliability of the database is largely dependent on the quality of the original maps from which it is built. Consequently, selection and inputting of appropriate maps have been crucial. Where multiple maps covered the same area, preference has usually been given to the most comprehensive and/or highest resolution map, unless its findings have been refuted in subsequent literature. Some areas, therefore, show specific sites of fine detailed mapping, within a region of more generalised coverage.

The procedure to add data to the map is outlined in Figure 2.1. Geomorphological maps from papers (e.g. Figure 2.2) were scanned and stored as raster graphics files (.tiffs). These were added as layers into ArcMap over a hillshade model derived from a 5 m resolution NEXTMap digital surface model (DSM). The maps were then georeferenced to the OS British National Grid system. Where maps had pre-existing gridlines or tick marks that could be used to create an overlying grid using Adobe Photoshop, this was a straightforward task as the coordinates of ground control points (GCPs) on the grid are known and can be inputted directly, and accurately, into ArcMap (Clark, 1997). These points were distributed evenly across the map, particularly around the edges, to reduce distortion. Usually a minimum of eight points were used, although this was not always possible for maps of limited coverage.
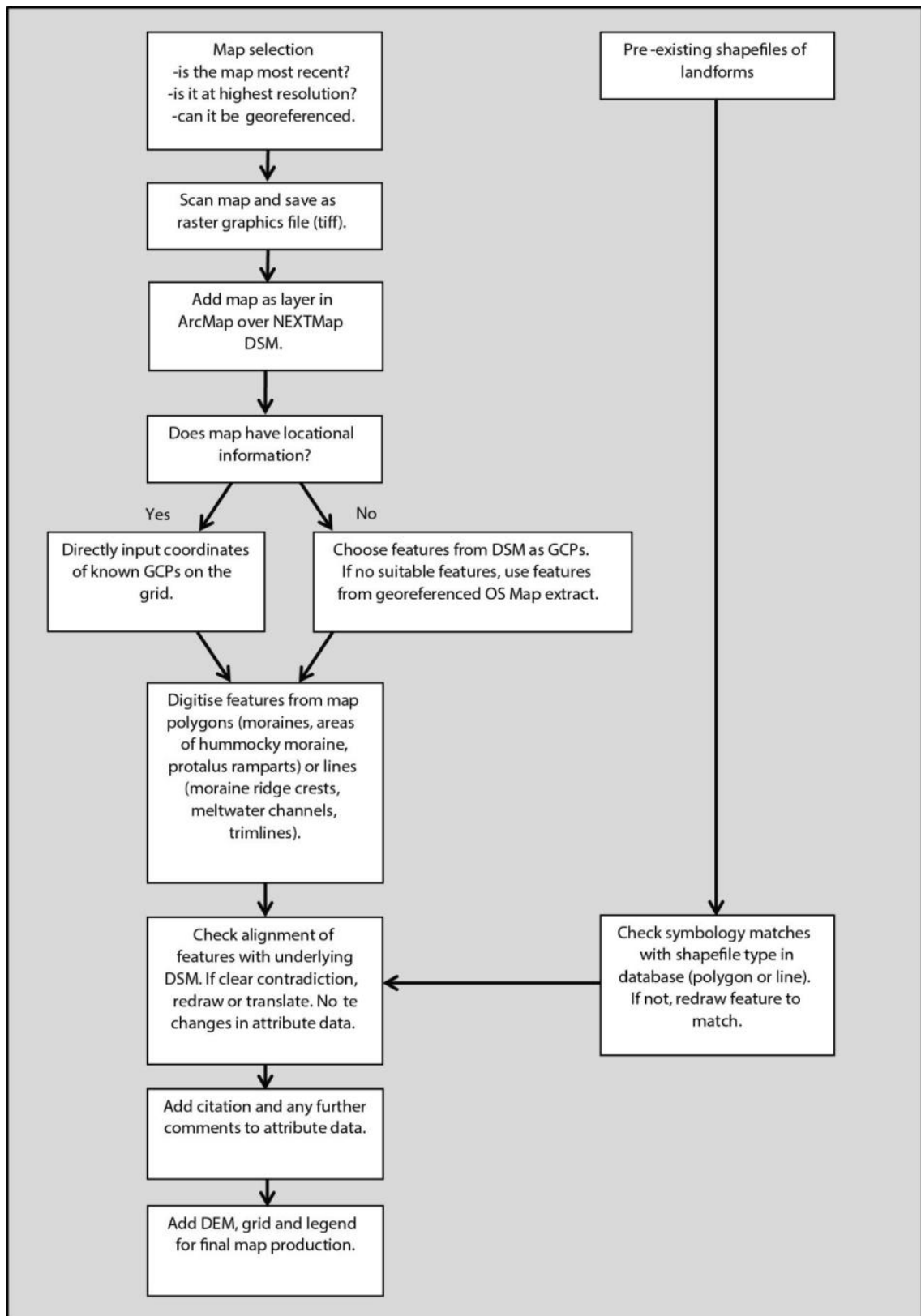
Figure 2.1 Flow chart showing the stages of GIS and map production
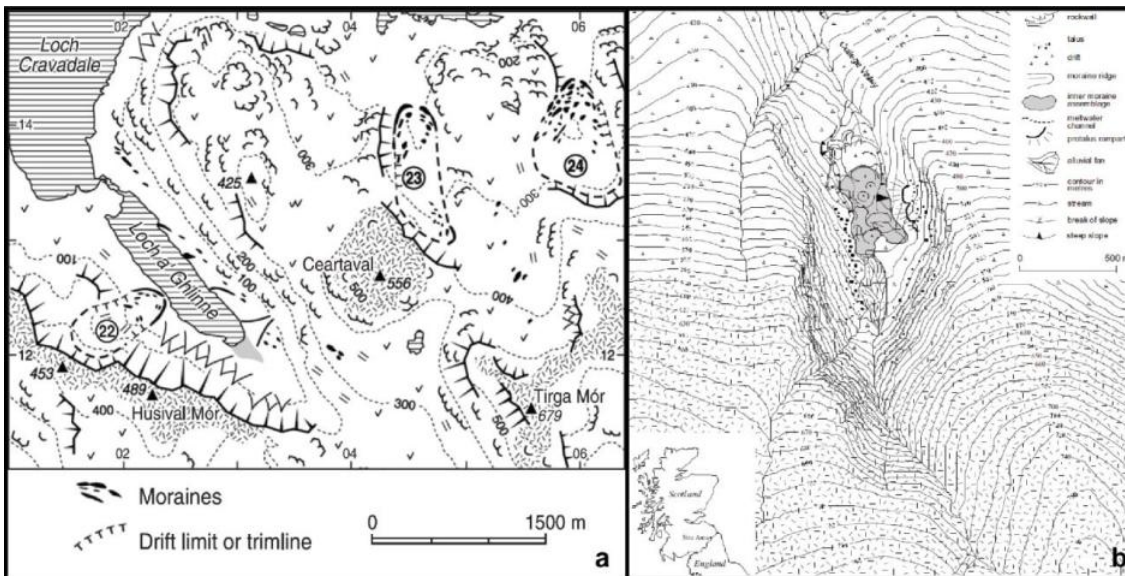
Figure 2.2 Examples of geomorphological maps used in the database a) The area around Ceartaval, Northern Harris. Reprinted from Quaternary Science Reviews, 26(25-28) Ballantyne, C.K., Loch Lomond Stadial glaciers in North Harris, Outer Hebrides, North-West Scotland: glacier reconstruction and palaeoclimatic implications., 3134-3149, Copyright (2007) with permission from Elsevier. b) The Bizzle Cirque, Cheviot Hills. Reprinted from Geografiska Annaler, Series A, 88(3), Harrison, S., Anderson, E. and Patel, D., The eastern margin of glaciation in the British Isles during the Younger Dryas: The Bizzle Cirque, southern Scotland., 199-207, Copyright (2006) with permission from John Wiley and Sons.

When maps did not provide such information, a different approach was required. OS maps (courtesy of Edina Digimap) were georeferenced and used to provide GCPs by matching prominent features on the OS map with the corresponding features on the geomorphological map. Features for GCPs were selected carefully, ideally with roads and buildings being used in preference to rivers, coastlines or water bodies, which might have changed position since the original mapping. However, the remote nature of many of the study locations often precluded use of anthropogenic features, leaving little choice for GCP selection. Our aim was to achieve RMS errors of 20 m or less but this was not always possible, particularly for older maps. Although great efforts were made to include all maps, wherever possible, some had to be excluded because they either did not contain sufficient information for their location to be derived, or they were at too coarse a scale to be useful. In total, some 216 individual geomorphological maps from 71 separate sources were found to be suitable.

### 2.2.3 Digitisation of layers

Once maps had been georeferenced in ArcMap, landforms were digitised as either lines or polygons into a series of layers organised by landform type, and saved as shapefiles (Figure 2.3). For the vast majority of evidence, this involved manual on-screen digitising of landforms. As the database is scale-independent, features were digitised at the maximum

resolution of the source maps. Where other researchers had produced shapefiles of geomorphology as part of their original mapping, these data were often provided by the authors for input into the database, and copied directly into the thematic layers. Other data were provided as unreferenced digital data (e.g. .tiff), which were then georeferenced and converted into shapefiles. Whilst the individual features are slightly pixelated and occasionally coalesced, this proved a very efficient way to rapidly ingest thousands of features from previous work into the GIS (Lukas and Lukas, 2006a, b).

As the accuracy of the database relies largely on the quality of the original maps, it is important that features can be traced back to their source. As such, the citation for each feature is included in its attribute table. This allows other researchers to examine the assumptions and justifications made by the original authors, and directly facilitates a closer examination of the existing literature. A full list of the references included in the map is available as supplementary material (in Appendix 1 of this thesis).

Hill-shaded renditions of the NEXTMap DSM were used to check the mapping and to provide topographic context for the geomorphology. This was useful for checking the quality of georeferencing and for identifying the correct position of landforms where original maps differed. However, the relatively small scale of LLS landforms is often below the 5 m resolution of the NEXTMap and so only larger features, such as end moraines or meltwater channels, were checked in this way. As the focus of this project was to compile the existing evidence rather than remap it, alterations (either redrawing or repositioning of landforms) have only been made in a limited number of cases where features were clearly incorrectly placed (e.g. where features which are clearly identifiable on the hill-shaded DSM are offset from their positions on the original maps). These alterations have been noted in the feature's attribute data.

In several cases, landforms originally attributed to ice limits that predate the LLS are also included in the original maps (Benn, 1990; Ballantyne, 2007a; Finlayson et al., 2011). In such cases, only features within or intersecting the authors' proposed glacier limits were included in the database. This approach relies heavily on the accuracy of the original glacial reconstructions to determine the age of the landforms which in some areas are speculative, often as a result of the paucity of absolute dates (Lukas, 2006). However, such restrictions were important to ensure that the LLS remained the focus of the project. Furthermore, as the Late Devensian Ice Sheet had previously overridden the areas later occupied by the LLS glaciers, it is possible that some locations within these limits may comprise palimpsest assemblages of younger landforms overlying older ones. Thus,

inclusion in this GIS does not necessarily mean that features were formed during the LLS but that they lay within the glacial limits and may have either survived beneath, or were modified by, the LLS ice.
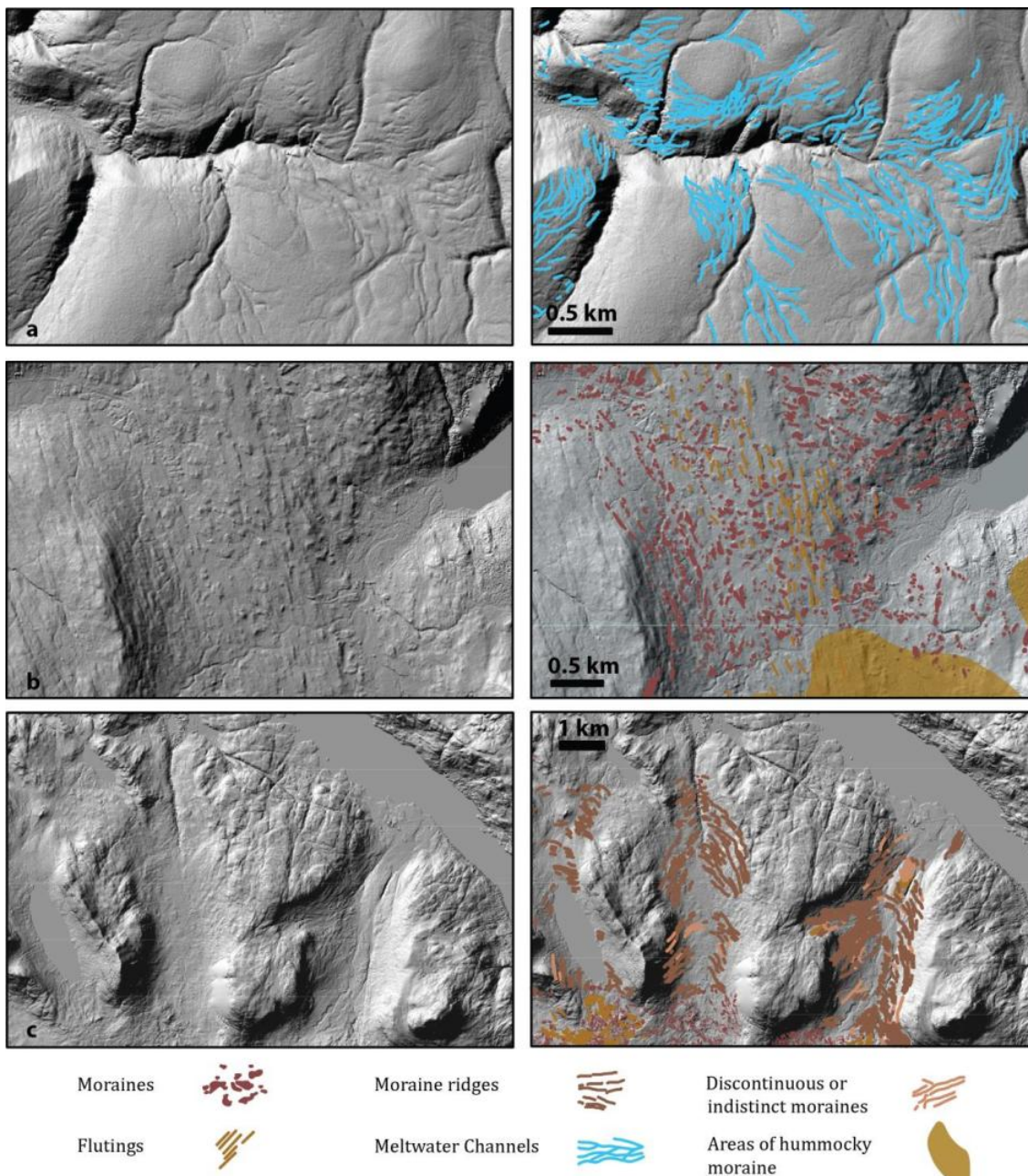


Figure 2.3 Examples of mapping from the GIS database showing a) Meltwater channels, Allt Gharbh Ghaig, Gaick, (after Sissons, 1974); b) Moraines, flutings and areas of hummocky moraine, Glen Sligachan, Isle of Skye (after Benn et al., 1992); c) Moraine ridges and discontinuous moraines, Strath Lungard and Glen Grudie, Wester Ross (Bennett, 1991; Bennett and Boulton, 1993b; McCormack, 2011). Legend has been modified slightly to aid identification of landforms. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

A practical problem encountered during digitisation was reconciling the differing scales and styles of mapping from the literature into one map. Maps range from coarse resolution sketches to highly detailed maps of individual sites of particular interest. Features which are classified as belonging to the same landform type are often represented differently between papers. The most widely reported LLS features were moraines, but numerous approaches to mapping these features have been adopted (Figure 2.3). Some researchers represent the shape and distribution of individual moraine mounds as polygons, whilst others record only ridge crests as line features or just general areas of 'hummocky moraine'. These different styles of mapping cannot be reconciled within a single layer in ArcMap, and so are split into separate layers accordingly ('Moraines (detail)', 'Moraine_Ridges' and 'Moraine_Hummocky_Area'). Generally, in areas where mapping of individual features overlapped areas of general 'hummocky moraine', only the detailed features were digitised to prevent the database becoming too cluttered. However, in instances where subsequent mapping shows an absence of features in locations where earlier publications reported evidence to be present, the older data have been retained around the newer evidence unless the presence of features is explicitly refuted in the latter paper or by the underlying NEXTMap data. Similar considerations were necessary where authors use different terminologies. Some areas originally mapped as 'hummocky drift' have been included in the 'Moraine_Hummocky_Area' layer when either more detailed mapping or the authors' descriptions indicated that moraines were found at the same location.

Striations and roches moutonneés have been combined into a single layer because some papers (Ballantyne, 2006, 2007a) represented these features with the same symbol, making it impossible to separate them into different layers. Similarly, the 'Drift_Limits' layer contains some features that were classified as 'Drift and/or boulder limits' in the original mapping, cf. Bendle and Glasser (2012). Although layers were informed by the most common way of representing features, inconsistencies arose very occasionally and, in order to prevent an impractical number of layers being created, features were redrawn to fit (e.g. polygon eskers were redrawn as lines).

### 2.2.4 Limitations

When synthesising over a century of research, gathered in an unsystematic way by a large number of researchers, a major consideration is the consistency and quality of the data collected. The original geomorphological maps were created using a variety of techniques (remote sensing, fieldwork or a combination of both) and, given the often subjective nature of landform identification, are based on the interpretations of individual workers.

Furthermore, advances in glaciological theory since publication of the original research may mean that some evidence is based on outdated ideas.

Secondly, it is conceivable, if not likely, that these LLS maximum extents were not reached synchronously (Golledge et al., 2008; Ballantyne, 2012). It is also possible that some features that were not formed by LLS glaciers may have been erroneously included. There is a relative scarcity of absolute dates on glacial landforms from the LLS in Britain and, consequently, some features that predate the LLS may have been included in this database if they were interpreted as LLS in age in the original study. Similarly, even when moraines have been inferred to predate the LLS, usually based on their more subdued appearance and greater size, the lack of absolute dates on these landforms means a LLS age cannot be ruled out. The glacial origin of some features has also been questioned, especially where moraines may appear similar in morphology to protalus ramparts or landslide blocks (Shakesby and Matthews, 1996). In cases of both uncertain age and genesis, this project assumes a LLS origin, especially if location appears consistent with the maximum extent of LLS glaciation, unless subsequent literature strongly refuted such an interpretation. Where these decisions have been made, this is indicated in the attribute data.

## 2.3 Discussion and implications

An initial inspection of the map (Figure 2.4) clearly shows the variable quality and unequal distribution of the source mapping. Some locations, for example the Isle of Skye, have been extensively studied (Ballantyne, 1989; Benn, 1992; Small et al., 2012) whilst others, including the southwest Scottish Highlands, have not been mapped to the same level of rigour and detail. Hence, this project identifies areas where further detailed study is required, guiding site selection for future research. It is hoped that this map and database will be the catalysts for the development of a complete coverage of mapping for all the former LLS glaciers in the British Isles, thereby facilitating a more holistic reconstruction of palaeoglaciology and palaeoclimate for the Younger Dryas in the region.

In some areas, there has been a general shift in our understanding of LLS glacier extent in that a more extensive ice cover has been recently proposed (Lukas and Bradwell, 2010; Brown et al., 2011, 2013; McDougall, 2013). The proposed limits of the West Highland Glacier Complex in Scotland have been refined, but are not too dissimilar to early reconstructions (Sissons, 1967a) and are supported by the results of numerical modelling (Golledge et al., 2008). However, at some locations, including both Sutherland in northwest Scotland and the English Lake District, more detailed remapping of geomorphology and an improved understanding of glaciological theory have been

employed to show that plateau icefields nourished extensive valley and cirque glacier systems (Rea et al., 1998; McDougall, 2001, 2013; Lukas and Lukas, 2006a, b; Lukas and Bradwell, 2010; Brown et al., 2011, 2013), thereby replacing earlier notions of more restricted, alpine styles of glaciation (Sissons, 1977a, 1980a).
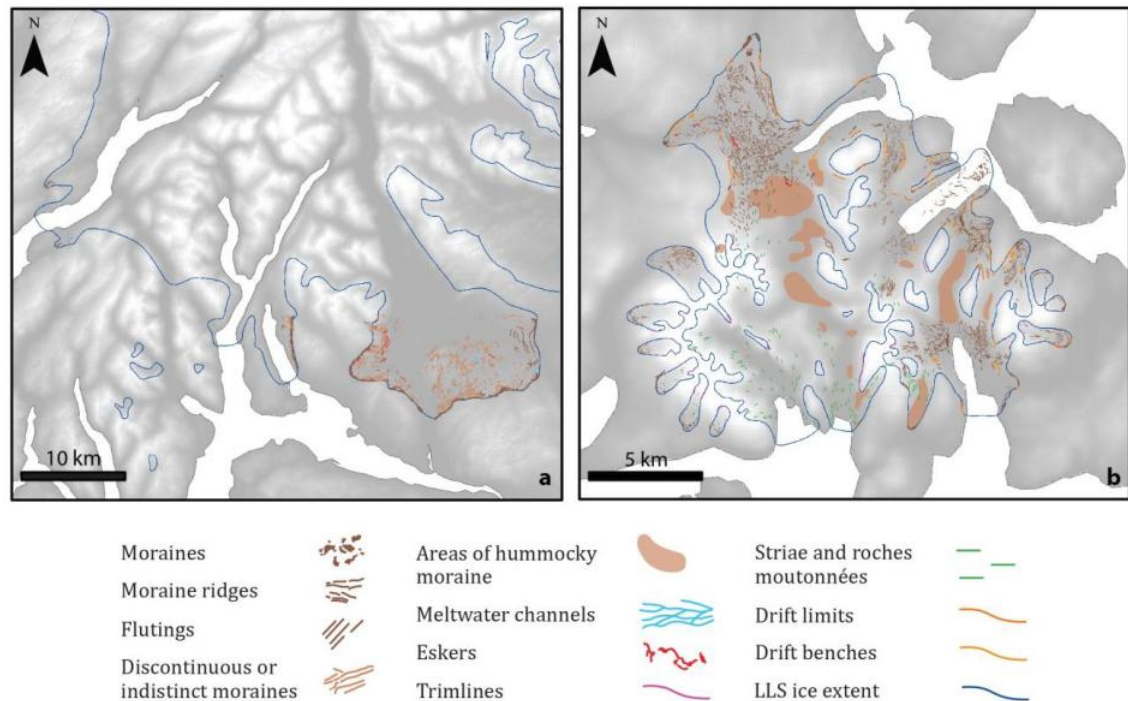


Figure 2.4 Examples of the variable quality and unequal distribution of geomorphological mapping a) The contrast between the detailed mapping of the Loch Lomond area in comparison to the surrounding southwest Scottish Highlands (after Sutherland, 1981; Rose, 1980, 1981). LLS ice extent adapted from Thompson (1972), Sutherland (1981) and Clark et al. (2004). b) Extensive and highly detailed mapping of the Isle of Skye (after Ballantyne, 1989; Benn, 1990, 1992, 1993; Benn et al., 1992; Dix and Duck, 2000). LLS ice extent adapted from Benn et al. (1992). GB Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA. UK outline reproduced from Ordnance Survey map data © Crown Copyright and Database Right 2014. Ordnance Survey (Digimap Licence).

Different landform assemblages appear to dominate different locations on the map, although some of this variation does result from diverse mapping styles in the literature. Many of the areas covered by the main mountain icefields are occupied by widespread 'hummocky moraine'. The moraines are sometimes organised into linear ridges, representing recessional moraines, or form a chaotic sea of mounds. These recessional moraines mark the retreat of the glaciers back to their source, for example, in the Uig Hills on the Isle of Lewis (Ballantyne, 2006). Elsewhere, such as Strath Mor and the upper Loch Ainort basin on the Isle of Skye, they appear to show that an earlier period of oscillatory retreat was followed by *in situ* stagnation of the ice (Benn et al., 1992). In the central Scottish Highlands, a rather different landform assemblage is apparent. The extensive plateau areas of the Gaick and Monadhliath show impressive sequences of meltwater

channels, suggesting the significant role of meltwater drainage in the behaviour of these ice masses (Sissons, 1974; Boston, 2012a).

Across Britain, but particularly at sites where environmental conditions were only marginally suitable for glacier development, the importance of glacier aspect is crucial. There is a clear predominance of glaciers with northerly to easterly aspects (Shakesby, 2007; Bendle and Glasser, 2012), which offer protection from insolation and are sites with a higher potential to receive nourishment of windblown snow from the west. This is particularly the case when plateaux or upland areas are found immediately southwest (upwind) of the glacier. At such locations, confidently identifying whether glaciers or snowbeds occupied certain locations has often challenged researchers (Shakesby and Matthews, 1993; Carr and Coleman, 2007a). The similarity in morphology between end moraines and protalus ramparts has required the use of other methods, such as glaciological modelling or sedimentological investigations, to differentiate between the two landform types (Shakesby and Matthews, 1993; Carr, 2001).

There are a number of ways in which the outputs from this project can be used. An obvious step is the compilation of a systematic evidence based reconstruction of the LLS ice masses. Although reconstructions of LLS ice extent in specific locations are common in the literature, this project aids comparison of the ice extent at larger scales, helping to identify trends in the distribution of ice. Furthermore, there is the potential to fine tune pre-existing reconstructions in areas that might have subsequently been remapped. The addition of absolute dates from the literature to this database will enable a more detailed understanding of the retreat dynamics of LLS ice. Although numerical modelling has previously been undertaken to determine the extent of LLS ice (Golledge et al., 2008), the outputs of this project provide a landform record which has the potential to aid in developing models to more accurately reconstruct glacier dynamics.

## 2.4 Conclusions
- This paper has compiled the glacial geomorphological evidence of LLS glaciation in Britain from the published literature into a GIS database that contains over 95,000 individual features.
- Where possible, original maps were collected and georeferenced into ArcMap and the evidence from these was digitised into a series of layers relating to landform type. Although great efforts were made to include all evidence, some maps did not provide sufficient geographic information to allow georeferencing and had to be

excluded. Similarly, it is possible that some sources have been overlooked during investigation of the literature.

- As the accuracy of the database relies heavily on that of the original publications, the source of each feature is included in its attribute data in the GIS.

- By overcoming the non-systematic nature of compilation of the evidence for LLS glaciation, it is hoped that this map and GIS database will form the catalyst for a range of potential investigations concerning the extent, retreat patterns and dynamics of LLS ice masses in Britain.

## 2.5 Software

Geomorphological maps were georeferenced and landforms were digitised in ESRI ArcGIS. Grids were added to some maps in Adobe Photoshop and the final map was produced in Adobe Illustrator.

## 2.6 Data

The geomorphology shapefiles created for this project are designed to be used in ArcMap. The files were released as supplementary material alongside this paper at the time of publication.

# Chapter 3

# The glacial geomorphology of the Loch Lomond Stadial in Britain: a review

## Abstract

This paper systematically reviews the glacial geomorphological evidence of the Loch Lomond Stadial (Younger Dryas) glaciation in Britain (12.9-11.7 ka). The geomorphology of sub-regions within Scotland, England and Wales is assessed, providing the most comprehensive synthesis of this evidence to date. The contrasting nature of the evidence at the local scale is reviewed and conceptual themes common to multiple sub-regions are examined. Advancements in glaciological theory, mapping technologies, numerical modelling and dating have been applied unevenly to localities across Britain, inhibiting a holistic understanding of the extent and dynamics of the Loch Lomond Stadial glaciation at a regional scale. The quantity and quality of evidence is highly uneven, leading to uncertainties regarding the extent of glaciation and inhibiting detailed analysis of ice dynamics and chronology. Robust dates are relatively scarce, making it difficult to confidently identify the limits of Loch Lomond Stadial glaciers and assess their synchroneity. Numerical models have allowed the glacier-climate relationships of the Loch Lomond Stadial to be assessed but have, thus far, been unable to incorporate local conditions which influenced glaciation. Recommendations for future research are made that will allow refined reconstructions of the Loch Lomond Stadial in Britain and contribute to a more comprehensive understanding of glacier-climate interactions during the Younger Dryas.

## 3.1 Introduction

Following the Last Glacial Maximum (LGM) between 26.5 and 19 ka (Clark et al., 2009) the Northern Hemisphere experienced an abrupt climatic downturn during a period known as the Younger Dryas (YD). This event, generally thought to have occurred between 12.9 and 11.7 ka (Rasmussen et al., 2006), is clearly marked by a decrease in the $\delta^{18}O$ isotope signal in Greenland ice cores, suggestive of up to 9°C cooling during this period (Alley, 2000; Carlson, 2013). A reduction in temperatures and in precipitation was experienced across most of the Northern Hemisphere, including North Africa and Asia (de Menocal et al., 2000; Nakagawa et al., 2003; Genty et al., 2006), but particularly in the North Atlantic and surrounding regions (Benway et al., 2010). Conversely, temperatures in the Southern Hemisphere warmed during this period, particularly at high latitudes, causing a global net cooling of approximately 0.6°C (Shakun and Carlson, 2010).

Although these effects were not experienced synchronously across all regions, the abrupt nature of the transition into cold temperatures during a period of increasing insolation supports the long-held assumption that the Younger Dryas was caused by disruption of the Atlantic meridional overturning circulation (AMOC), which moderates the region's climate (Carlson, 2010; Golledge, 2010). It is generally accepted that this disruption resulted from an influx of freshwater into the North Atlantic (Broecker, 2006; Carlson and Clark, 2012) although there has been much debate as to the causes of this, ranging from: glacial outburst floods of meltwater northwards along the Mackenzie River system (Murton et al., 2010); to an extra-terrestrial impact causing destabilisation of the Laurentide Ice Sheet (Firestone et al., 2007); to a catastrophic break up and expulsion of palaeocrystic sea ice from the Arctic Ocean (Bradley and England, 2008). More recently the hypothesis that the retreat of the Laurentide Ice Sheet allowed meltwater to drain eastwards along the St Lawrence River into the North Atlantic (Johnson and McClure, 1976; Teller et al., 2002) has been proposed (Carlson, 2013). In a recent review, Carlson and Clark (2012) concluded that this eastwards routing of runoff from the Laurentide Ice Sheet into the St Lawrence River was the most likely cause of the Younger Dryas cooling and argued that hypotheses against this conflicted with constraints on meltwater discharge, involved incorrect interpretation of geologic data or inaccurate ice sheet models.

The resulting climatic deterioration caused the expansion of circum-Atlantic ice masses. In North America, the southern margin of the Laurentide Ice Sheet readvanced to beyond the southern shore of Lake Superior, forming a 1000 km long near-continuous moraine system between Duluth and North Bay, Ontario (Lowell et al., 1999). Further east the Saint

Narcisse moraine complex marks the southeastern limits of the ice between the Ottawa and Saguenay Rivers (Occhietti et al., 2004; Occhietti, 2007), and small glaciers in Nova Scotia were reactivated (Mott and Stea, 1993). Likewise, the Icelandic Ice Sheet readvanced towards coastal sites, truncating raised shorelines and depositing moraines (Ingólfsson et al., 2010). Similarly, the Scandinavian Ice Sheet readvanced, extending almost to the Norwegian coastline, covering all but the southern regions of Sweden and Finland, forming a continuous series of moraines around the ice sheet (Andersen et al., 1995; Mangerud et al., 2011).

Although it is difficult to match the timing of this event across global records for this period, Great Britain experienced a fall in mean July air temperatures from 11°C at the end of the interstadial, to 7.5°C at the beginning of the YD (Brooks and Birks, 2000), with a maximum drop of 10°C in mean annual temperatures in Scotland (Golledge et al., 2008). This caused glaciers to regrow to form an extensive icefield along the length of the western Highlands, flanked by numerous satellite icefields, ice caps, valley and cirque/niche glaciers in surrounding upland areas (Golledge, 2010). The existence of a phase of valley glaciation following the retreat of the British-Irish Ice Sheet was recognised during the 19th Century (Forbes, 1846; Jolly, 1868; Geikie, 1878), but most early accounts were purely descriptive. This period of readvance became known as the Loch Lomond Stadial (LLS), based on Simpson's (1933) identification of the limits of this phase of glaciation in the Lomond Valley, southwest Scotland. Subsequently, Charlesworth (1955) reconstructed ice marginal positions, identifying a 'Moraine Glaciation' stage during the retreat of the last British-Irish Ice Sheet, the extent of which matches more recently proposed reconstructions of the LLS glaciers. However, no field mapping was included in this study and it was only in the 1970s that systematic mapping of the glacial landforms attributed to this period was conducted, most notably by J.B. Sissons (e.g. 1972, 1974, 1977a, b).

Much of the published output on LLS glaciation in Britain has consisted of maps of the geomorphology of specific locations, leading to a fragmented and spatially inconsistent body of research compiled over many years by numerous researchers. To reconcile these various databases, Bickerdike et al. (2016) recently compiled the evidence for the glacial geomorphology of the Loch Lomond Stadial in Britain, using the published literature and collating the relevant information in a GIS database and accompanying glacial map (Bickerdike et al., 2016); this is a similar approach to that taken for the compilation of the evidence associated with the last British Ice Sheet by Clark et al. (2004) and Evans et al. (2005). Whilst previous studies have detailed the maximum extent of LLS glaciers from

published reconstructions (Golledge, 2010), Bickerdike et al. (2016) were the first to detail the geomorphological record of the LLS glaciers at this scale and make all evidence available in a GIS format. Evidence within the GIS is organised into a series of thematic layers, each containing a different landform type, and comprising moraines (including individual moraine mounds and ridges, fluted moraines and areas of hummocky moraine), drift and boulder limits, drift benches, meltwater channels, eskers, periglacial trimlines, striations and roches moutonneés (Bickerdike et al., 2016).

The compilation of the geomorphological evidence based upon an extensive search of the published literature has enabled a systematic evaluation of a substantial LLS palaeogeographic database, identifying issues associated with both areal coverage and mapping quality. This paper concentrates on critically reviewing the evidence, assessing different regions in turn; specifically Scotland, England and Wales are subdivided into sub-regions, as detailed in Figure 3.1. An example of the complete map, as available in Bickerdike et al. (2016), is shown in Figure 3.2. The final section of this paper draws together conceptual themes common to all or several of the sub-regions, including the quality and quantity of evidence, uncertainties regarding the age or glacial origin of landforms, changing views on the proposed style of glaciation, the findings of numerical modelling experiments and the approximate extent of LLS glaciation.

The fragmented and spatially inconsistent nature of the previous mapping is clear; whilst high resolution geomorphological maps exist for some localities, others have been the focus of little to no research and our understanding of the LLS glaciation in these areas is poor. Advances in mapping technologies and glaciological theory have aided in better constraining the extent of LLS glaciers and there has been a general shift away from alpine to icefield styles of glaciation (e.g. Rea et al., 1998; McDougall, 1998, 2001, 2013; Lukas and Bradwell, 2010; Brown et al., 2011, 2013). Thermomechanical modelling forced by a locally scaled Greenland Ice Core Project (GRIP) temperature series (Golledge et al., 2008), which has been able to closely replicate the extent of LLS glaciation inferred from empirical limits, further supports this more extensive style of glaciation but deviates from the limits at local scale at sites where factors such as windblown snow were probably critical to glacier development. Furthermore, the relative paucity of LLS dated features has led to uncertainty when differentiating between LLS and older landforms, potentially leading to incorrect age assignments to landforms, which has repercussions when these landforms are then used to establish the extent of glaciation. In some locations the limits of LLS glaciation have been reconstructed with confidence, based on a combination of clear geomorphological evidence and robust absolute dating. However, this is not the case

for many areas, particularly where the evidence of LLS glaciation may be patchy or is equivocal and, in such cases, caution should be exercised when making inferences from the proposed ice limits. The implications of these limitations for the current understanding of LLS glaciation are highlighted in this paper and recommendations for future research to remedy such shortfalls these are proposed. Specific local scale problems, where alternative interpretations have been proposed for the geomorphology and its age, are outlined and a justification provided for the published interpretation that is preferred for inclusion in the GIS. The regional compilation within the GIS database is then employed to reconstruct the extent of the LLS glaciation, based upon the literature available up to the census end date of June, 2014.
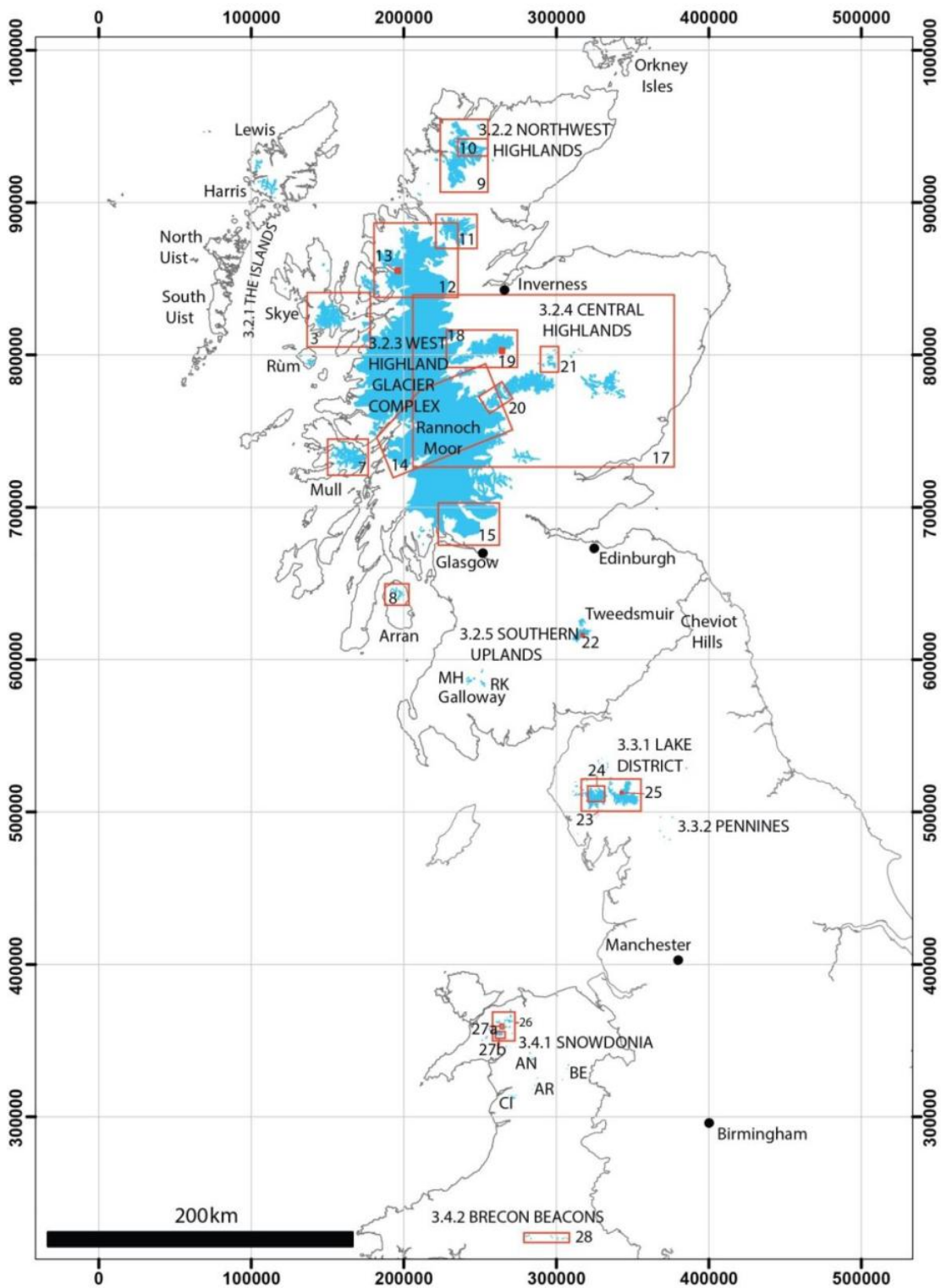
Figure 3.1 Location map of regions where LLS glacial geomorphology is present, as discussed in this paper, showing the approximate extent of the ice. The location of figures is is indicated in red or is indicated later in the paper. MH: Merrick Hills, RK: Rhinns of Kells AN: Arenig Mountains, AR: Aran Mountains, BE: Berwyn Mountains, CI: Cadair Idris. Britain coastline reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence).
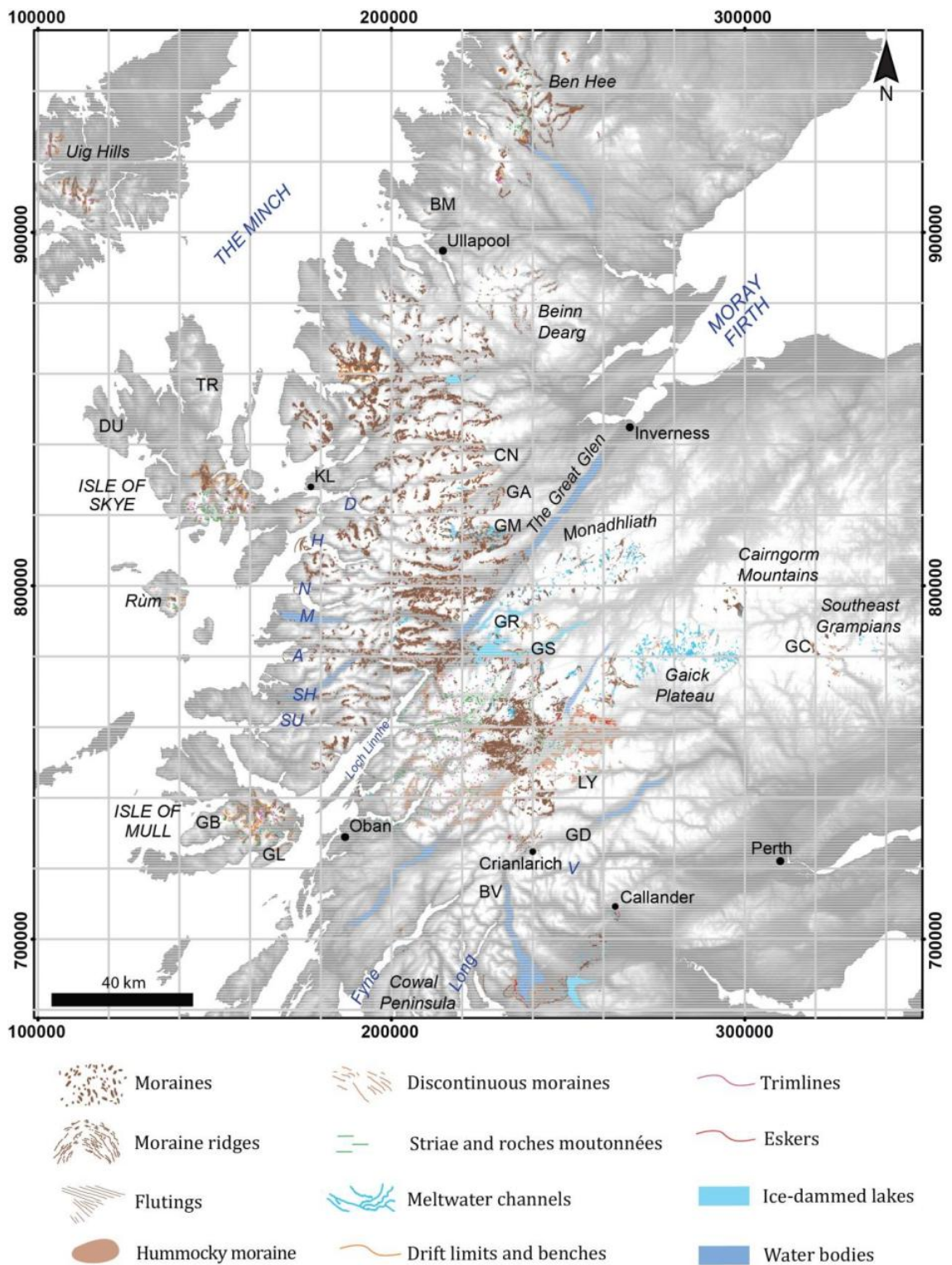
Figure 3.2 LLS glacial geomorphology of the Highlands and islands of Scotland, adapted from Bickerdike et al. (2016). The map shows the distribution of landforms relating to the West Highland Glacier Complex which covered most of the Western Grampian Mountains and was flanked by satellite icefields in neighbouring upland areas. BM: Ben Mor Coigach, TR: Trotternish Peninsula, DU: Duirinish Peninsula, KL: Kyle of Lochalsh, D: Loch Duich, H: Loch Hourn, N: Loch Nevis, M: Loch Morar, A: Loch Ailort, SH: Loch Shiel, SU: Loch Sunart, GB: Gribun, GL: Glen Libidil, CN: Glen Cannich, GA: Glen Affric, GM: Glen Moriston, GR: Glen Roy, GS: Glen Spean, LY: Glen Lyon, GD: Glen Docharty, V: Loch Voil, BV: Ben Vorlich, GC: Glen Clova. British coastline and present-day waterbodies reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence). GB SRTM Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA.

## 3.2 Geomorphological evidence in Scotland

### 3.2.1 The Islands

A range of glaciation styles are represented by the geomorphological evidence preserved on the islands around Scotland's coastline. Substantial icefields developed in the mountainous areas of the isles of Skye and Mull (Ballantyne, 1989, 2002a; Benn et al., 1992), which lie in closest proximity to the main West Highland Glacier Complex. In contrast, more marginal islands, such as the Isle of Arran (Ballantyne, 2007b) and the Outer Hebrides (Ballantyne, 2006, 2007a) supported only valley glaciers whilst glaciation on the Orkney Isles was restricted to favourably situated cirques (Ballantyne et al., 2007).

### 3.2.1.1 Orkney

Hoy, in the Orkney Isles, represents the most northerly site of glaciation in the British Isles during the LLS but the extent of ice was restricted to two small cirques, Enegars and Dwarfie Hamars, in the upland area in the northwest of the island (Hall, 1996; Ballantyne et al., 2007). At Enegars a large terminal moraine, 300 m wide and 8 m high, comprising poorly sorted, coarse gravels (Sutherland, 1996a), marks the maximum extent of a small glacier which developed in the cirque. Similarly, a complex of moraine ridges and hummocks between 3 and 6 m in height are found at Dwarfie Hamars. The large volume of sediment comprising this 250 m wide ridge probably resulted from a combination of slow glacial retreat that reworked an abundant supply of either pre-existing debris or material supplied by rockfalls (Ballantyne et al., 2007). It has been suggested that both glaciers retreated slowly and actively from their maximum extents before potentially undergoing later uninterrupted retreat, as indicated by large outer moraines, inner recessional ridges and an absence of ridges in the upper cirques (Ballantyne et al., 2007). Cosmogenic isotope dating of boulders on these moraines gave ages between 11.9±1.2 ka and 10.4±1.7 ka at Enegars and 12.4±1.5 ka and 11.6±1.6 ka at Dwarfie Hamars, respectively (Ballantyne et al., 2007), strongly supporting a LLS age in accord with the significantly more developed scree slopes found only outside the proposed limits (Sutherland, 1996b).

Whilst other authors previously argued for more extensive glaciation of Orkney (Charlesworth, 1955; Rae, 1976), it has subsequently been argued that the thick drift which is found at other potential sites was deposited during ice sheet glaciation and that the presence of cirque glaciers at Enegars and Dwarfie Hamars results from their favourable topographic settings downwind of large potential snow-contributing upland areas (Sutherland, 1996b; Ballantyne et al., 2007). Based on this evidence, not only was ice probably restricted to these sites on Hoy but it also seems implausible that any other

locations in the Orkney Isles would have been glaciated, given the low, subdued topography of the rest of the archipelago.

### 3.2.1.2 Outer Hebrides

Much of the generally low-lying rugged landscape of the Outer Hebrides was last glaciated by a locally nourished ice cap during the LGM (Flinn, 1978; von Weymarn, 1979; Stone and Ballantyne, 2006). The exceptions are the mountainous areas of western Lewis and Harris where Geikie (1878) identified moraines formed during a subsequent phase of valley glaciation following LGM deglaciation. Although Charlesworth (1955) proposed that extensive valley glaciers had drained from the Harris mountains, covering much of the island, and from the upland areas of western Lewis and South Uist, more recent geomorphological mapping suggests a much more restricted ice cover during the LLS (Ballantyne, 2006, 2007a).

Accounts of LLS glacial geomorphology in the Uig Hills, western Lewis (Peacock, 1984; Ballantyne, 2006) have been in general agreement, with Ballantyne (2006) reconstructing four valley glaciers from the geomorphological evidence in the area. The Raonasgail and Tamanisdale glaciers flowed north and south, respectively, through the trough between two parallel mountain ridges, leaving an almost complete coverage of nested recessional hummocky moraines, the limits of which were recorded by von Weymarn (1979), although he attributed them to ice stagnation of uncertain age. Two further glaciers (the Dibadale and Suainaval glaciers) drained from the eastern ridge, though moraines in these valleys are concentrated around the glacier limits and ice-scoured bedrock and glacially-transported boulders dominate the valley floors. There is some uncertainty as to whether the Suainaval glacier was connected to the rest of the ice mass, in a similar fashion to ice in upper Glen Dibadale. Peacock (1984) believed that the source area was to the west of the mountain ridge, whereas Ballantyne (2006) put the source in the cols on the north and south flanks of Tahaval based on trimlines and the alignment of lateral moraines in Glen Raonasgail. This evidence suggests that the ice was insufficiently thick to allow the Raonasgail glacier to override the mountain ridge and coalesce with the Suainaval glacier.

A greater extent of LLS glacier coverage has been proposed for the hills of northern Harris. Three glaciers occupied the through-valleys of Glen Ulladale-Glen Chliostair, Glen Meavaig and Glen Langadale, with a further seven cirque glaciers in the surrounding area (Ballantyne, 2007a). Extensive spreads of hummocky moraine are present in both the trunk valleys and smaller tributaries and are arranged into belts of abundant hummocks and intervening moraine-free areas, suggesting that the retreat of the glaciers was episodic with periods of stability or slow retreat (Ballantyne, 2007a; cf. incremental

stagnation of Eyles, 1979). Whilst the presence of moraines that extend to the glaciers' source area accords with the evidence found in the Uig Hills (Ballantyne, 2006), the organisation of moraines in northern Harris into hummocky moraine bands was not reported for that area. The termini of the glaciers on northern Harris are marked either by abrupt limits to the outermost hummocky moraine band (as at the northern limits of the Ulladale and Langadale glaciers) or by clear end moraines, sometimes arranged in chains or multiple ridges (as formed by the Meavaig glacier and the cirque glaciers at An Coire, Gleann Dubh and Gleann Bhearrary).

Despite good agreement with earlier mapping for some areas (including Glen Meavaig, north Glen Ulladale and An Coire), features in Glen Dibidale and Glen Laxadale of the Sgaoth Aird hills have recently been argued to predate LLS glaciation (Ballantyne, 2007a). Similarly, limits at Beinn Losgaintir and on Loch Sealg's northern shore (von Weymarn, 1979) are at elevations too low to reconcile with the higher termini of the LLS glaciers in the hills.

There is an absence of absolute dates on proposed LLS features in the Outer Hebrides, but it has been possible to use the geomorphological evidence to infer a LLS age for the phase of valley glaciation (Ballantyne, 2006, 2007a). For example, the orientation of ice directional features in the valleys contrasts with those indicative of LGM ice flow beyond the proposed limits, suggesting that the valleys were glaciated subsequent to the LGM in a separate readvance event. Furthermore, 'hummocky moraine', which is generally associated with LLS glaciation, is present within these limits and periglacial phenomena, such as frost-weathered bedrock and solifluction lobes, are restricted to areas outside of the hummocky moraine. This suggests that the areas of moraines were protected from exposure to LLS cold conditions by the presence of glacier ice. Cosmogenic isotope dates from the high cols suggest that ice sheet deglaciation persisted until 14.5-14.1 ka (Stone and Ballantyne, 2006), suggesting that the LLS represents the only subsequent cold period during which this readvance could have occurred.

### 3.2.1.3 Isle of Skye

The Isle of Skye (Figure 3.3) contains some of the finest examples of LLS glacial geomorphology in Britain and, consequently, mapping of this area has been comprehensive. During the LGM, ice from the mainland was deflected to the north and south around an independent ice dome centred on the Cuillin mountains, which dominated the centre of the island (Ballantyne, 1989). Although the existence of a later phase of local glaciation during the LLS has long been recognised (Forbes, 1846; Harker, 1901; Charlesworth, 1955), the first detailed geomorphological mapping of the area was

conducted by Sissons (1977b) who reconstructed nine cirque and valley glaciers in the Cuillins and four in the Eastern Red Hills (Figure 3.4a). However, more detailed remapping of the geomorphology prompted reconstruction of a much more extensive icefield in the Cuillins and surrounding areas (Ballantyne, 1989; Benn, 1992; Benn et al., 1992) (Figure 3.4b).
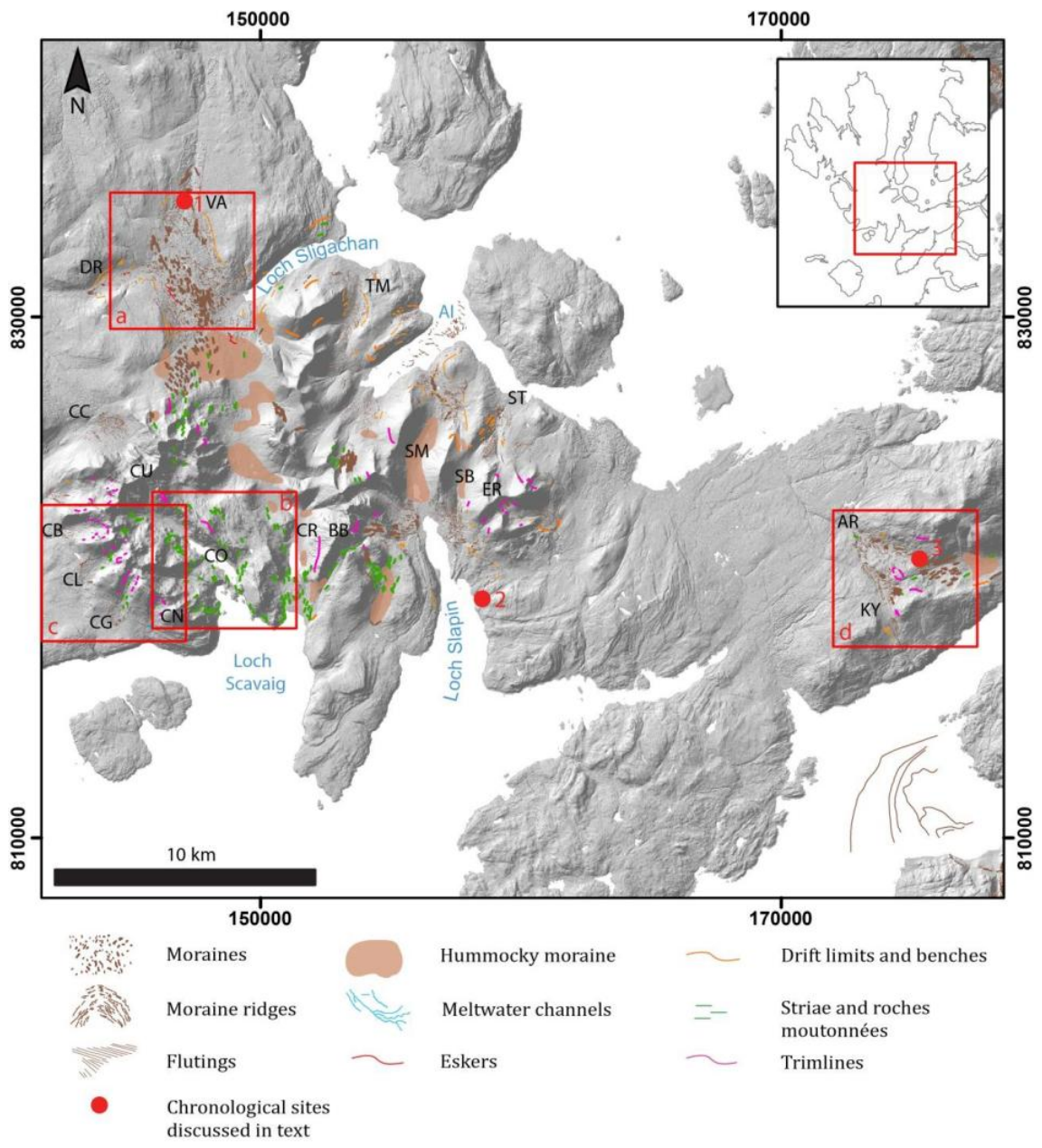


Figure 3.3 Geomorphological mapping compiled for the glacial map for the Isle of Skye (after Ballantyne, 1989; Benn, 1990, 1992, 1993; Benn et al., 1992; Dix and Duck, 2000). Location boxes included for Figure 3.5. VA: Varragill, DR: Drynoch, TM: Gleann Torra-mhichaig, AI: Loch Ainort, CC: Coire na Creiche, CU: Cuillin Hills, CB: Coire na Banachdich, CL: Coire Lagan, CG: Coire a' Ghrunnda, CN: Coire nan Laogh, CO: Loch Coruisk, CR: Creitheach, BB: Blà Bheinn, SM: Strath Mor, SB: Strath Beag, ER: Eastern Red Hills, AR: Glen Arroch, KY: Kyleakin Hills. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.
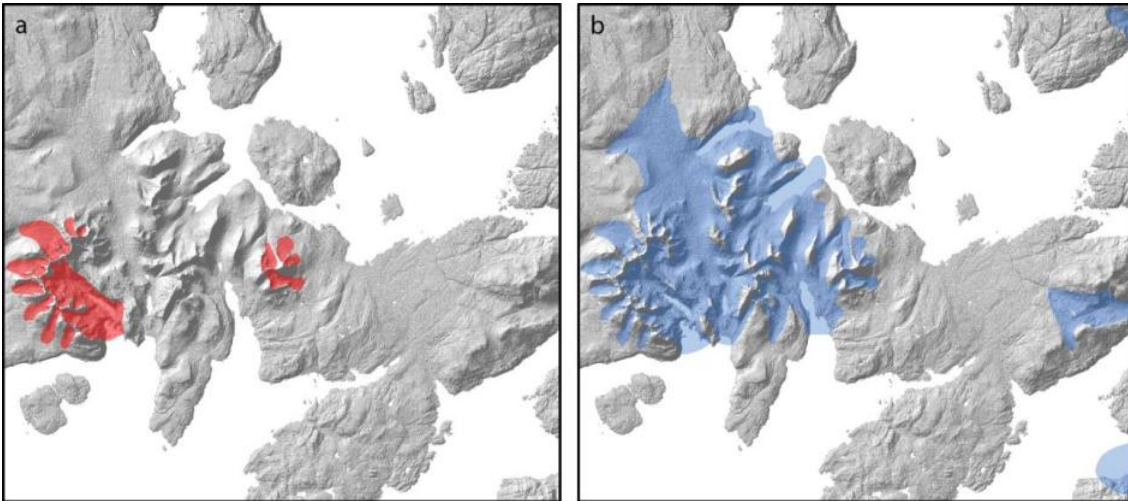
Figure 3.4 The extent of LLS glaciation on the Isle of Skye a) after Sissons (1977b) and b) as inferred from the subsequent geomorphological mapping, after Ballantyne, 1989; Benn, 1990, 1992, 1993; Benn et al., 1992; Dix and Duck, 2000. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The LLS icefield occupied the mountainous area of the Cuillins and Red Hills, covering an area of approximately 150 km$^2$, and showed pronounced north-south asymmetry, resulting partly from the low gradient piedmont lobe which extended northwards from the Cuillins down Glens Sligachan, Drynoch and Varragill (Ballantyne, 1989) (Figure 3.3). The hummocky moraines which mark the extent of these northern outlet glaciers are arranged into a complex pattern of transverse recessional moraines, chaotic mounds and flutings (Benn et al., 1992) (Figure 3.5a). Sedimentological investigations have determined that this area is underlain by a compact, sheared till that deformed under low shear stresses beneath the LLS glaciers and hence, resulted in a low gradient ice lobe (Ballantyne, 1989; Benn, 1992). These features were deemed by Sissons (1977b) to predate the LLS but pollen stratigraphies from within this area were found to date from the Early Flandrian onwards (Figure 3.3, chronological site 1), supporting the notion that these features were formed during the LLS (Benn et al., 1992).

The extent of glaciers which drained the eastern sector of the icefield is also demarcated by hummocky moraines. Examples of these are found in Gleann Torra-Mhichaig where cross-valley pairs of sub-parallel moraine ridges on the valley sides suggest that this glacier underwent active retreat, although conical mounds on the valley floor further indicate that some sediment-covered areas of ice decayed *in situ* (Benn, 1990; Benn et al., 1992). Similarly, whilst intermittent recessional moraines are present along the sides and floor of Loch Ainort, these are replaced by isolated hummocks at the head of the loch and sedimentary exposures suggest they are ice stagnation moraines (Benn et al., 1992).

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Bathymetric surveys have revealed that a sequence of moraines is present at the opening of Loch Ainort (Figure 3.3), suggesting that the glacier was approximately 800 m longer than had previously been inferred from the onshore evidence (Dix and Duck, 2000). The sequence comprises an outer belt of De Geer moraines followed by an inner area of more hummocky relief with sporadic moraines, indicating that deglaciation involved an initial phase of active retreat followed by more uninterrupted retreat with only occasional stillstands or readvances (Dix and Duck, 2000). Similarly, the evidence in nearby Strath Mor suggests a two-phase retreat, the lower valley having clear transverse belts of larger moraine ridges and intervening smaller mounds, whilst the upper valley is occupied by a thick drift sheet comprising irregular hummocky topography and dissected kame terraces (Benn et al., 1992) indicative of ice stagnation.

This pattern of deglaciation is further replicated for the Slapin glacier, which drained the Eastern Red Hills and the Blà Bheinn mountain area (Figure 3.3). The limit of this lobe is marked by a series of prominent end moraines on the eastern loch shore, dated to an average age of 11.5±0.7 ka (Small et al., 2012) (Figure 3.3, chronological site 2), which continue to the western shore as a boulder ridge on the loch floor (Benn, 1990). Glacial features are largely absent from the area immediately inside the moraines, which has been modified for settlement. However, from approximately 1.5 km inside the limit, well-developed, sharp-crested moraine ridges occupy the slopes of Strath Beag (Figure 3.6a) whereas the lower flatter ground is dominated by recessional moraines formed by chains of large mounds and ridges (Benn et al., 1992). These features continue up-valley until transitioning into undulating drift, supporting the two-phase retreat observed elsewhere on the island. The depositional evidence of the Slapin glacier contrasts with that mapped for the southwards flowing Coruisk and Creitheach glaciers, which consists primarily of ice-moulded bedrock and striae with a general absence of drift cover (Figure 3.5b). In some locations, such as the northwest shore of Loch Scavaig (Figure 3.3), reconstruction of the margins of these glaciers has only been possible by identifying where striations aligned with LLS ice flow overprint older erosional features indicative of LGM ice flow (Ballantyne, 1989).
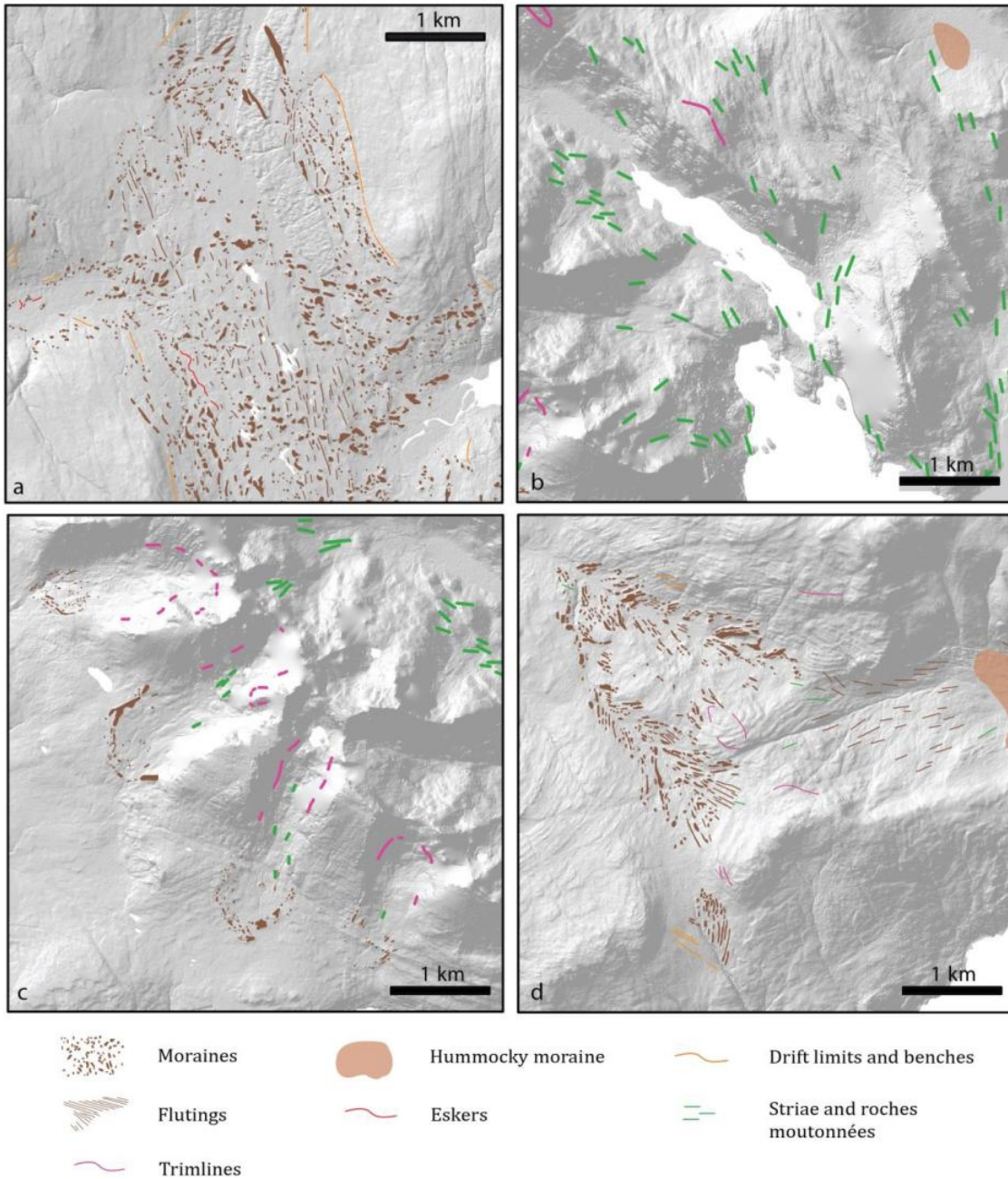
Figure 3.5 Examples of the varied LLS glacial gemorphology of the Isle of Skye. a) Hummocky moraine in Glen Sligachan, comprising transverse recessional moraines, flutings and chaotically arranged mounds (after Benn et al., 1992). b) Erosional evidence, including striae, roches moutonnées and trimlines at Loch Coruisk (after Ballantyne, 1989). c) Terminal moraines and trimlines, indicating the extent of the Cuillin cirque glaciers of Coire na Banachdich, Coire Lagan, Coire a' Ghrunnda and Coire nan Laogh (after Benn, 1990 and Benn, 1992). d) Recessional moraines marking the extent of the small Kyleakin icefield in Glen Arroch (after Ballantyne, 1989; Benn et al., 1992). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre

Figure 3.6 Examples of different styles of LLS moraines. a) Recessional hummocky moraines in Strath Beag, Isle of Skye, Scotland. b) Drift limit on Conic Hill on the eastern shore of Loch Lomond, Scotland. c) Recessional moraines associated with the terminus of the Lomond Lobe at Gartness, Scotland. The outermost feature changes from a flat-topped feature comprising laminated silts and glaciofluvial sands and gravels in the south, formed subaqueously in a proglacial lake, into a sharp-crested ridge of till in the north. d) Moraines at Stake Pass in the Lake District, England, indicating active retreat of the Langstrath glacier up onto the plateau above. e) Large lateral moraine on the northern shore of Llyn Llydaw, Snowdonia, Wales (feature Y on Figure 26b). This feature was formed during a phase of re-stabilisation of the retreating Llydaw glacier. f) The Fan Hir ridge, Brecon Beacons, Wales. This large, linear feature, which runs parallel to the escarpment behind it, has been identified as both a LLS aged moraine and a protalus rampart. Photograph A from H.L. Bickerdike and B-F from D.J.A. Evans.

The main Cuillin icefield was flanked by ten cirque glaciers, seven along the western flanks of the Cuillin ridge and three in the Eastern Red Hills (Figure 3.3). The limits of the majority of these glaciers are characterised by chains of moraine ridges and boulders arranged in a belt at the glacier terminus, with the upper cirques dominated by ice-moulded bedrock and striations (Sissons, 1977b; Ballantyne, 1989; Benn, 1990) (Figure 3.5c). The extent of the glaciers in Coire Lagan and Coire na Banachdich has been disputed;

Sissons (1977b) reconstructed the limits of these glaciers from lateral moraines, boulder limits and the extent of hummocky moraine, but Ballantyne (1989) extended these limits to incorporate flutes ('fluted moraine'), boulder spreads and hummocky drift beyond the Sissons (1977b) limits. Benn (1990) argued for a more restricted extent, closer to Sissons' (1977b) original reconstruction, suggesting that the more subdued outer ridges at Coire na Banachdich predated the LLS and that Ballantyne's reconstruction required ice to flow up onto higher ground at Coire Lagan and that the glacier did not extend this far. The arrangement of ridges in both the Cuillin and Eastern Red Hills cirques suggests that retreat, at least initially, was active. In some cirques, such as Coire Lagan, as few as two retreat positions, both near the terminus, have been identified (Benn, 1990). In contrast, in Coire na Creiche (Benn, 1993), recessional moraines cover much of the valley floor and sides for 2 km inside the margin and record up to 10 retreat positions (Benn, 1990). The discrepancy between these patterns suggests a complex and varied style of deglaciation across the island.

Aside from the main Cuillin icefield, three peripheral areas of Skye nourished glaciers during the LLS. Ballantyne (1989) reconstructed three valley glaciers in the Kyleakin Hills (Figure 3.3) but found no evidence for the icefield and eight valley glaciers proposed previously by Charlesworth (1955). Of these valleys, especially clear evidence is found in Glen Arroch where recessional moraines, form 2-3 m high cross-valley pairs of chevron-shaped bands pointing down-valley (Benn, 1992; Benn et al., 1992) (Figure 3.5d). A sediment core from near the head of the glen (Figure 3.3, chronological site 3) revealed a transition from basal minerogenic to organic sediments containing a complete early Flandrian pollen succession, indicating that Glen Arroch was completely ice-free by the end of the LLS (Benn et al., 1992).

LLS glaciers have also been proposed to have formed along the Trotternish Escarpment of northern Skye (Figure 3.2). Charlesworth (1955) and Anderson and Dunham (1966) both proposed that substantial glaciers had developed along the escarpment during readvances after ice sheet deglaciation but neither presented field evidence. Subsequently, much less extensive glaciation has been inferred by Ballantyne (1990) who found evidence for just two small cirque glaciers in the lee of the escarpment. The end moraines which mark these glaciers are relatively subdued in their appearance and in places are challenging to identify amongst the thick peat and landslide blocks of the surrounding area (Charlesworth, 1955; Ballantyne, 1990). A further small glacier has been reconstructed in Glen Osdale on the Duirinish Peninsula (Figure 3.2) as inferred from an arcuate band of drift hummocks which are continued up-valley by drift limits (Ballantyne and Benn, 1994).

Recessional moraines located inside the outermost ridge suggest that, like the other cirque glaciers on Skye, the initial retreat of this glacier was active.

In summary, the wealth of evidence on Skye, strongly refutes the Sissons (1977b) model of restricted glaciation and supports the more extensive icefield configuration proposed by Ballantyne (1989), Benn (1990, 1992, 1993), Benn et al (1992) and Dix and Duck (2000). Dated evidence from Loch Slapin and the pollen stratigraphy from Kyleakin, indicating that these areas were glaciated during the LLS, is particularly persuasive as both sites fall well beyond the Sissons (1977b) limits. If this dating control can be extrapolated to other undated valleys on the island, the same style of fresh moraines would appear to represent the LLS limit. However, the presence of offshore moraines in Loch Ainort (Dix and Duck, 2000) that lie well beyond onshore features previously regarded as the LLS limit, offers the possibility of an even more extensive LLS glaciation. Further bathymetric surveying of other sites where LLS glaciers terminated beyond the present coastline is necessary to determine whether this is the case everywhere around the island, although absolute dating is still required to confirm a LLS age.

*3.2.1.4 Rùm*

In comparison to its neighbour Skye, the glacial history of the Isle of Rùm (Figure 3.2) has been the focus of relatively limited research. Whilst it has long been acknowledged that a phase of local glaciation occurred on the island (Charlesworth, 1955), Ryder and McCann (1971) argued that the scarcity of terminal and lateral moraines combined with post-glacial modification of morainic deposits made identifying the outer limits of these glaciers difficult. Detailed mapping of LLS landforms on the island was first conducted by Ballantyne and Wain-Hobson (1980) who reconstructed nine glaciers in the Rùm Cuillin and a further two in the Western Hills. Six of these glaciers were reconstructed from complete or fragmented end moraines between 2 and 15 m high whereas two coalescent valley glaciers in Glen Harris and Glen Dibidil were inferred from drift limits and fluted or hummocky moraines (Ballantyne and Wain-Hobson, 1980). Relict periglacial features, such as the blockfields on the summits of the western hills and the screes in the Rùm Cuillins (Ryder and McCann, 1971), are absent within the inferred glacial limits, suggesting that these areas were protected from periglacial conditions during a relatively recent phase of ice occupancy. Thus, whilst there are no absolute dates on the features demarcating the limits, it is reasonable to interpret these features as LLS in age.

*3.2.1.5 Isle of Mull*

Like the Isle of Skye to the north, the Isle of Mull (Figure 3.7) supported an independent ice dome during the LGM which deflected mainland ice around it, before nourishing a mountain icefield during the LLS. Bailey et al. (1924) first recognised that, whilst the north and west of the island showed evidence of LGM glaciation, the valleys in the central and eastern mountainous area were characterised by hummocky deposits from a phase of valley glaciation and inferred the extent of ice from striae, erratic carry and the distribution of moraines. Although subsequent accounts presented more detailed geomorphological mapping (Gray and Brooks, 1972; Ballantyne, 2002a), the proposed extent of LLS glaciation on Mull has remained broadly similar.

Ice drained north from the cirques on the flanks of Sgurr Dearg and the Beinn Talaidh-Corra-bheinn ridge to form the Ba and Forsa outlet glaciers to the northwest and north, respectively (Figure 3.7). The lower slopes of both valleys are covered by nested lateral moraines, chains of recessional moraines and thick drift, which terminates abruptly against bare slopes above, but the termini of both glaciers are obscured by outwash (Ballantyne, 2002a) (Figure 3.7). This has caused some uncertainty as to whether the Forsa glacier extended beyond the current shoreline (Bailey et al., 1924; Synge, 1966; Gray and Brooks, 1972) or just short of it (Ballantyne, 2002a), the latter being favoured herein. In its upper reaches, the accumulation area of the Forsa glacier coalesced through a low breach with the accumulation area of the Glen More and Spelve Don glaciers, which drained ice from the southern flanks of the Beinn Talaidh-Corra-bheinn ridge and Sgurr Dearg (Figure 3.7). Charlesworth (1955) suggested that the Glen More glacier lobe reached the current coastline, but based on the position of fragmented end and lateral moraines. Gray and Brooks (1972) argued that the glacier terminated a kilometre up-valley, which was later confirmed by Ballantyne (2002a). The neighbouring Spelve-Don lobe glacier seems to have deposited few moraines, contrasting with the almost complete coverage of morainic material in the northern valleys, and much of the area is characterised by ice-moulded bedrock (Figure 3.7). The outer limit of the glacier is marked by occasional fragmented end moraine complexes and sections in the Loch Don end moraine (Figure 3.7) have revealed that this feature is a thrust moraine, formed by the LLS glacier advancing over the underlying Lateglacial glaciomarine sediments (Benn and Evans, 1993).
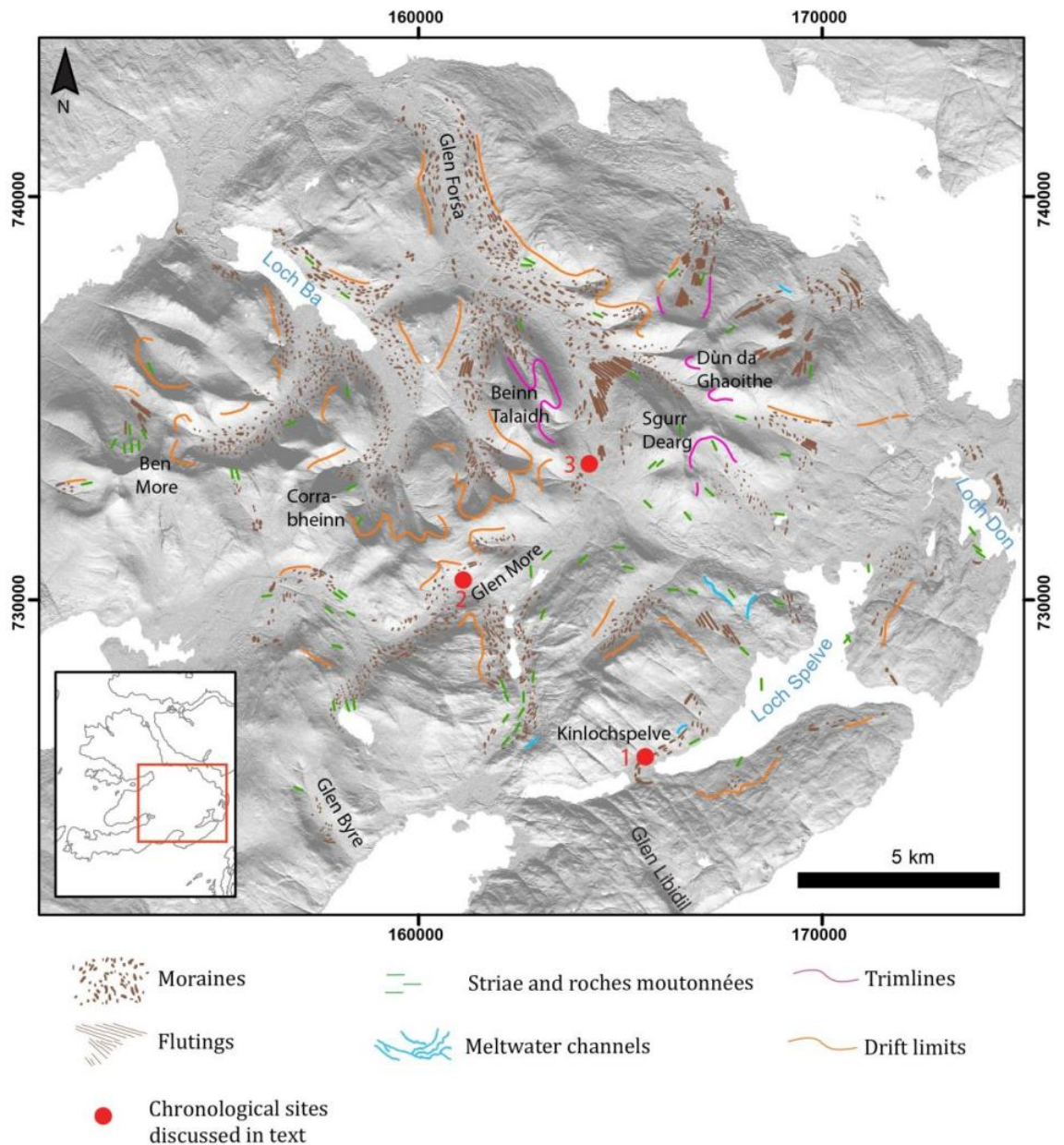
Figure 3.7 Geomorphological mapping compiled for the glacial map for the Isle of Mull (after Ballantyne, 2002a). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Six cirque glaciers flank the Mull icefield (Figure 3.7). The two largest cirque glaciers, which formed to the northeast of the Dun da Ghaoithe ridge, and a smaller glacier in Glen Byre to the south of the Glen More lobe formed clear sequences of nested recessional moraines whilst the three glaciers along the northwest periphery of the icefield produced only drift limits, striae and occasional hummocky moraines (Ballantyne, 2002a). The contrast between glaciation of only three small cirque glaciers on the flanks of the highest peak, Ben More, compared to the vast Spelve-Don lobe which extended to present eastern coastline has been attributed to the eastwards redistribution of windblown snow and further accounts for the low equilibrium line altitudes (ELAs) of those glaciers which were

situated downwind of potential snow-contributing upland areas, such as the Glen Byre glacier (Ballantyne, 2002a). Further small cirque glaciers were proposed to have developed in Glen Libidil to the south of the main icefield and in the cirque east of Ben More (Bailey et al., 1924; Gray and Brooks, 1972), and at Gribun (Dawson et al., 1987), approximately 6 km west of Figure 3.7. However, Ballantyne (2002a) found no evidence for LLS glaciers at the former two sites and argued that the arcuate ridge at Gribun represents a rockslope failure. Given the substantial distance of Gribun from the main icefield, the low elevation of the site, and the lack of a snow-contributing area upwind of it, the presence of a glacier at this site seems unlikely, and consequently, these three refuted cirque glaciers were excluded from the GIS database in Bickerdike et al. (2016).

Chronological control is somewhat restricted on Mull. Radiocarbon dating of shell fragments from sediments comprising the Kinlochspelve moraine supported a LLS age for this feature (Figure 3.7, chronological site 1), substantiated by pollen analysis of the same sediments (Gray and Brooks, 1972). The presence of only Flandrian sediments from infilled kettle hole basins within the limits in Glen More (Figure 3.7, chronological sites 2 and 3) further supports a LLS age for this period of local glaciation (Walker and Lowe, 1982). Although no other dates exist, the absence of mature talus slopes and Lateglacial raised shorelines from inside the proposed limits and their presence outside further strengthens this argument (Bailey et al., 1924; Synge, 1966; Ballantyne, 2002a).

### 3.2.1.6 Isle of Arran

Although only a relatively small body of research has been published regarding the LLS on the Isle of Arran, two contrasting styles of glaciation have been proposed for the island (Figure 3.8). Evidence for a phase of local glaciation can be found in the valleys of the mountainous northern half of the island, which is characterised by granite peaks surrounded by cirques and glaciated valleys. Gemmell (1973) noted that many of these valleys were occupied by locally derived grey till that he attributed to the decay of ice during an early local readvance, arguing that it was probably supplemented and pushed up into moraine ridges in some valleys during the LLS (Figure 3.8b). Based on the distribution and elevation of raised shorelines and the location of these moraines, Gemmell (1971, 1973) proposed that LLS glaciers had developed in the western and central glens of Catacol, Iorsa, Easan Biorach and Diomhan (Figure 3.8a and b), connecting through the cols to form a substantial icefield. Similarly, the Glen Sannox and Rosa valley glaciers joined through the col and were surrounded by numerous cirque glaciers (Figure 3.8b). Additionally, Gemmell (1973) noted the presence of fresh, steep-sided, boulder strewn moraines in the high cirques and believed they dated from a final readvance or stillstand at the end of the LLS.

Ballantyne (2007b) argued that, whilst moraines were present in the lower reaches of Glen Rosa, Glen Sannox, North Glen Sannox, Coire nam Meann, Gleann Easan Biorach, Glen Catacol and Glen Iorsa, these features were formed by an early stage of glaciation and that the extents of LLS glaciers were marked by the (usually) fresher and less fragmented, boulder-strewn moraine ridges found in the upper cirques (Figure 3.8c). This glacial reconstruction contrasted with Gemmell's (1971, 1973) interpretation. The clearest discrepancy between the two reconstructions is the extension of the glacier in Glen Iorsa. In Gemmell's (1973) earlier reconstruction it reaches the lower valley, but Ballantyne (2007b) argued the glacier terminated approximately 7 km further up-valley in a series of subdued, often peat-covered, moraine ridges. Furthermore, Ballantyne (2007b) argued that the neighbouring Tanna glacier had terminated at the outermost of a series of indistinct chains of recessional moraines, far short of coalescing with the Iorsa glacier as proposed by Gemmell (1973). Similarly, in both Gleann Easan Biorach and Glen Catacol, where Gemmell (1973) proposed two coalescent glacier lobes, the revised limits have been placed at the outermost limit of nested chains of moraine ridges, rather than at a ridge near the mouth of Glen Catacol, which Ballantyne (2007b) argued is a bedrock feature. Gemmell (1973) also proposed that valley glaciers had extended down to approximately 30 m OD in Glen Sannox and Glen Rosa and had deposited the outwash at the mouths of these valleys. However, Ballantyne (2007b) argued that the limit of the LLS glaciers is significantly further up-valley where a boulder-strewn lateral moraine cross-cuts older lateral moraines. This style of moraine was also found in upper Glen Sannox, the cirques at the head of North Glen Sannox, Coire a' Bhradain, and Coire Lan and Coire nan Larach (Figure 3.8a), where such moraines mark the limit of bouldery debris (Ballantyne, 2007b).

In the absence of chronological control it is not possible to refute the notion that the features beyond the limits suggested by Ballantyne (2007b) are of LLS age and hence potentially formed early during this period, particularly given that a two- (or three) phase LLS has been inferred at other locations based upon moraine distribution patterns (e.g. Benn et al., 1992; Brown et al., 2013; Boston et al., 2015). However, Gemmell's (1973) mapping of moraines was too coarse in its resolution for these features to be entered into the GIS. Numerical modelling by Golledge et al. (2008) indicated that Arran was glaciated by a substantial icefield, similar to that suggested by Gemmell (1971, 1973), which reached its maximum extent around 12.7 ka, but had deglaciated by 12.3 ka. Given the accordance of the model with the empirical limits in other locations, Arran certainly merits further research to establish a robust chronological control on its glacial landforms.
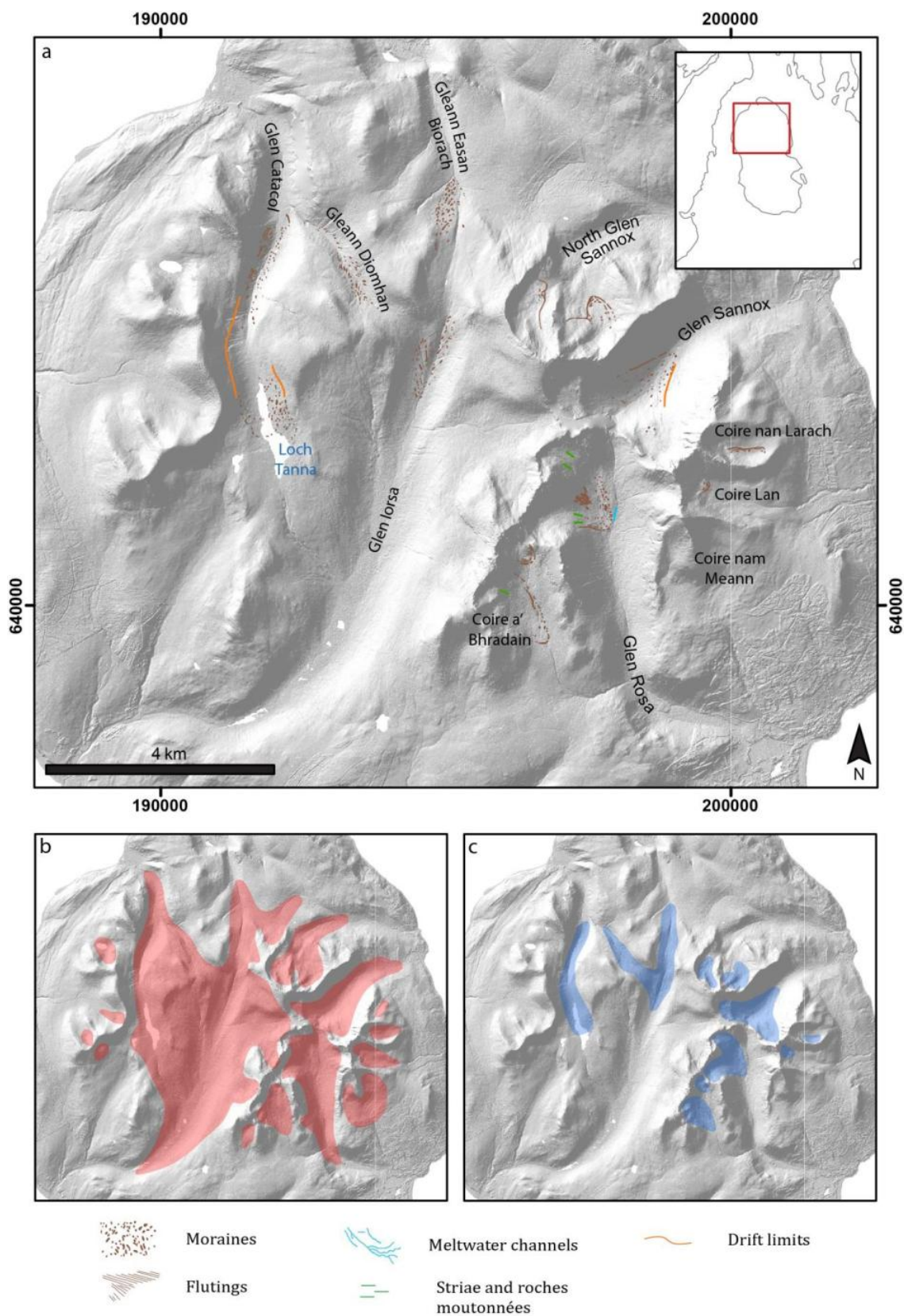
Figure 3.8 a) Geomorphological mapping compiled for the glacial map for the Isle of Arran (after Ballantyne, 2007b). The extent of LLS glaciation on Arran as reconstructed by Gemmell (1973) (b) and Ballantyne (2007b) (c). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC EarthObservation Data Centre.

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### 3.2.2 Northwest Highlands

The geomorphology of the icefields and cirque glaciers that flanked the northern sector of the main icefield in the northwest Highlands has been relatively well recorded but remapping of these features (e.g. Lukas and Lukas, 2006 a, b; Finlayson et al., 2011) has suggested that the extent of LLS glaciation in this region was far more extensive than previously proposed. A range of styles of glaciation were present in this area; an extensive icefield was nourished in the northern mountains around Ben Hee (Figures 3.9 and 3.10), stretching as far south as Glen Oykel (Lukas and Bradwell, 2010) with a small icefield forming on the Applecross Peninsula and an ice cap occupying the Beinn Dearg massif (Figure 3.11). Cirque glaciers formed in topographically favourable locations such as the cirques of An Teallach (Sissons, 1977a; Ballantyne, 1987) (Figure 3.12). Although chronological control remains fairly sparse within this region, dating in specific locations has been utilised to determine the extent of LLS glaciation (Bradwell, 2006; McCormack et al., 2011). In addition to evidence of LLS glaciation, landforms relating to the Wester Ross Readvance, an ice sheet oscillation thought to have occurred between approximately 18.0 and 16.5 ka (Everest et al., 2006), are present at sites within this region.

### 3.2.2.1 Ben Hee

Until the turn of the 20th century, the remoteness and relative inaccessibility of the far northwest Highlands meant research into the glacial history of this area was comparatively limited. Following mapping in the early 20th century by the British Geological Survey (Read et al., 1926; Read, 1931), some 50 years elapsed before the next study of glacial geomorphology in the area was undertaken (Sissons, 1977a). Unlike the early accounts, which assumed that maximum glaciation had been followed by a period of independent valley glaciers and a subsequent period of cirque glaciation, Sissons (1977a) argued that the valley and cirque glaciers were contemporaneous and reconstructed a series of 70 glaciers in the NW Highlands during the LLS, of which 30 occupied the hills of the far north. Recent remapping of this area determined that landforms attributed to the LLS were far more widespread, prompting reconstruction of a 340 km$^2$ mountain icefield stretching between Foinaven in the north and Glen Oykel in the south, covering five times the area of Sissons' (1977a) glaciers (Lukas and Bradwell, 2010) (Figure 3.9).
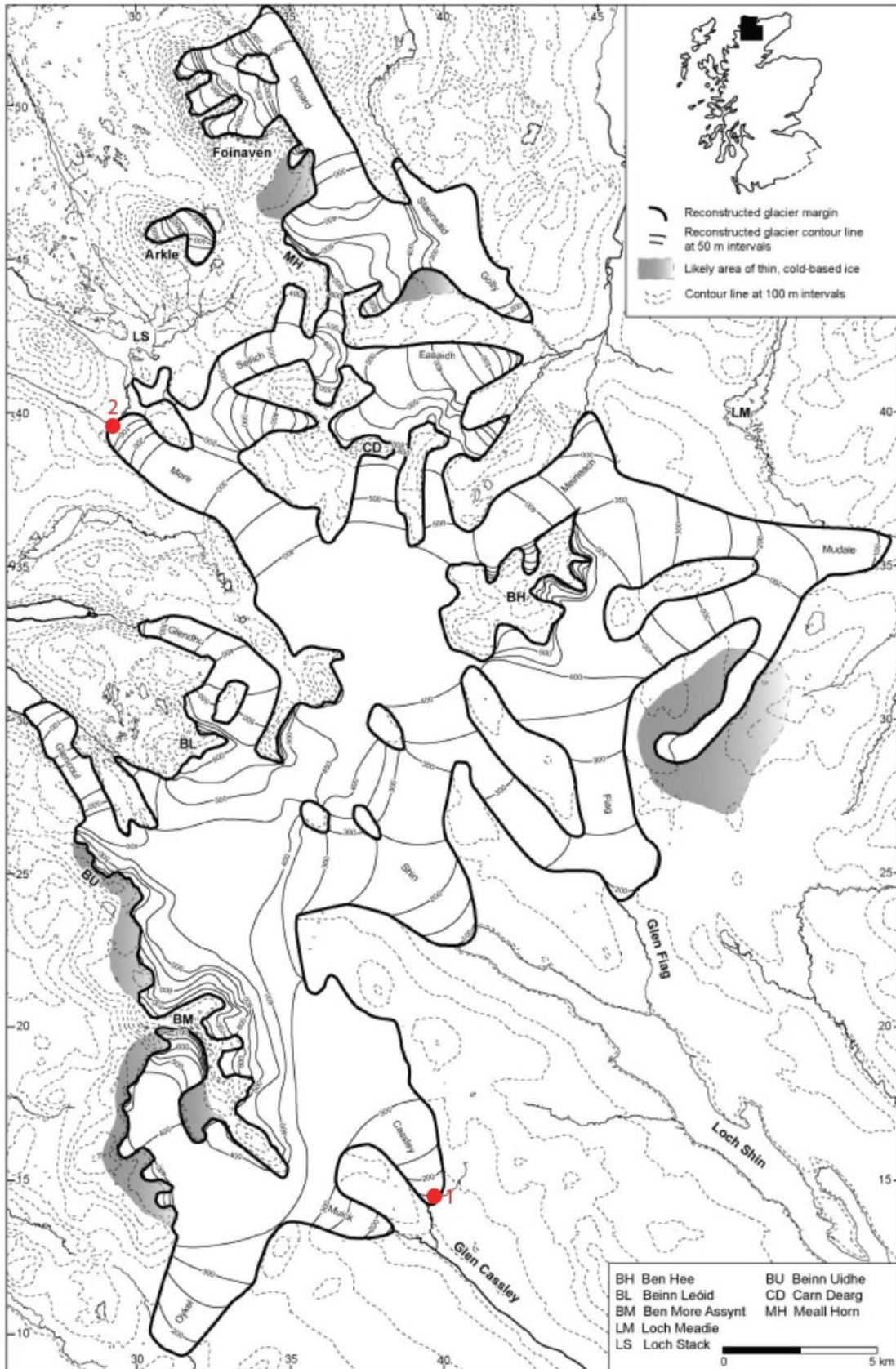
Figure 3.9 Three-dimensional reconstruction of the mountain icefield in the Ben Hee area. Chronological sites discussed in the text are indicated by the red dots. Adapted from Journal of Quaternary Science, Vol. 25(4), Lukas, S. and Bradwell, T., Reconstruction of a Lateglacial (Younger Dryas) mountain ice field in Sutherland, northwestern Scotland, and its palaeoclimatic implications, 567-580, Copyright (2010), with permission from John Wiley and Sons.

The two models of glaciation of this area are in close agreement for the northern sector of the icefield, where both Sissons (1977a) and Lukas and Bradwell (2010), using reasonably similar maps of hummocky moraines, flutings and meltwater channels, reconstructed glaciers draining the eastern cirques of Foinaven and coalescing in the valley below and then flowing northwards (Figure 3.9). However, the presence of moraines beyond Sissons' (1977a) limits and the direction of ice flow indicated by roches moutonnées was inferred by Lukas and Bradwell (2010) to show that, during the LLS, the Dionard glacier had connected through the col to the south to the main icefield and to ice in Glen Golly to the east (Figure 3.9). The central sector of the icefield comprised a series of largely topographically constrained but connected valley glaciers, which occupied an area deemed by Sissons (1977a) to have been almost entirely ice-free. The area is dominated by clear sequences of moraines, which often extend the entire length of the valleys and usually end in isolated terminal moraines. The mounds and ridges that fill these valleys are usually between 5 and 15 m in height and are clearly arranged in arcuate chains which trend obliquely down-valley (Lukas and Benn, 2006; Lukas and Lukas, 2006a, b) (Figure 3.10). Sections in these features show evidence of deformation from ice readvance subsequent to initial moraine formation in two thirds of those features sampled (Benn and Lukas, 2006) and the frequency of retreat positions was used to estimate that moraine formation occurred on average every 3-11 years for the larger glaciers, and every 7-23 years for the smaller glaciers during the second half of the LLS (Lukas and Benn, 2006).

Less detailed geomorphological mapping is available for the southern sector of the icefield. Sissons (1977a) used the extent of hummocky moraine and end moraine fragments to reconstruct four glaciers on the low ground northeast of the ridge formed by Ben More Assynt and Beinn Uidhe (Figure 3.9). Lawson (1986) made minor changes to these limits, but generally agreed on this configuration of glaciers. However, subsequent mapping reported the presence of moraines beyond these limits. For example, although boulder-strewn moraines terminate at the Fionn lochs, it was reported that hummocky moraines extended a kilometre further northeast to the shores of Loch nan Sgaraig and similarly moraines were present in Glencoul as far as the head of the loch (Bradwell, 2006) (Figure 3.9). It was inferred from these features that a substantial body of ice formed in the lee of the Ben More ridge, coalescing to the north with the Shin glacier proposed by Lukas and Lukas (2006a, b) and to the south with the Cassley and Muick glaciers (Figure 3.9), where sequences of recessional moraines were mapped (Bradwell, 2006). The Glen Cassley site is of particular significance as cosmogenic isotope dating of boulders on the outermost moraine gave uncorrected ages of 10.1±1.0 ka and 11.2±1.0 ka (Bradwell, 2006) (Figure 3.9, chronological site 1). This accords with earlier radiocarbon dates of 11,230 to 10,680 from the proglacial delta at the terminus of the More glacier (Figure 3.9, chronological site

2) and with the presence of Lateglacial tripartite sequences in eight lochs just beyond the limits (Pennington, 1977).
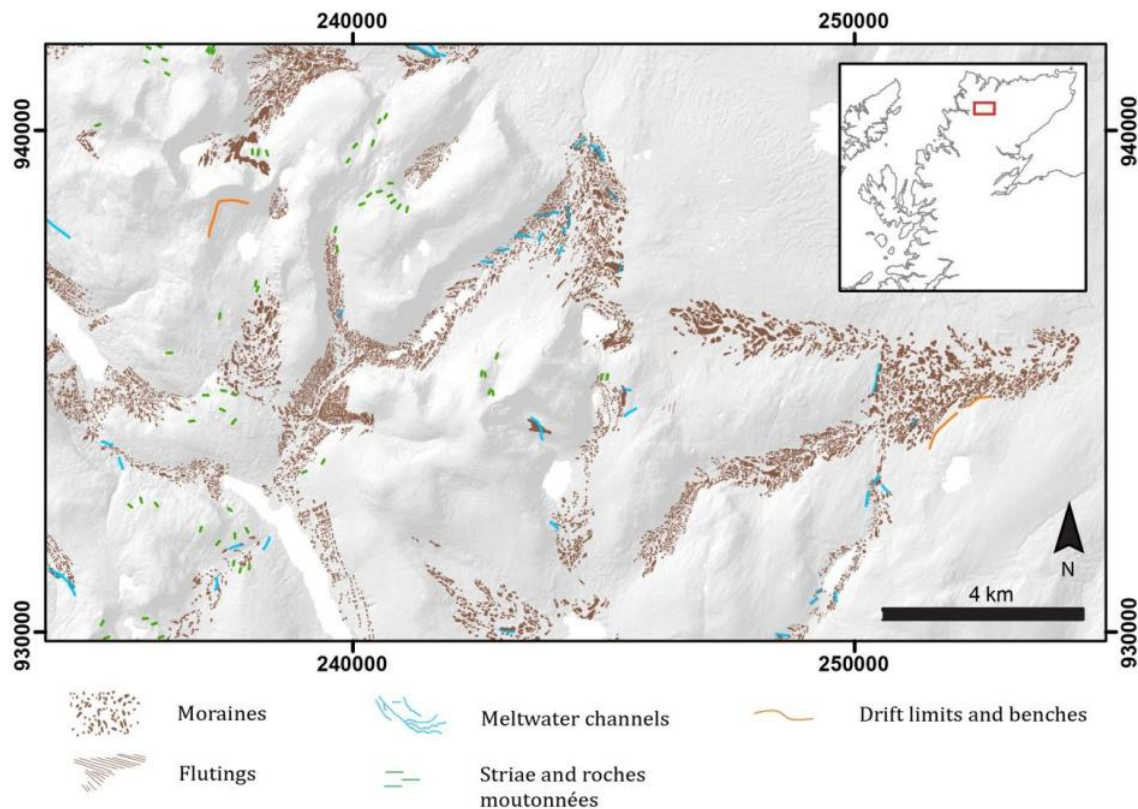


Figure 3.10 Geomorphological mapping of the central section of the Ben Hee icefield (after Lukas and Lukas, 2006a, b) showing the dense coverage of recessional moraines formed by extremely active retreat of these glaciers. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

A series of small, independent glaciers have been reconstructed in areas peripheral to the main icefield. Many of these have not featured in published mapping since Sissons' (1977a) original account, but have not been refuted in published literature as of LLS age, and thus were included in the GIS database of Bickerdike et al. (2016). These glaciers have been reconstructed from end and lateral moraines which enclose areas of hummocky moraine and generally comprise small cirque glaciers of north to easterly aspect, situated in the lee side of topographic ridges. Where the geomorphology is ambiguous, the origin of some of these features has been questioned. For example, Mills and Lukas (2009) proposed that, rather than preserving evidence of three cirque glaciers, the cirques east of Ben Hope might actually contain evidence of rockslope failures and/or rock glaciers. As yet, no evidence firmly refuting a LLS glacial origin has been published. Although much of this mapping is less detailed than that for the main icefield, a range of retreat types appear to be represented; large end moraines mark the extent of the glaciers at Cul Mor (approximately 15 km to the west of the Oykel glacier) and Ben More Coigach, where they

form multiple ridges, whereas several retreat positions are marked by recessional moraines at Arkle (Sissons, 1977a; Lawson and Ballantyne, 1995; Lukas, 2006) (Figure 3.9).

The combination of high resolution geomorphological mapping and sedimentary evidence for the Ben Hee icefield has allowed a much more detailed understanding of the extent and dynamics of LLS glaciers in this area than for much of Britain, and suggests that these glaciers underwent active retreat throughout deglaciation (Benn and Lukas, 2006). Dating of landforms in strategic locations has been used to confirm the morphostratigraphic principles outlined by Lukas (2006), allowing identification of landform criteria for a LLS age and enabling a LLS age to be inferred for similar features in adjacent valleys and, hence, for the whole icefield. Less certain is the existence of LLS cirque glaciers, generally restricted to topographically favourable sites, where either the age or origin of landforms have been disputed (e.g. Mills and Lukas, 2009). Further research to map these features in more detail and to assess their sedimentological composition may be necessary to determine whether these sites were glaciated during the LLS.

### 3.2.2.2 Beinn Dearg

During the LLS, an ice cap developed on the Beinn Dearg massif (Figure 3.11), its outlet glaciers reaching 2.5 km from the main West Highland Glacier Complex. Charlesworth (1955) envisaged that this area was occupied by a substantial ice mass during his Stage M 'mountain' glaciation, which, although undated, is broadly correlative in its extent with many accepted LLS limits. Sissons (1977a) suggested that just 12 cirque glaciers and a further two small valley glaciers had formed in the area (Figure 3.11a). Recent geomorphological mapping of Glen Alladale (Figure 3.11a) suggested that this earlier reconstruction was an underestimation (Finlayson and Bradwell, 2007) and, after reinvestigation of the whole area, Finlayson et al. (2011) reconstructed a highly asymmetrical LLS ice cap drained by particularly extensive outlet glaciers to the north and southeast, but with only small cirque glaciers in the western sector (Figure 3.11a).

The valleys occupied by ice from the Beinn Dearg ice cap are characterised by an abundance of closely-spaced, recessional hummocky moraines, usually between 2 and 5 m high (Finlayson et al., 2011), although the coverage of moraines is not continuous throughout the valleys, as was the case around Ben Hee. The termini of the glaciers are frequently marked by an abrupt transition between the sharp-crested features in the upper valleys and larger, more subdued features in the lower valleys (Figure 3.11b), which, based on their morphology, Finlayson et al. (2011) attributed to pre-LLS glaciation. An exception occurs at Loch a' Gharbhrain (Figure 3.11a) which is dammed by a large and

prominent terminal moraine, Cnoc a' Mhoraire, some 800 m long, 200 m wide and up to 25 m high (Sissons, 1977a; Gordon, 1993; Finlayson et al., 2011). During recent mapping, lateral moraines were observed trending up-slope to the plateau surface, such as in Gleann Mor (Figure 3.11a), and several valleys were found to have ice-moulded bedrock at their heads, suggestive of ice from the plateau feeding the valley glaciers. Furthermore, the position and asymmetrical nature of the moraine complex is consistent with an ice source on the plateau. Cosmogenic isotope dating of boulders on the moraine complex in a northern tributary of Glen Alladale (Figure 3.11a, chronological site 1) gave ages between 12.9±1.6 and 10.6±1.0 ka and indicated that the whole valley was glaciated during the LLS (Finlayson and Bradwell, 2007).
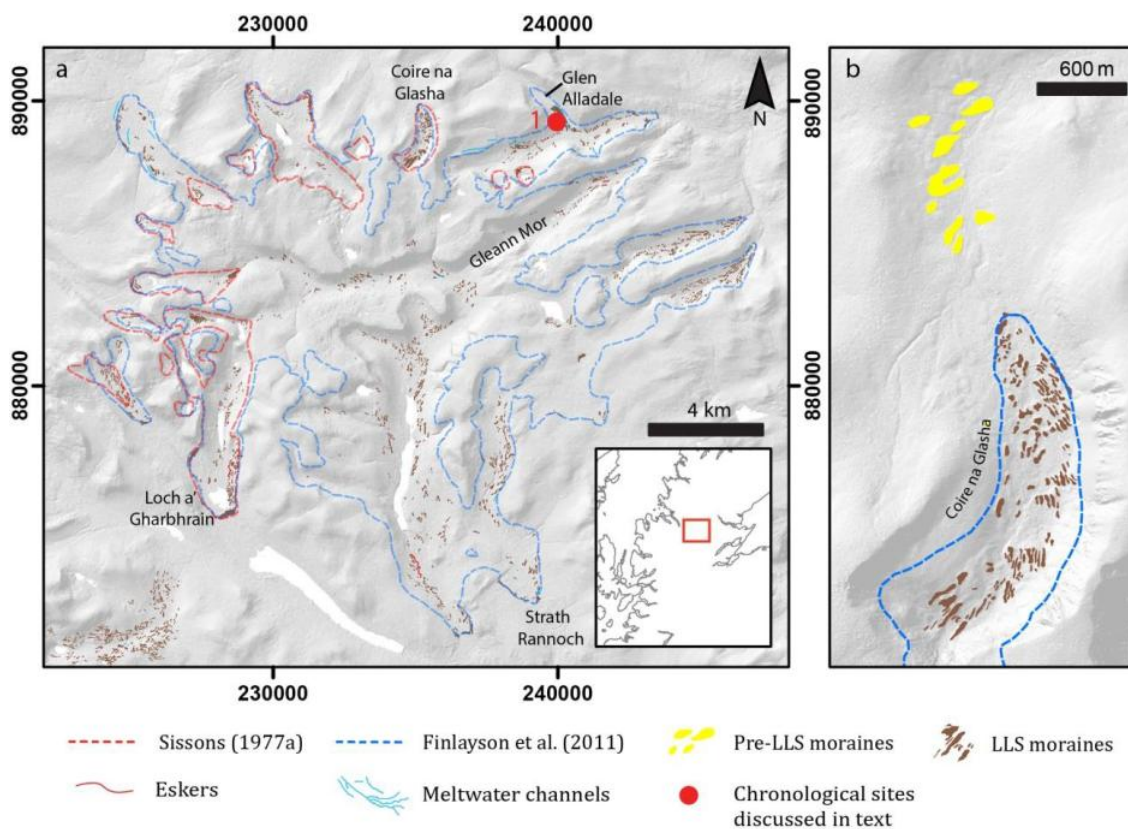


Figure 3.11 a) The glacial landforms of the Beinn Dearg massif (after Finlayson et al., 2011). The presence of moraines beyond the limits proposed by Sissons (1977a), as indicated in red, and has led to reconstruction of a much more extensive ice cap and outlet glaciers (Finlayson et al., 2011) (in blue). b) The contrast between the smaller, more sharp-crested recessional moraines attributed to the LLS and the larger more subdued features thought to predate the stadial. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Given the similarity of the geomorphology in Glen Alladale and its neighbouring valleys, a LLS age seems very reasonable for the resulting plateau ice cap reconstructed by Finlayson et al. (2011). This assumption is further supported by the tendency for numerical models to produce a plateau ice cap on the Beinn Dearg massif under a variety

of LLS climate scenarios (Golledge et al., 2008), although the underestimation and overestimation of the volume of ice in the east and west sectors, respectively, suggests that windblown snow, a factor not included in the model, was likely an important control on glacier build-up (Finlayson et al., 2011).

### 3.2.2.3 An Teallach

Six small LLS glaciers developed in the north- and east-facing cirques of the An Teallach massif (Figure 3.12), just to the north of the West Highland Glacier Complex. The extents of these glaciers are marked by end, lateral and hummocky moraines. Particularly clear examples are found at Glas Tholl where a 30 m high end moraine encloses an area of thick drift with hummocky moraine mounds and flutes (Sissons, 1977a; Ballantyne, 1987). The large volume of drift within the limits contrasts starkly with the bare bedrock beyond and is attributed to rockfall from free faces in the cirque (Benn, 1989a). Although single end moraines delimit four of the six cirque glaciers, the extent of the glacier at Mac is Mathair (approximately 2 km north of Figure 3.12) is marked by a series of five recessional moraine ridges. The outermost of these ridges truncates one of a series of older, larger drift ridges from ice sheet retreat and confirms that the cirque glaciers represent a later readvance of likely LLS age (Ballantyne, 1987).

### 3.2.2.4 Applecross

Independent glaciers in the northwest Highlands existed in the hills of the Applecross Peninsula but the extent of these glaciers has been somewhat debated. From multiple end moraine ridges, lateral moraines and flutings, Robinson (1977, 1987) reconstructed three valley glaciers flowing south-eastwards from the Applecross Hills into Loch Kishorn (Figure 3.12), the largest of which deposited an arcuate boulder belt in the loch, and possibly reached the southern shore of the loch. Mapping of moraines by Bennett (1991) supported these lateral limits but did not report the presence of trimlines or evidence from which the vertical extent of the ice could be inferred. Although Robinson thought the southern glaciers were connected through breaches in the watershed to two glaciers draining the northern side of the ridge, Jones (1998) argued from the presence of trimlines in the valley heads that the glaciers were confined to the valleys and did not flow over the breach. Subsequently, McCormack et al. (2011) argued that there were no clear trimlines in the valleys and dated the exposure of bedrock in the breach to a mean of 11.9±0.5 ka (Figure 3.12, chronological site 1), supporting the notion that the breach experienced glacial erosion during the LLS.

Although the more restricted ice model accords more closely with the results of numerical modelling (Golledge et al., 2008) it is likely that windblown snow, which is not accounted

for by the model, would have played a significant role in the distribution of glaciers in Applecross. The consistency of LLS ages from within the breach and the presence of a pre-LLS age of 17.5±1.2 ka (McCormack et al., 2011) on the heavily weathered bedrock knoll above inspires confidence that these ages are reliable. Geomorphological mapping by Bennett (1991) provides the greatest level of detail and most recent complete coverage but also includes landforms which predate the LLS; hence the limits of LLS glaciation proposed by McCormack (2011) and Robinson (1987) were used by Bickerdike et al. (2016) to differentiate between the LLS and older features.

### 3.2.3 West Highland Glacier Complex

The West Highland Glacier Complex (Figures 3.1, 3.2 and 3.12) comprises three main sectors, over which the quality and coverage of published mapping varies considerably. The entirety of the northern sector of the icefield has been mapped (Bennett, 1991), and much of the Great Glen Region has been remapped recently by Turner et al. (2014), although not all landform types were recorded. Further south the mapping is extensive but less detailed (Thorp, 1984), and whilst it is possible to reconstruct the lateral extent of the glaciers in the Rannoch Moor area, the level of detail of these maps is insufficient to allow the dynamics of the glaciers to be inferred. Generally, mapping coverage of the southernmost section remains poor and although the termini of the Lomond and Menteith lobes (Figures 3.2, 3.15 and 3.16) are well-constrained (Rose, 1980, 1981; Wilson, 2005), much of this region requires comprehensive and systematic geomorphological mapping before the extent and dynamics of these glaciers can be confidently identified.

### 3.2.3.1 Torridon to Loch Linnhe

The glaciers of the West Highland Glacier Complex dominated the landscape of Scotland during the LLS, stretching between the Slioch Massif in the north and Loch Lomond in the south (Figure 3.2). For the northern half of the icefield, Charlesworth (1955) suggested that continuous ice had reached just short of the northern coastline of Scotland and had covered the width of the country between Inverness and Kyle of Lochalsh (Figure 3.2), although no field evidence was presented. Conversely, Sissons (1967b) presented a LLS icefield of much reduced area, more confined to the Western Grampian mountains. Geomorphological data for much of the area between Torridon and Loch Linnhe was compiled from mapping by Bennett (1991), whose principal aim was to investigate the distribution of hummocky moraine to reconstruct the retreat patterns of the main icefield on a scale not attempted since Charlesworth's (1955) study.
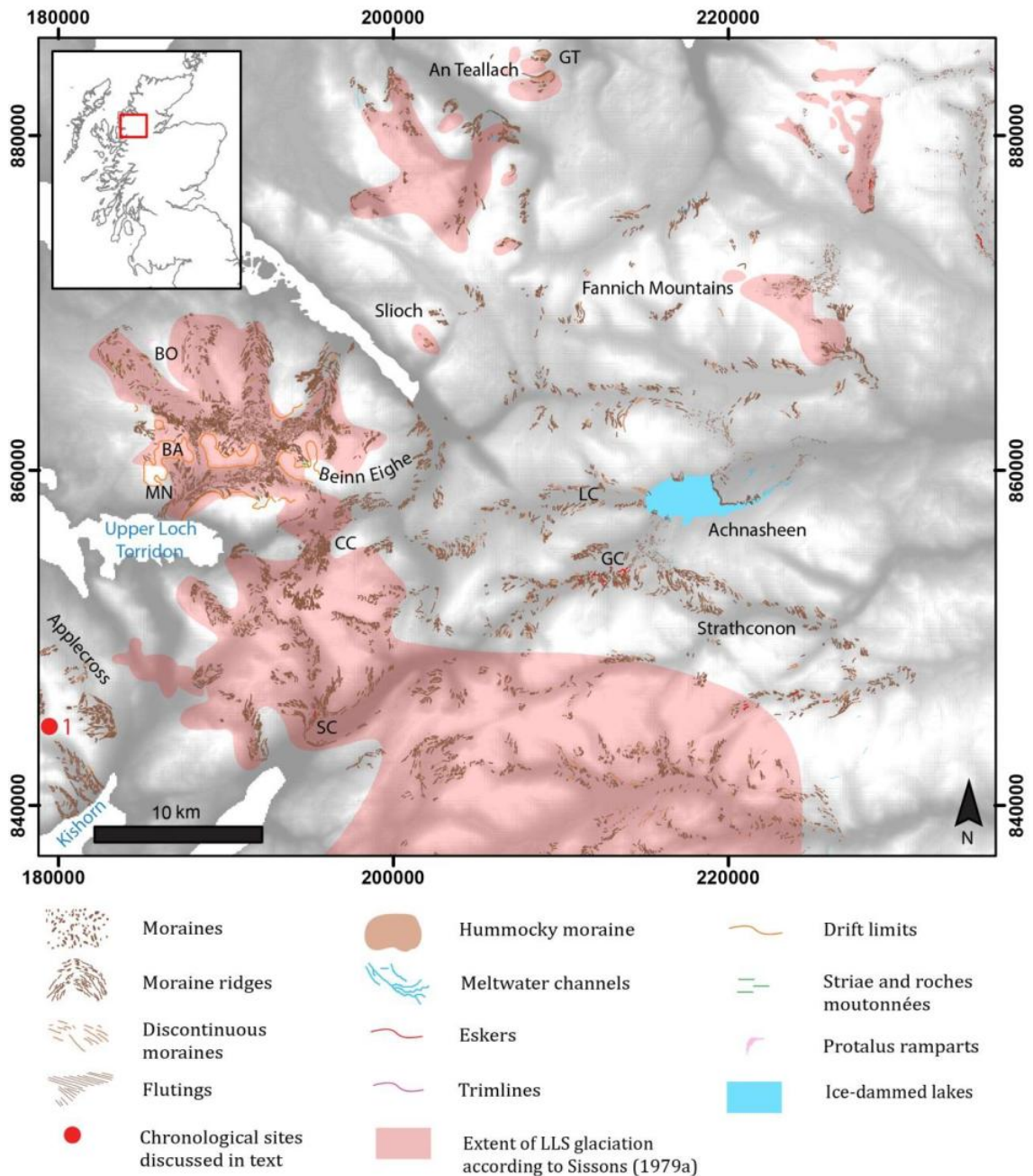
Figure 3.12 The northern sector of the West Highland Glacier Complex (after Sissons, 1977a, 1982; Ballantyne, 1986, 1987; Bennett, 1991, Bennett and Boulton, 1993b; Wilson and Evans, 2000; Wilson, 2005; Finlayson et al., 2011; McCormack, 2011). Sequences of recessional moraines extend well beyond the limits of glaciation proposed by Sissons (1979a), including at Achnasheen. The continuous nature of these features back to Sissons' limits cannot be reconciled with formation during two separate phases of glaciation, supporting a LLS age for all these features. GT: Glas Tholl, GL: Glen Carron, LC: Loch a' Chroisg, BO: Baosbheinn, BA: Beinn Alligin, MN: Coire Mhic Nobuil, CC: Coire a' Cheud-chnoic, SC: Strath Carron. Present-day water bodies reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence). GB SRTM Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA.

This distribution of hummocky moraine indicates that LLS glaciers extended across the entire width of the Highland mountains, from the hills just south of An Teallach in the west, to the Fannich Mountains in the east (Figure 3.12). The greatest difference between this reconstruction and previous interpretations was the inclusion of LLS glaciers in the vicinity of Achnasheen (Figure 3.12). The glaciofluvial terraces and associated moraine ridges observed at this site have been attributed to the presence of an ice-dammed lake in the valley, enclosed to the east by ice sourced in the Fannich Mountains and to the west by ice from Glen Carron and the valley occupied by Loch a' Chroisg (Geikie, 1901; Sissons, 1982; Benn, 1989b). In earlier interpretations (Sissons, 1982), it was argued that these features predated the LLS because the required configuration of ice did not match with the previously reconstructed LLS glacier limits to the west. However, Bennett (1991) suggested that the moraines at Achnasheen could be traced continuously westward to those limits proposed by Sissons (1977a) with no evidence of another possible LLS limit, thereby implying that the features at Achnasheen were formed by glaciers during the LLS (Bennett and Boulton, 1993b) (Figure 3.12). This interpretation is further supported by the absence of Lateglacial sediments from within the limits of the Achnasheen glaciers (Sissons, 1982).

The reinterpretation of the Achnasheen features as being of LLS age, contributed to landforms in valleys to the south of this site also being assigned to the stadial. Bennett (1991) and Bennett and Boulton (1993a) modified Sissons' (1979a) proposed ice limit in this region. They argued that a large lobe of ice had covered much of the Strathconon area (Figure 3.12), where Sissons (1979a) had suggested the limits were very uncertain, and refined the limits of LLS glaciers in Glen Cannich, Glen Affric and Glen Moriston. These updated limits are generally in good agreement with mapping by Peacock et al. (1992). As with the majority of glaciers reconstructed by Bennett (1991) and Bennett and Boulton (1993a), the extents of these eastern glaciers were reconstructed predominantly from sequences of recessional moraines on the valley sides and floors. Tipping et al. (2003) suggested that there is little depositional evidence within the area of Glen Affric, particularly in the upper valleys, and attributed this to fast-flowing ice, whilst also highlighting that trimline evidence appears to suggest conflicting thicknesses of ice within this valley. The limits of the Glen Moriston glacier are well-preserved, comprising a series of end and lateral moraine ridges, meltwater channels and associated areas of "water-washed rock" bedrock that Sissons (1977c) argued had been stripped of drift cover by meltwater during flood events. Sissons (1977c) also inferred from the presence of lake shorelines, lake sediments, terraces and overflow channels within Glen Doe that an ice-dammed lake had formed at this site.

Topographically constrained valley glaciers developed around the mountains of Baosbheinn, Liathach and Beinn Eighe and in at least one instance, truncated moraines from the earlier Wester Ross Readvance (Ballantyne, 1986) (Figure 3.12). Both the northern and southern flowing glaciers in this area formed clear sequences of recessional moraines (Bennett and Boulton, 1993a, b), with particularly clear examples in Coire Mhic Nobuil (Figure 3.12) where the contribution of debris from fresh faces produced especially prominent moraines below Beinn Alligin (McCormack, 2011). Although recessional moraines in the lower valleys indicate that retreat was initially active, deposits are more chaotic in appearance where upper reaches of these glaciers coalesced in the breach, suggesting that a later phase of ice stagnation may have occurred (McCormack, 2011).

The floor of nearby Coire a' Cheud-chnoic ('Valley of a hundred hills') (Figure 3.13) is covered by morainic mounds arranged in flutings parallel to the valley axis, the asymmetric long profiles of which give the impression of chaotically arranged, conical mounds (Hodgson, 1982, 1987; Wilson and Evans, 2000; Wilson, 2005). In the lower valley, the flutings overprint a series of valley-transverse ridges (Figure 3.13). Although these features have previously been attributed to ice stagnation processes (Sissons, 1967a; Eyles, 1983), the overprinting relationship of the flutes and transverse ridges has been noted to suggest that the features were formed by a LLS glacier overriding older cross-valley, ice sheet retreat moraines (Hodgson, 1987; Wilson and Evans, 2000). Given the widespread nature of recessional moraines in this region it seems reasonable to assume that active retreat was typical of the glaciers which comprised the northwestern sector of the main LLS icefield throughout deglaciation (Bennett and Boulton, 1993a). Landforms associated with glacial stagnation are uncommon; esker fragments and irregular mounds of sand and gravel were observed in Glen Ling and Strath Carron (Figure 3.12) by Bennett (1991) but were inferred to represent small areas of ice stagnation at the margins of actively retreating glaciers.

Further south, particularly along the west coast, the margins of the West Highland Glacier Complex are less clear (Figure 3.2). A conspicuous reduction in drift coverage occurs south of Loch Morar and recessional moraines are scarce or absent from the banks of Lochs Nevis, Morar, Ailort, Shiel, Sunart and Linnhe (Bennett and Boulton, 1993a). This absence of drift has been attributed to very active glaciers flowing through these valleys and calving in the sea lochs, removing deposits from the valley sides and leaving no depositional evidence on the steep valley sides (Bennett, 1991). Seismic surveying of these sea lochs revealed the presence of large moraine ridges across the mouths of Loch Nevis and Loch Ailort, which were believed to date to the LLS and correlated to the onshore

mounds mapped at the head of Loch Morar (Boulton et al., 1981). McIntyre et al. (2011) identified moraine ridges in Loch Hourn (Figure 3.2) extending 3.5 km beyond the proposed LLS limit, and suggested the outermost ridges had only been occupied briefly by a tide-water glacier margin which then retreated rapidly to stabilise at the main moraine sequence within the basin. Although the age of these features is not made explicit, it is emphasised that onshore evidence may underrepresent the extent of these outlet glaciers, and suggested that they are of LLS age. This is in contrast to the conclusions reached by Dawson (1988) who argued that Lochs Ailort and Moidart were not overridden by LLS ice because fragments of the Main Lateglacial Shoreline were present; these features remain undated however, and this evidence appears to be undermined by the bathymetric records.
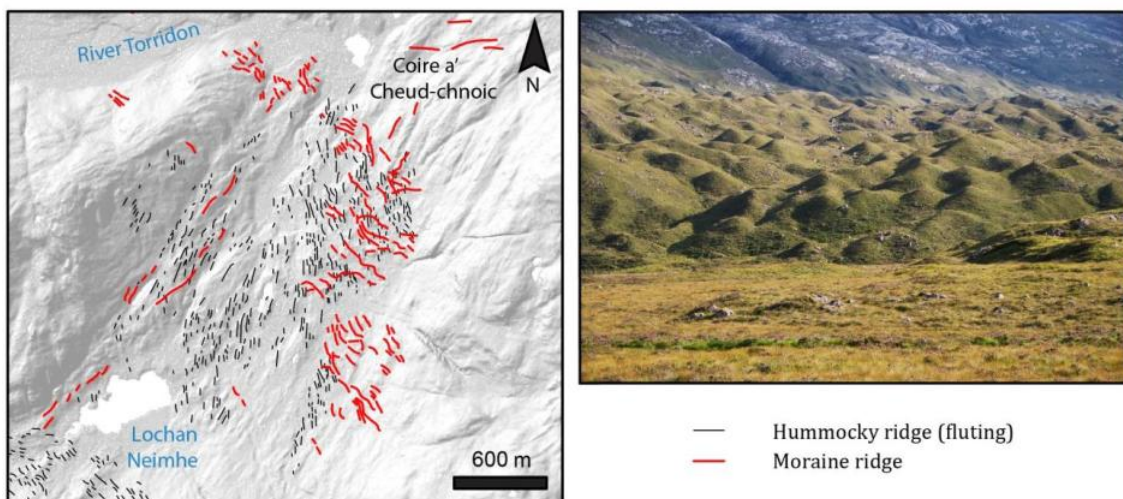


Figure 3.13 Coire a' Cheud-chnoic (after Wilson and Evans, 2000; Wilson, 2005). Although when viewed from ground level these features appear chaotically arranged, from aerial imagery it is clear that this pattern comprises linear elements, parallel to ice flow, which appear to overrun a series of transverse ridges believed to predate LLS glaciation. Photograph from H. Kinley. Underlying hill- shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The patchy nature of the evidence along the west coast, coupled with the potential that the termini of many of these glaciers extended offshore, makes reconstructing the maximum positions of west coast margins of the West Highland Glacier Complex challenging. Whilst numerical modelling suggests that the Hourn and Nevis glaciers terminated at the mouths of the lochs (Golledge et al., 2008), at other sites along the west coast there is much greater discrepancy, with modelling indicating much more extensive glaciation (Golledge et al., 2008). Given that it seems possible that LLS glaciers extended further than suggested by the onshore evidence, this area requires further research, both to identify whether more extensive limits were reached and, if so, to ascertain whether these represent an initial but short-lived advance or a temporally more sustained position.

*3.2.3.2 Rannoch Moor*

The glaciers in the northwest highlands coalesced across the Great Glen with ice nourished in the western Highland mountains and Rannoch Moor to the south (Figure 3.1), which acted as the accumulation centre for much of the main icefield. The lateral limits of the LLS glaciers in this area are generally well accepted (Thorp, 1984; Turner et al., 2014) although much of the area, particularly in the west, has not been mapped in detail since the 1980s.

Ice drained west along Glens Nevis, Leven and Coe and coalesced to form a substantial glacier which descended to sea level in Loch Linnhe (Figure 3.14). Two further ice lobes terminated at the heads of Loch Creran and Loch Etive (Figure 3.14) but end moraines are absent, or at best fragmentary at these sites; consequently the extent of these glaciers has been partly inferred from the presence of glacial outwash (Thorp, 1984). There has been some uncertainty as to whether this outwash was deposited proglacially beyond the maximum extent, as suggested by McCann (1966), or was laid down during deglaciation inside the outermost limit (Peacock, 1971; Thorp, 1984).

The termini of the eastward flowing glaciers are more clearly defined. Both the Treig and Ossian glaciers (Figure 3.14) are well-constrained by end and lateral moraines which enclose areas of hummocky moraine, whilst lateral moraine fragments, meltwater channels, eskers and an abundance of roches moutonneés were used to reconstruct the unusually wide Rannoch glacier (Thorp, 1984) (Figure 3.14). The clear limits of the coverage of hummocky moraine, which is often continuous between major valleys, such as between Glen Dochart and Loch Voil (Figure 3.2), largely informed the reconstruction of seven eastward flowing outlet glaciers from the main icefield between Loch Rannoch and the Teith Valley (Thompson, 1972).

Larger uncertainty surrounds the vertical extent of the glaciers in this area and earlier research favoured a thick ice mass. Thompson (1972) argued that most glaciers flowed along the valleys of the area but were nourished above an altitude of 900 m, although this value was extrapolated from the limits of moraines in the valleys, and Horsfield (1983) suggested that the ice cap had extended above 1000 m OD. Conversely, based on the altitude of periglacial trimlines on mountain spurs and in cols, Thorp (1981, 1984, 1986) reconstructed a laterally similar, but much thinner icefield, punctuated by 60 nunataks and with an ice-shed at approximately 750 m OD (Figure 3.14). Thorp (1984) argued that the consistency of the pattern of trimlines and its agreement with the other geomorphological evidence indicated that this evidence was a reliable indicator of the

vertical extent of the ice. However, if trimlines are interpreted to represent englacial thermal boundaries rather than the maximum elevation of ice (Fabel et al., 2012; Ballantyne and Stone, 2015), the reconstruction by Thorp becomes a minimum scenario.
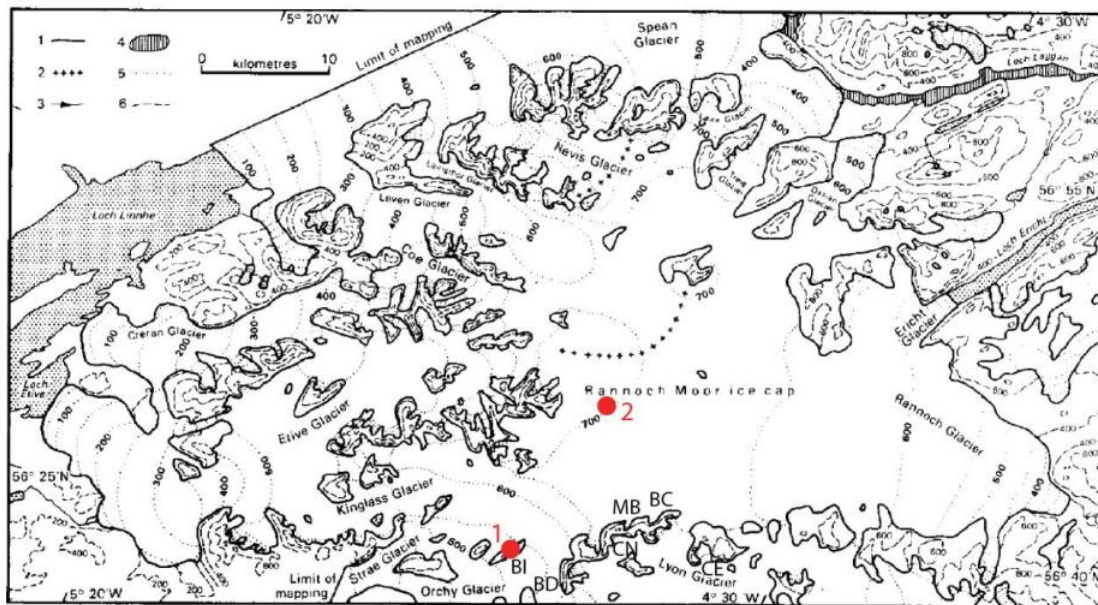


Figure 3.14 The extent of LLS glaciation on Rannoch Moor as inferred from the elevation of trimlines. Chronological sites discussed in the text are indicated by red dots. Adapted from Boreas, Vol. 15(1), Thorp, P.W., A mountain icefield of Loch Lomond Stadial age, western Grampians, Scotland, 83-97. Copyright (1986) with permission from John Wiley and Sons. 1) Ice limit, 2) Ice-shed, 3) Overflow from ice-dammed lake, 4) Ice-dammed lake, 5) Glacier surface contour, 6) Land surface contour at 200 m interval. Additional place names BI: Beinn Inverveigh, BD: Beinn Dorain, CN: Coire nan Clach, MB: Meall Buidhe, BC: Beinn a' Chreachain, CE: Coire Eoghannan.

To resolve the uncertainty regarding the vertical extent of the icefield, Golledge and Hubbard (2005) remapped the area around Glen Lyon and identified ice-smoothed bedrock in three cols on the Beinn Dorain-Beinn a' Chreachain ridge (Figure 3.14) at the head of the valley, indicating that the ice overrode the ridge at 744, 750 and 813 m.a.s.l. Although Thorp (1984) argued that ice-directional evidence from the LGM was present above the trimlines in this area, it seems unlikely that the erosion in the cols predates the LLS as the neighbouring summit of Meall Buidhe (970 m) has clearly been frost-weathered. Similarly, the presence of erratics from Beinn Heasgarnich on the Beinn a' Chreachain col at 924 m.a.s.l. agrees with LGM ice flow and indicates that the LLS did not reach this elevation. However, just below the ridge, 2-3 m high boulder moraines were observed in Coire nan Clach at 820-830 m (Golledge and Hubbard, 2005). This evidence suggests an ice surface at approximately 900 m.a.s.l. in this area. Furthermore, cosmogenic exposure dating of glacial erratics and bedrock on Beinn Inverveigh (Figure 3.14, chronological site 1) between 623 and 581 m gave ages between 12.9±1.5 ka and 11.6±1.0 ka (Golledge et al., 2007). These dates imply that Beinn Inverveigh was overrun by ice, at

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least to this elevation, during the LLS and that Thorp's maximum ice surface, which lies below this elevation, is incorrect.

In the area between Glen Lyon and Glen Lochay, Golledge (2007) observed several landsystem elements diagnostic of topographically unconstrained ice caps. These included the oblique alignment of moraines across valleys, asymmetric deposition of till in valleys, preservation of older deposits, streamlining of surfaces in high cols and a radial flow pattern indicated by ice-directional landforms. In areas such as Coire Eoghannan smaller, ice-directional features were superimposed onto larger, older ridges and in other areas pre-existing sediments were simply remoulded into arcuate cross-valley ridges rather than creating new landforms (Golledge, 2006). This perhaps conflicts with mapping of hummocky moraines on Rannoch Moor, which Wilson (2005) argued were aligned (albeit weakly) to indicate that the Rannoch Moor glacier was nourished by ice flowing into the basin from the mountains to the west. This would suggest that Rannoch Moor was a centre of convergence for glaciers flowing into the basin, rather than ice inception, as would be expected for an ice cap. However, this mapping focused on a comparatively small area and, as such, the patterns cannot necessarily be applied to the whole area. To the south of the area, around Glen Falloch, landform evidence is more indicative of topographically concordant flow (Golledge, 2007).

The style and timing of retreat of these glaciers remains uncertain. During the LLS, outlet glaciers from the main icefield advanced into Glen Roy and Glen Spean (Figure 3.2), creating a series of ice-dammed lakes (Palmer et al., 2010; Turner et al., 2014), the shorelines of which form the famous Parallel Roads of Glen Roy. Cosmogenic nuclide dating of bedrock on the 325 m shoreline gave mean formation ages of between 11.9±1.5 and 11.5±1.1 ka (Fabel et al., 2010). This implies that the Spean glacier did not reach its maximum position until the latter part of the LLS, although Ballantyne (2012) recalibrated these dates to suggest that LLS glaciers were retreating from 12.0 ka onwards. The later dates support the results of the Lochaber Master Varve Chronology (Palmer et al., 2010), from which it was inferred that the Spean glacier had only advanced sufficiently far to initiate damming of the 260 m lake by 12,165 a BP, indicating that these glaciers were continuing to advance relatively late during the LLS (Palmer et al., 2010). Similarly, numerical modelling by Golledge et al. (2008) indicated that although most of the LLS glaciers had reached their maxima by the mid-stadial (around 12.5 ka), the Rannoch Moor glacier was not at its maximum extent until 12.1 ka, suggesting that such a late advance is not unrealistic. However, this evidence conflicts with radiocarbon dates on plant macrofossils from the base of a series of cores taken from basins on Rannoch Moor

(Bromley et al., 2014), the proposed centre of the icefield (Figure 3.14, chronological site 2). These dates gave a most probable estimate of 12,262±85 cal a for the colonisation of the area by vegetation, indicating that ice reached its maximum extent in this area very early in the stadial (Bromley et al., 2014). Since this date would require a remarkably rapid glacier advance and retreat if the LLS glaciers were to reach their accepted limits, Bromley et al. (2014) suggested that the LLS glaciers either grew from pre-existing ice masses, as proposed by Bradwell et al. (2008) in the northwest Highlands, or advanced prior to the onset of the stadial and retreated under warmer summer conditions during the stadial. More recently, cosmogenic isotope dating of boulders situated on the crest of a moraine which impounds Bromley et al.'s (2014) core sites, gave a best estimate age of 11.5±0.6 ka (Small and Fabel, 2016). These ages cannot be reconciled with Bromley et al.'s dates and support the notion of glacier ice remaining on Rannoch Moor until the close of the LLS or early Holocene.

The combination of evidence, including the elevation of ice-smoothed bedrock and locally sourced erratics, the cosmogenic isotope dates and the presence of the landsystem elements described above, is impossible to reconcile with Thorp's (1981, 1984, 1986) reconstruction of a thinner icefield and strongly supports the thicker, domed ice cap model favoured by Golledge and Hubbard (2005). This configuration accords well with numerical modelling by Golledge et al. (2008) but conflicts with the dates collected by Bromley et al. (2014). The conflicting chronological evidence poses some important questions about the timing and style of LLS glaciation, which will require more chronological data and landform mapping on Rannoch Moor (cf. Wilson, 2005) to resolve.

### 3.2.3.3 Loch Lomond and the Trossachs

The southwest sector of the main icefield remains understudied and little published literature on the glaciation of this area during the LLS is available. Sutherland (1981) mapped various glacial features at specific sites within the area, but this study was focused primarily on the raised shorelines and much of the area still lacks detailed geomorphological mapping. At Ardentinny, on the shores of Loch Long (Figure 3.2), and at Furnace and Strathlachlan, on Long Fyne, hummocky moraine and outwash marks the termini of glaciers in these sea lochs (Sutherland, 1981) but much of the area is either free of deposits or geomorphological evidence has been obscured by forestry or infrastructure. Lateglacial raised shorelines were observed outside these limits, but were absent from within, indicating a LLS age for this period of glaciation. Sutherland (1981) identified periglacial features above approximately 750 m OD on Beinn Ime and 800 m OD on Beinn Vorlich, suggesting that these summits formed nunataks above the surrounding ice. The existence of small cirque glaciers at Black Craig, Corrachaive, Corarsik, Stronlonag and

Creag Mhor on the Cowal Peninsula (Figure 3.2), beyond the icefield limits, was inferred from end moraines and cross-cutting striations. However, these features were not mapped by Sutherland (1981) and no glacial reconstruction was presented.

Likewise, with the exception of the Menteith and Callander lobes, the glaciers which drained the southeast sector of the West Highland Glacier Complex have been largely overlooked in the published literature. It is thought that LLS glaciers occupied Glens Lochay and Dochart and almost coalesced with each other at the head of Loch Tay, where the series of large hummocks that occupies the upper valley comes to an end (Wilson, 2005). Similarly, topographically constrained glaciers draining the main icefield terminated at the head of Loch Earn and independent valley glaciers were nourished in the hills to the south of the loch (Sissons, 1976). It has also been proposed that a glacier developed in Glen Almond, based on the apparent presence of hummocky moraine within the glen and its tributary valleys (Thompson, 1972). However, the location of this glacier (approximately 15 km east of the main icefield), its apparent absence of a clear source area and the lack of dating to confirm a LLS age, result in great uncertainty regarding the presence of a LLS glacier in this valley. Indeed, the glacial reconstruction for much of the area between the Rannoch and Callander lobe relies solely on the limits proposed by Thompson (1972) and represents a region that urgently requires more detailed and systematic mapping.

At the southern end of the main icefield, a large piedmont lobe drained the Loch Lomond basin while other glaciers terminated beyond the present marine shoreline in Gare Loch and Loch Long (Figure, 3.15). Much of the maximum extent of the Loch Lomond lobe is demarcated by a continuous belt of well-developed moraine ridges, the area inside of which is largely occupied by widespread, till hummocks (Rose, 1981) (Figures 3.6b and 3.15). The presence of large areas of outwash and eskers along the western shore of the loch has been interpreted as indicative of glacier thinning and the production of large volumes of meltwater and glaciofluvial sediment (Rose, 1980). This terrain contrasts with that found outside the limit which is characterised by northwest to southeast aligned drumlins formed by the earlier, Late Devensian Ice Sheet and periglacial features on steeper slopes. Near the terminus of the Lomond glacier (Figure 3.15), the ice flowed a short distance down Glen Fruin whilst the majority inundated the lowland area south of the loch, terminating in a belt of moraine ridges which runs broadly west to east between Alexandria and Craighat. The limit then turns sharply north towards Gartness and takes the form of a series of six recessional moraine ridges (Rose, 1981), the outermost of which changes along its length from a flat-topped feature comprising laminated silts and

glaciofluvial sands and gravels in the south to a sharp-crested ridge of till in the north (Figures 3.6c and 3.15). It is believed that the southern part of this moraine was formed subaqueously in a proglacial lake formed when the advancing Loch Lomond piedmont lobe blocked the drainage of the Endrick Water, deforming sediments on the proximal side of the ridge (Benn and Evans, 1996; Evans and Rose, 2003a, b; Benn et al., 2004; Evans and Wilson, 2006a).
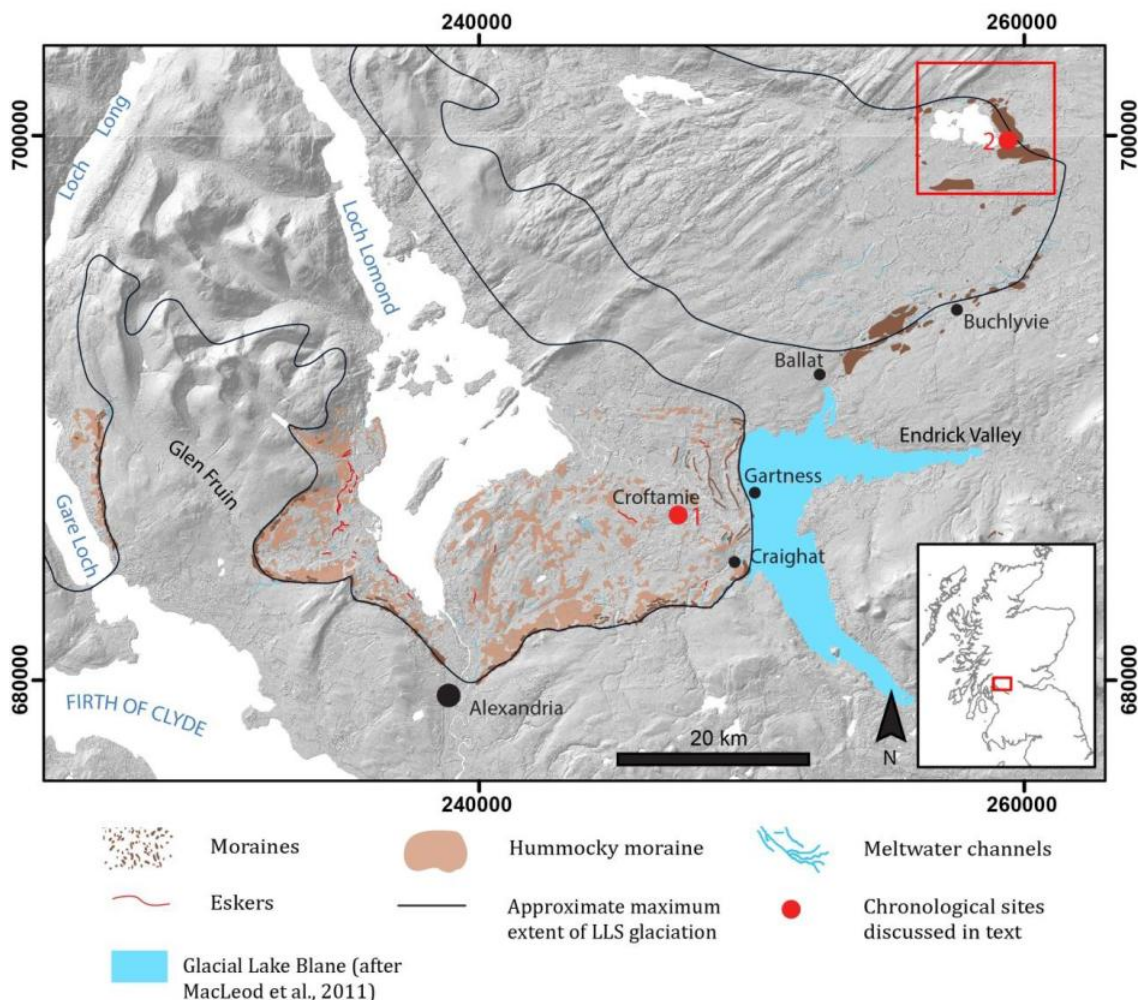


Figure 3.15 The geomorphology of the Gare Loch, Lomond and Menteith ice lobes (after Rose, 1980, 1981; Wilson, 2005). The approximate LLS glacier extent is included to demonstrate the restriction of mapped geomorphological evidence to the termini for these glaciers. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The glacial history of the area has been well recorded in a number of stratigraphic exposures, the most important of which is the official type site for LLS glaciation at Croftamie (Rose et al., 1988; Rose and Lloyd-Davies, 2003) (Figure 3.15, chronological site 1). The stratigraphic succession for the whole basin (Table 3.1) comprises: the basal LGM Wilderness Member/Till; overlain sequentially by the Killearn Member (glacier lake sediments); Lateglacial glaciomaine Clyde Beds (Linwood and Paisley Members); a layer of organic detritus (Croftamie Member), radiocarbon dated at 12,115-11,627 cal a BP

(MacLeod et al., 2011); topped by LLS deposits comprising the Blane Member (glaciolacustrine), Gartocharn Member (till) and Drumbeg Member (glaciofluvial and deltaic). Croftamie is a unique site in Britain for the presence of Lateglacial plant remains between tills which thereby allow age assignments of LGM and LLS (Evans and Rose, 2003a). The lower till was deposited by the last British Ice Sheet and was overlain by the Killearn Member silts, deposited into proglacial Lake Blane formed during deglaciation by glaciers damming the drainage of the Endrick and Blane valleys (Figure 3.15). As global sea levels rose, a marine transgression occurred, inundating the area and laying down the Clyde Bed marine sediments. Readvance of the Lomond glacier during the LLS once again dammed the Endrick and Blane valleys, causing glacial Lake Blane to reform and deposition of the Blane and Drumbeg Members by the emplacement of the Gartocharn (till) Member (Rose, 1981; Benn and Evans, 1996; Phillips et al., 2002; Evans and Rose, 2003a, b; Benn et al., 2004). The radiocarbon dates for the organic material at Croftamie (Rose et al., 1988) suggest that the Loch Lomond glacier reached its maximum extent very late in the stadial, where it remained for at least 260 years, as indicated by a varve sequence from Lake Blane (MacLeod et al., 2011). These older dates are consistent with those from Lochaber, and potentially suggest that larger glaciers, such as the Spean and Loch Lomond lobes, might have responded more slowly to mass balance changes and the changing climate, explaining why they continued to advance while smaller glaciers were retreating (Lowe and Palmer, 2008).

Table 3.1 Lithostratigraphic units in the Clyde Valley Formation, showing the oldest units at the base. The three columns reflect the changing stratigraphic nomenclature used for these deposits over the last thirty years. Adapted from The Quaternary of the Western Highland Boundary: Field Guide, Evans, D.J.A. and Rose, J., Late Quaternary stratigraphy of the Western Highland Boundary, 21-29. Copyright (2003) with permission from Quaternary Research Association.

| Rose (1980) | Hall et al. (1998) | Sutherland (1999) | |
|---|---|---|---|
| Blane Valley Silts | Blane Water Formation | Blane Member | |
| Rhu Sands and Gravels | Drumbeg Formation | Drumbeg Member | |
| Gartocharn Till | Gartocharn Till Formation | Gartocharn Member | |
| Blane Valley Silts | Blane Water Formation | Blane Member | Clyde Valley Formation |
| Organic detritus and palaeosol | | Croftamie Member (12.1-11.6 ka BP) | |
| | Killearn Formation | Killearn Member | |
| Clyde Beds | Linwood & Paisley Formations | Linwood & Paisley Members | |
| Gartness Silts | ? Killearn Formation (part) | ? Killearn Member (part) | |
| Wilderness Till | Wilderness Till Formation | Wilderness Member | |
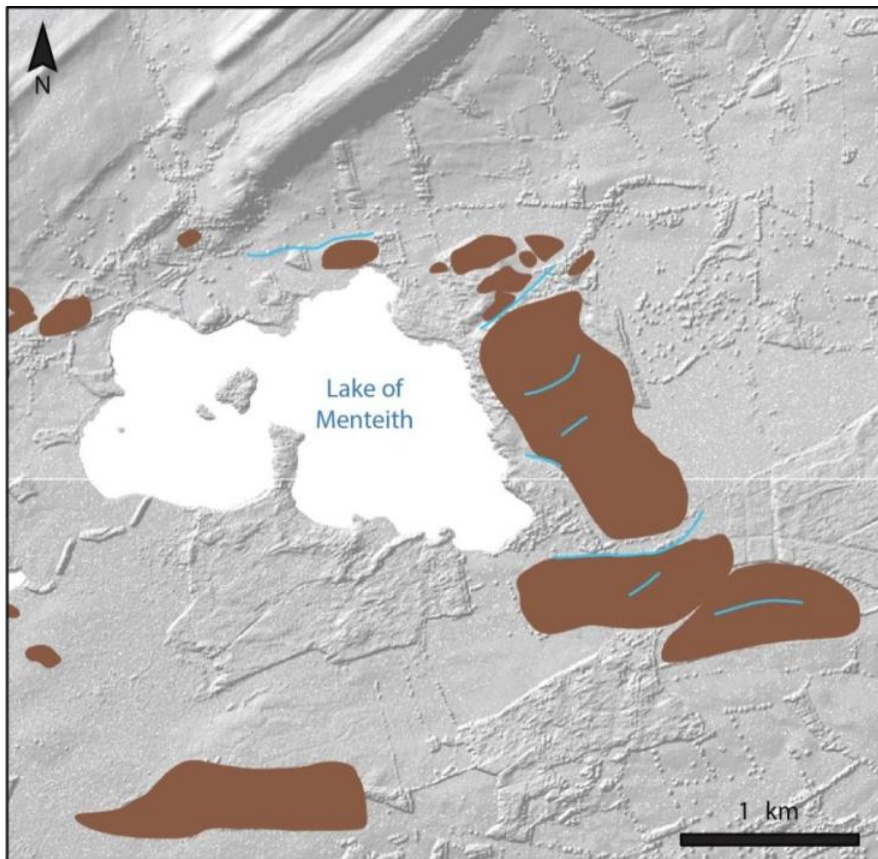
Figure 3.16 The Menteith end moraine (after Wilson, 2005). Legend in Figure 3.15. Note the similarity in size and shape between the moraine and eastern section of the lake. Location on Figure 3.15. Underlying hill-shaded images were derived from NEXT Map DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

To the north, the extent of the Menteith glacier is marked along its southern and eastern margins by a series of latero-frontal moraine ridges from west of Buchlyvie to the eastern shore of the Lake of Menteith (Wilson, 2005; Evans and Wilson, 2006b) (Figures 3.15 and 3.16). Water from Lake Blane drained into the Forth Basin through an overflow channel at Ballat, and under the Menteith glacier, in a complex of meltwater channels associated with the moraines (Rose, 1980, 1981; Evans et al., 2003; Wilson and Evans, 2003; Wilson, 2005). In contrast to the evidence found at Loch Lomond, there is an absence of glacial features within this limit, apart from a few possible kettle holes indicative of ice stagnation (Wilson, 2005). A large, 30 m high moraine ridge marks the limit along the eastern shore of the Lake of Menteith (Smith, 1993; Evans and Rose, 2003a) (Figure 3.16). Although the ridge was previously believed to be a push moraine (Smith, 1993), Evans and Wilson (2006b) proposed that this ridge comprises sediments excavated from the lake and that the two represent a hill-hole pair. The presence of hill-hole pairs and possible crevasse-fill sediments along this margin suggests formation by a fast-flowing, possibly surging glacier. Such an explanation is entirely consistent with a glacier flowing over deformable sediments, such as the marine beds which underlie this area, particularly if subglacial drainage of meltwater from Lake Blane through these sediments contributed to high pore-

water pressures (Evans and Rose, 2003a; Evans and Wilson, 2006b). Shells from the moraine were radiocarbon dated to 11,800±170 ^{14}C BP (Sissons, 1967) (Figure 3.15, chronological site 2), indicating a LLS age for the formation of the moraine.

The apparent limit of LLS glaciation in the upper Teith Valley is marked by a terminal moraine approximately 2.5 km down-valley of Callander (Thompson, 1972) (Figure 3.2). The moraine is particularly clear on the northern side of the valley, where it forms a boulder-strewn, arcuate ridge comprising sands and gravels (Merritt et al., 2003), whilst on the southern side it consists of a series of till-surfaced mounds (Thompson, 1972). The site is of particular significance as stratigraphic analysis from a borehole 200 m outside the terminal moraine on the valley's southern side has revealed the presence of an especially long Lateglacial sequence of lacustrine silts, overlain by sands and gravels and capped by a diamicton (Merritt et al., 1990). This diamicton, thought to be a supraglacial or proglacial flow till, is relatively unconsolidated and contains a higher proportion of material from the upper Teith Valley than is typically found in tills beyond the Callander terminal moraine, supporting the idea that it was formed by a readvancing valley glacier (Merritt et al., 2003).

Dating of organic material from near the base of the sequence gave a radiocarbon age of 12,750±70 ^{14}C BP, supporting the notion that the overlying till was deposited during the LLS (Merritt et al., 1990, 2003). Within the sequence, a transition from Lateglacial Interstadial to LLS deposits is reflected in both the pollen and coleopteran records. The former indicates that a substantial volume of sediment accumulated in this area subsequent to the onset of LLS conditions but prior to the readvance glacier overran the site (Merritt et al., 1990). A similar sequence of sediments was observed on the northern valley side, although the diamicton which capped this sequence is more compact and was interpreted as a subglacial till (Merritt et al., 2003). The combination of sedimentological, stratigraphic, palynological and chronological evidence strongly supports the idea that the LLS Teith glacier continued to advance until relatively late during the stadial, when it overran the area of gravelly outwash and deposited the locally-derived till (Merritt et al., 1990; 2003). It then retreated slightly and stabilised to form the terminal moraine ridge mapped by Thompson (1972). The presence of early Flandrian sediments in kettle holes inside the Callander moraine (Lowe, 1978), further suggests that the Teith glacier may have held its position at the terminal moraine for a relatively short period. Evidence from the Callander glacier indicates that there may be a mismatch between the limits of LLS glaciation as inferred from the extent of till deposits versus those inferred from terminal moraine positions. Thus, this example stresses the importance of using multiple lines of evidence when identifying the limits of LLS glaciation.

### 3.2.4 Central Highlands

During the LLS, a series of independent satellite icefields formed in the upland areas to the east of the West Highland Glacier Complex (Figure 3.1). Although glacial landforms are widespread in this region, in the absence of reliable chronological control, the age of these landforms has been disputed, prompting some authors to argue that areas such as the Cairngorms and the Gaick and Drumochter hills were almost ice free (Sugden, 1970; Lukas, 2003; Merritt et al., 2004a) or there was a combination of icefields and valley glaciers (Sissons, 1974; Benn and Ballantyne, 2005). In the last decade several of these sites have been mapped or remapped and, like the proposed extents in the northwest Highlands, there has been a general shift towards a reconstruction of more extensive ice coverage (Finlayson, 2006; Standell, 2014; Boston et al., 2015), which is also supported by numerical modelling (Golledge et al., 2008) (Figure 3.17).
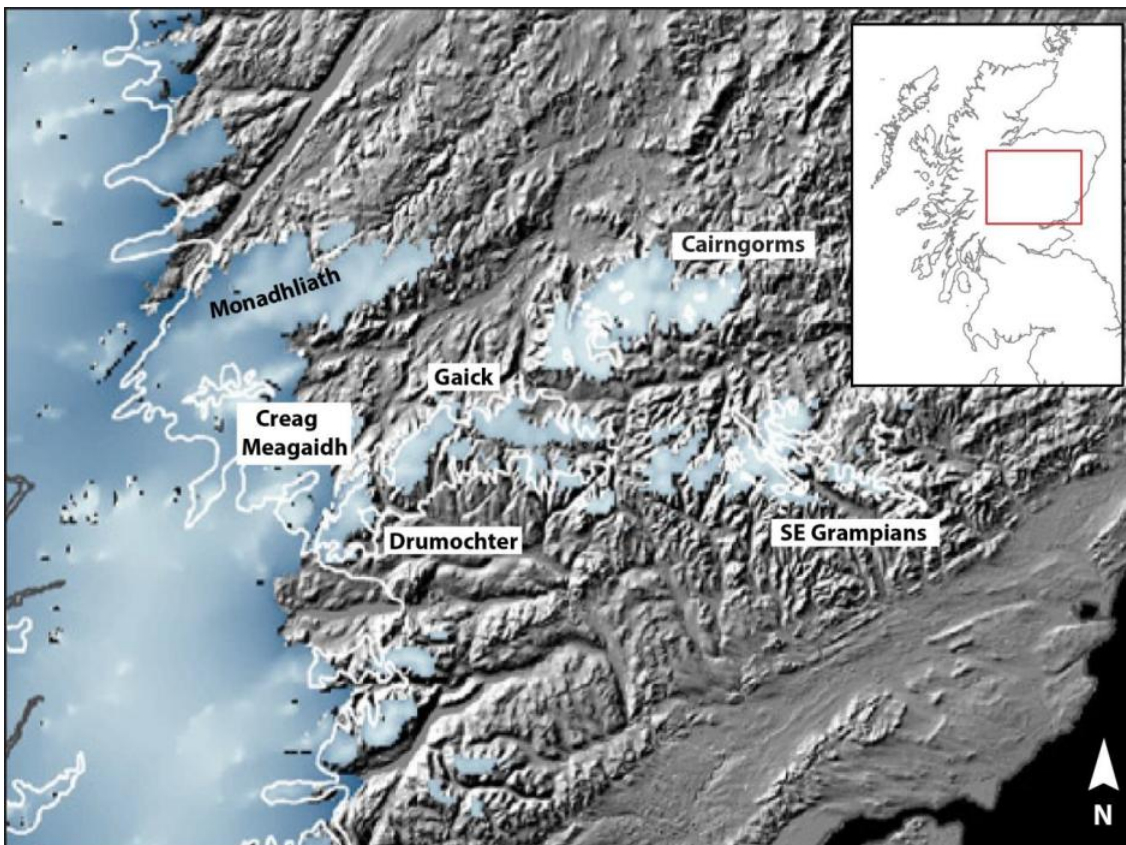


Figure 3.17 The extent of LLS glaciation in the central Grampians, adapted from Golledge et al. (2008). The solid white line indicates the extent of empirically based reconstructions, superimposed over the results of numerical modelling (from Golledge et al., 2008). The modelling supports a more extensive coverage of ice, although since publication an extensive ice field has been reconstructed on the Monadhliath mountains which accords well with the modelled extent.

*3.2.4.1 Monadhliath*

The glacial history of the Monadhliath Mountains (Figure 3.17) remained largely unstudied until the last decade. Numerical modelling (Golledge et al., 2008) indicated that the area had nourished ice on the upland plateau, but these limits remained unsupported by empirical evidence until the first systematic geomorphological mapping of the region by Boston (2012a, b). It was proposed that, during the LLS, the upland plateau nourished two coalescing icefields which were drained by outlet glaciers in many of the valleys (Boston, 2012b; Boston et al., 2013). The extents of these glaciers were reconstructed from sequences of recessional moraines (which are well preserved in 17 of 53 valleys) and, more commonly, from ice-marginal meltwater channels (Boston et al., 2013) (Figure 3.18). The recessional moraines often record ice retreat back onto the plateau where further small moraines were identified in small proto-valleys (Boston, 2012a). Combined with the presence of roches moutonneés and ice-moulding along the plateau edges (Boston and Trelea-Newton, 2013), this geomorphological evidence strongly supports a plateau style of glaciation (cf. Rea and Evans, 2003, 2007; Evans et al., 2006) on the Monadhliath. Since chronological control is lacking for most of the area, Boston (2012a, b) used a morphostratigraphic approach to identify two types of moraines (smaller ridges with a fresher appearance, or large more subdued forms) that were then used to define the LLS glacier limits (Figure 3.19).

Prior to Boston's work, only specific sites within the Monadhliath had been studied. Benn and Evans (2008) argued that a small ice cap had developed to the north of Glen Roy and that a glacier had terminated in Glen Turret close to the Turret Fan (Figure 3.18). However, Boston et al. (2013) suggested that this feature and the moraines in the lower valley predated the LLS and placed the limit further up-valley. Moraines were found that suggested LLS glaciers had advanced over fans deposited in the 350 and 325 m lakes, and it was argued that these glaciers reached their maxima late in the LLS (Boston et al., 2013). In the southeast of the icefield, Young (1978) briefly reported that Gleanns Lochain, Ballach and Chaorainn (Figure 3.18) hosted sequences of hummocky moraines at their heads. Trelea (2008) mapped evidence for glaciers in each of these valleys, plus a fourth glacier of unknown age in Gleann Fionndrigh, but was unable to determine whether these were fed by the plateau or were confined to the valley, since profile modelling suggested both configurations were plausible (Trelea-Newton and Golledge, 2012).

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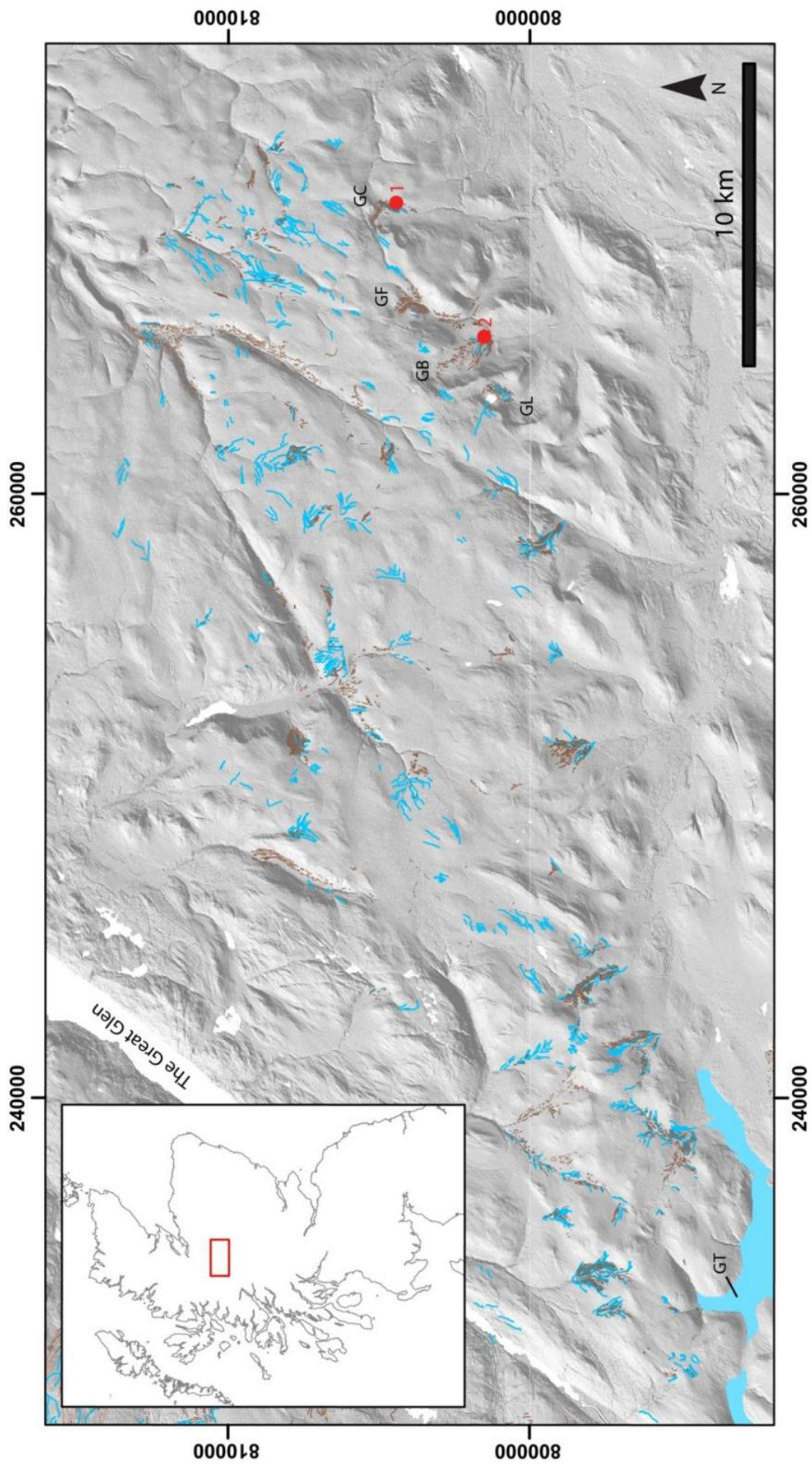
Figure 3.18 Geomorphological mapping of the Monadhliath mountains (after Boston, 2012a). Legend as in Figure 3.15, the pale blue area represents the ice-dammed lake in Glen Roy (after Turner et al., 2014). Whilst recessional moraines are apparent in around one third of the valleys, the majority of the icefield was reconstructed from ice marginal meltwater channels. GT: Glen Turret, GL: Gleann Lochain, GB: Gleann Ballach, GF: Gleann Fionndrigh, GC: Gleann Chaorainn. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The reconstructions presented by Boston et al. (2013), Trelea (2008) and Trelea-Newton and Golledge (2012) are in broad agreement, but are difficult to reconcile with cosmogenic isotope dates from Gleann Chaorainn, where cosmogenic isotope dating of boulders and bedrock gave ages of 19.3±1.1 ka, 16.8±0.9 ka and 13.7±0.8 ka (Gheorghiu and Fabel, 2013, recalibrated from Gheorghui et al., 2012). These dates suggest that the valley was last glaciated prior to the LLS, but Gheorghiu and Fabel (2013) suggested that only the middle age is reliable, the oldest and youngest being subject to nuclide inheritance and exhumation. In nearby Gleann Ballach (Figures 3.18 and 3.19) two concentric, arcuate end moraine ridges, some 300 m apart, mark the possible limits of LLS glaciation. Trelea-Newton and Golledge (2012) favoured the inner moraine ridge as the limit and Boston et al. (2015) used the outer; both were dated to the LLS at 11.4±0.6 ka and 13.0±0.7 ka respectively (Gheorghiu and Fabel, 2013), potentially indicating that LLS glaciation here occurred in two-phases.
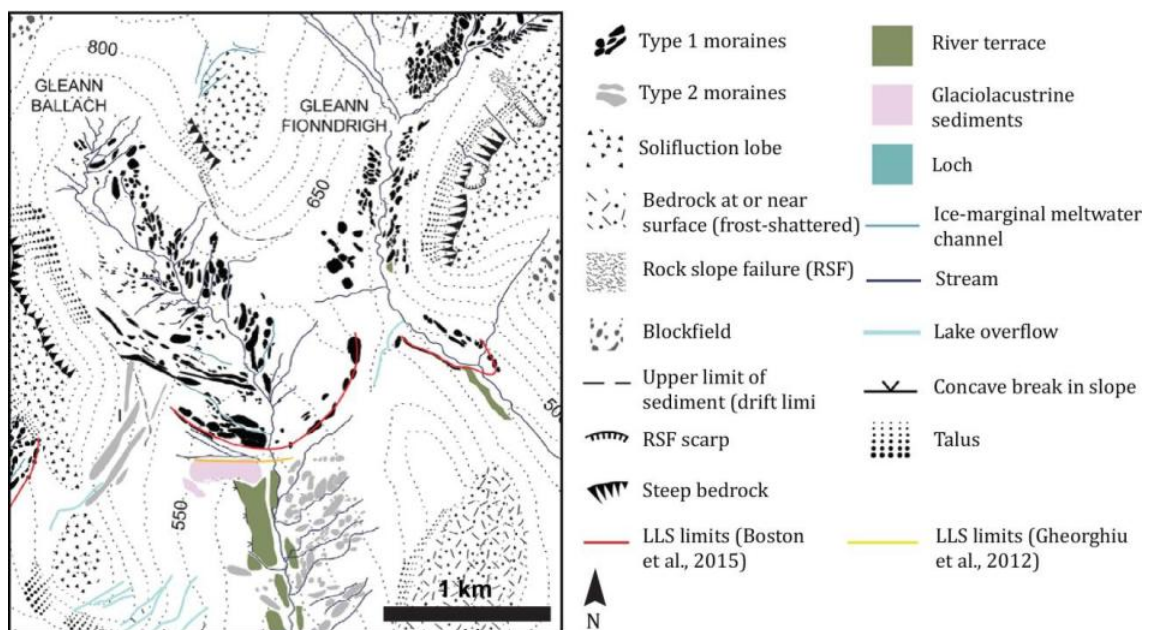


Figure 3.19 Geomorphological mapping of Gleann Ballach and Gleann Fionndrigh (adapted from Boston et al., 2015). A morphostratigraphic approach was used to determine the limits of LLS glaciation.

The presence of a plateau icefield over the Monadhliath resolves the previous discrepancy between the empirical limits of LLS glaciation (e.g. Golledge, 2010) and those inferred from numerical modelling (Golledge et al., 2008). The morphostratigraphic approach used by Boston et al. (2015) strongly supported a LLS age and the consistency of evidence between valleys allows confidence in this interpretation. However, establishing absolute chronological control has been problematic. Given the suspected unreliability of at least some of the dates in Gleann Chaorainn (Gheorghui et al., 2012; Gheorghiu and Fabel, 2013) and the persuasive geomorphological evidence, it seems unreasonable to rule out the presence of LLS ice in this valley. Further systematic dating of these valleys, and others with the same landform assemblage, might overcome these uncertainties.

### 3.2.4.2 Creag Meagaidh

Situated in the area surrounded by the Monadhliath to the north, the main icefield to the west and Gaick Plateau to the east, the glacial geomorphology of the Creag Meagaidh massif (Figure 3.17) has largely been overlooked. Sissons (1979b) used drift limits and end moraine fragments to reconstruct a valley glacier in Coire Ardair and three very small cirque glaciers in the surrounding area. Remapping of the area using a morphostratigraphic approach revealed that many of the valleys had sequences of sharp-crested, closely-spaced recessional moraines deposited during a phase of local glaciation which succeeded ice sheet retreat (Finlayson, 2006). The resulting glacial reconstruction comprised a small icefield in the west and three independent glaciers (in Glen Ardair, Coire Choille-rais and Coire Bhanain) although Finlayson (2006) highlighted that this was a minimum reconstruction and that these glaciers could have been connected to the rest of the icefield if cold-based ice had developed on the plateau. In the absence of dates, Finlayson (2006) used the hummocky nature of these moraines (which contrast with larger subdued forms found beyond the limits), the restriction of periglacial features to areas beyond the reconstructed margins, and the abundance of boulders and thick drift coverage inside these limits to infer a LLS age.

### 3.2.4.3 Drumochter and Gaick

There are two areas of uncertainty with regard to LLS glaciation in the Gaick and Drumochter area (Figures 3.17 and 3.20) that relate to the extent of glaciation across this region and the timing at which this occurred. Both Charlesworth (1955) and Horsfield (1983) proposed that the entire area was covered by an extension of the main West Highland Glacier Complex. Subsequently, Sissons (1974) mapped the area and suggested that a more restricted ice cap had developed on the Gaick Plateau (Figure 3.17), the southwest outlet glaciers of which flowed down Coire Mhic-sith, and the neighbouring Allt

an Stalacir', into the Drumochter Pass and, ultimately, coalesced with an icefield nourished on the West Drumochter Hills (Sissons, 1980b).

For the Gaick reconstruction, Sissons (1974) used a morphostratigraphic approach, inferring that the limits of this local glaciation coincided with contrasts between landform assemblages. Areas within the limits are characterised by sharp-crested, boulder-strewn, hummocky moraines and an abundance of small ice-marginal meltwater channels between 1 and 4 m deep (Sissons, 1974; Merritt, 2004a). These channels are found both associated with the moraines in the valleys and also in large numbers on the plateau. Although the moraines were recorded simply as areas of hummocky moraine rather than as individual mounds or ridges, it was argued that the landforms within the limits recorded the retreat of outlet glaciers back to their sources on the plateau during the LLS. These features contrasted with the much larger channels and major glaciofluvial deposits, including kames, found beyond the limits, and had a mutually exclusive relationship with periglacial phenomena, leading Sissons (1974) to infer a LLS age.

Merritt et al. (2004b) identified a series of problems with this glacial reconstruction, including implausible ice surface contours and accumulation centres, particularly for the western sector, for which no detailed geomorphological evidence was presented. It was instead suggested that the evidence around Drumochter represented southwestward ice sheet retreat. Because the Drumochter moraines fall outside the Rannoch Moor end moraine, which has been accepted as the outer limit of LLS glaciation, Lukas (2003) and Merritt et al. (2004b) argued that these features must predate the LLS. Furthermore, given the apparent lack of clear limits between these landforms and those on the Gaick Plateau, it was argued that these features must all represent a single deglaciation event of the last ice sheet and that LLS glaciation was restricted to a few small cirque glaciers, for example in Coire Chais and Coire Cam (Merritt, 2004b). This interpretation is consistent with the notion of a very restricted LLS glaciation in the neighbouring Cairngorms at this time (Everest and Kubik, 2006). The discovery of glaciolacustrine sediments in Coire Mhic-sith (Figure 3.20), on the eastern slopes of the Drumochter Pass has been inferred to show the presence in this valley of a lake dammed to the southwest by ice (Lukas and Merritt, 2004; Benn and Ballantyne, 2005), which is difficult to reconcile with Sissons' (1980b) model of coalescent ice between the Gaick and Drumochter.
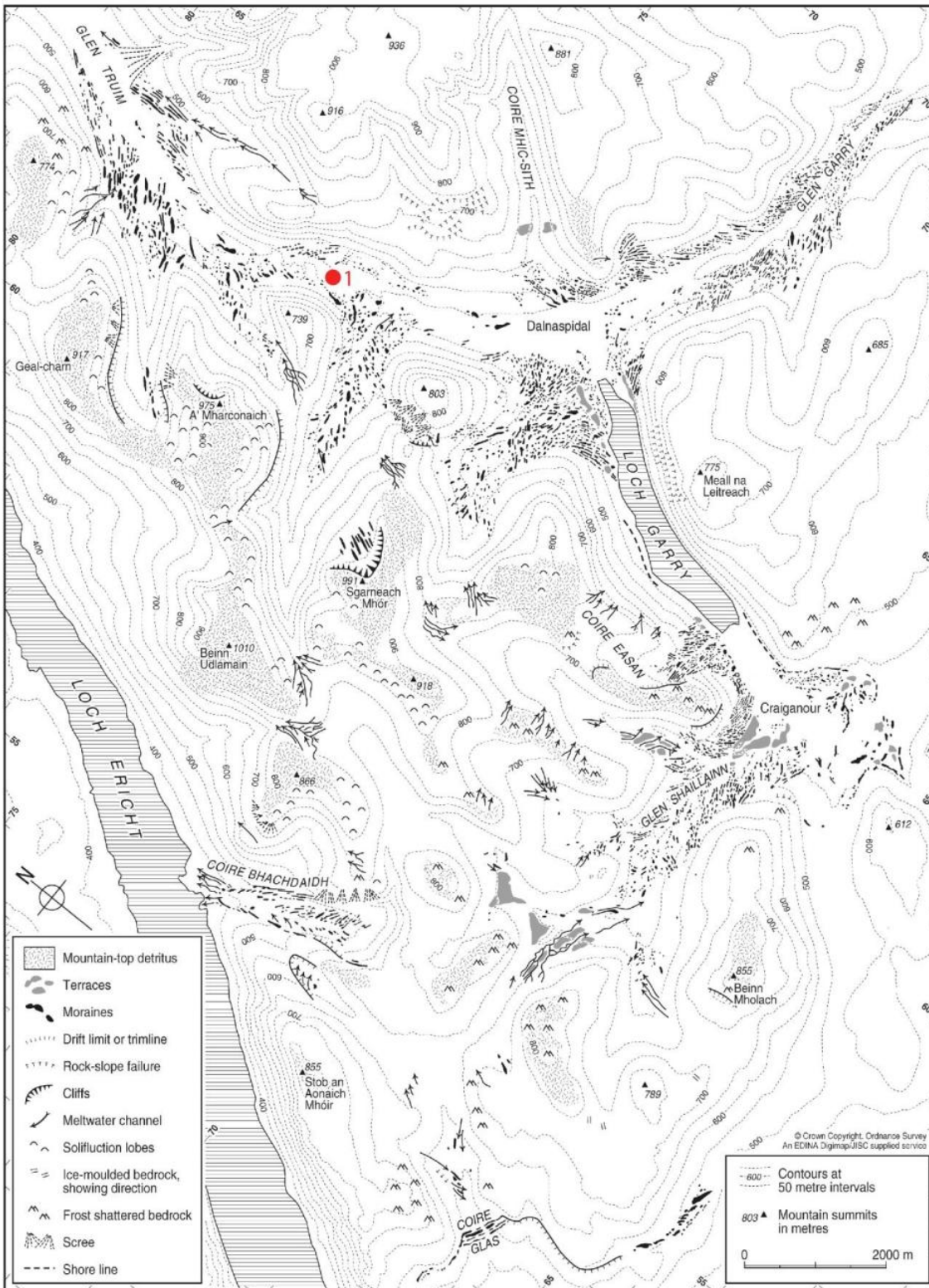
Figure 3.20 Geomorphological mapping of the West Drumochter Hills. Adapted from Journal of Quaternary Science, Vol. 20(6), Benn, D.I. and Ballantyne, C.K., Palaeoclimatic reconstruction from Loch Lomond Readvance glaciers in the West Drumochter Hills, Scotland, 577-592. Copyright (2005) with permission from John Wiley and Sons. Trimlines and the orientation of moraines in the valleys indicate active retreat of the ice up-valley to a central source rather than southwestwards retreat of the last ice sheet. The red dot marks the location of dated Holocene sediments (Walker, 1975).

The valleys of the West Drumochter Hills are dominated by hummocky, recessional moraines, which trend obliquely downslope. Particularly clear examples are found in Glen Garry and Glen Truim (Figure 3.20). Crucially, in each of these valleys, moraines and/or trimlines suggest that ice retreated up-valley towards a central source rather than to the southwest as favoured by Merritt et al. (2004a, b). Moraines are less apparent in the southwest facing valleys, a pattern which Benn and Ballantyne (2005) attributed to snowblow away from these valleys. Terraces appear to show the existence of a lake dammed by the Garry, Shallain and Easan glaciers (Figure 3.20), indicating that these lobes reached their maximum positions contemporaneously and that the Shallain glacier began to retreat before the Garry glacier (Benn and Ballantyne, 2005). Such a retreat pattern is indicative of the deglaciation of an independent icefield in the area, rather than progressive southwesterly retreat of the last ice sheet. Absolute dates are limited in this area to a single kettle hole in the Drumochter Pass (Figure 3.20, chronological site 1), where Holocene sediments, with a basal date of 10.5 ka cal BP were found (Walker, 1975). The absence of mature periglacial features from inside the limits of the West Drumochter glaciers supports a LLS age but more dating is essential to resolve these debates.

The Gaick Plateau represents one of the most understudied regions with regard to LLS glaciation, and thus, understanding of the extent and dynamics of LLS glaciation in this area is currently limited. In particular, it is uncertain how Sissons' (1974) original mapping can be reconciled with the more recent mapping of the Drumochter Hills, and whether these ice masses coalesced or were separate. In the absence of high resolution mapping it is not possible to determine whether the areas of hummocky moraine mapped by Sissons (1974) comprise recessional moraines indicative of active retreat, or stagnation terrain. Whilst modelling by Golledge et al. (2008) indicates that this area supported LLS glaciers, an updated empirically based glacier reconstruction is required to validate these findings.

### 3.2.4.4 Cairngorms and southeastern Grampians

The extent of LLS glaciation in the Cairngorm Mountains (Figures 3.17 and 3.21) has been widely debated, largely because it has proved difficult to confidently differentiate between features formed by LLS glaciers and those formed during earlier phases of ice sheet deglaciation. Sugden (1970) mapped the geomorphology of the area, noting that many of the valleys contain topographically constrained drift deposits usually arranged in distinct ridges which trend obliquely down the valley. However, based on their sandy composition, Sugden (1970) identified these features as eskers, rather than as hummocky moraines, and consequently argued that they represented ice stagnation at the close of the LGM. It was proposed that during a subsequent phase of deglaciation, ice had occupied many of

the main Cairngorm valleys and had formed a small plateau glacier on Monadh Mor and upper Glen Eidart (Figure 3.21), where moraines were identified, before undergoing widespread stagnation (Sugden, 1970). Arcuate terminal moraines were found in eight favourably sited cirques within the area and were thought to represent a later readvance. The presence of Late Devensian sediments within an area of hummocky moraine at Loch Builg (18 km east of Figure 3.21) in the eastern Cairngorms (Clapperton et al., 1975) demonstrated that this area had escaped LLS glaciation and suggested that other areas characterised by these landforms in the Cairngorms may have also remained unglaciated (Sugden and Clapperton, 1975).

This notion of limited LLS glaciation in the Cairngorms, potentially restricted to eight cirques, contrasts with Sissons' (1979c) subsequent remapping and glacial reconstruction (Figure 3.21). The two proposed ice extents are broadly comparable for the eastern and central Cairngorms but major differences are found between reconstructions for the western hills. Sissons (1979c) proposed that valley glaciers had occupied Garbh Coire, Glen Geusachan and Glen Eidart (Figure 3.21), which Sugden (1970) believed had been occupied by only small cirque glaciers. The cirque glaciers were reconstructed from the position of arcuate terminal moraines or boulder spreads, whereas sequences of recessional hummocky moraine mark the extents of the three valley glaciers (Sissons, 1979c; Standell, 2014). Given that the age of peat from within the margins of the cirque glacier beneath Braeriach precluded a Little Ice Age origin for the glacier (Buckley and Willis, 1972), Sissons (1979c) argued that the most plausible age for this readvance was the LLS. This interpretation was supported by Bennett and Glasser (1991) who broadly agreed with Sissons' (1979c) limits and inferred from the recessional moraine ridges in Glen Geusachan, that this glacier had undergone active retreat (Figure 3.21).

Cosmogenic nuclide dating of boulders on lateral moraines at the outermost margin of the Glen Geusachan glacier (Figure 3.21, chronological site 1) yielded ages between 16.5±1.0ka and 12.2±0.7 ka, with a weighted mean of 13.6±0.3 ka (Everest and Kubik, 2006), indicating that this area likely escaped LLS glaciation. However, Standell (2014) argued that the features dated by Everest and Kubrik (2006) were incised drift, rather than moraines, and that the true LLS limit was slightly further up-valley, marked by the hummocky moraine below Devil's Point (Figure 3.21, chronological site 2). Dates on these moraines, of between 10.1±0.5 and 11.4±0.5 ka, support the notion that Glen Geusachan was glaciated during the LLS, but that the ice was less extensive than proposed by Sissons (1979c) or Bennett and Glasser (1991). Standell further proposed the existence of a previously unmapped glacier in Coire Etchachan (Figure 3.21) but was unable to determine its age since different isotope production rates gave results indicative of both a

LLS and pre-LLS age. A LLS age for the cirque glaciers has been strongly supported by cosmogenic dates from the boulder sheet in Coire an Lochain (Kirkbride et al., 2014) (Figure 3.21, chronological site 3) but this small glacier is not necessarily coeval with the larger valley glaciers.
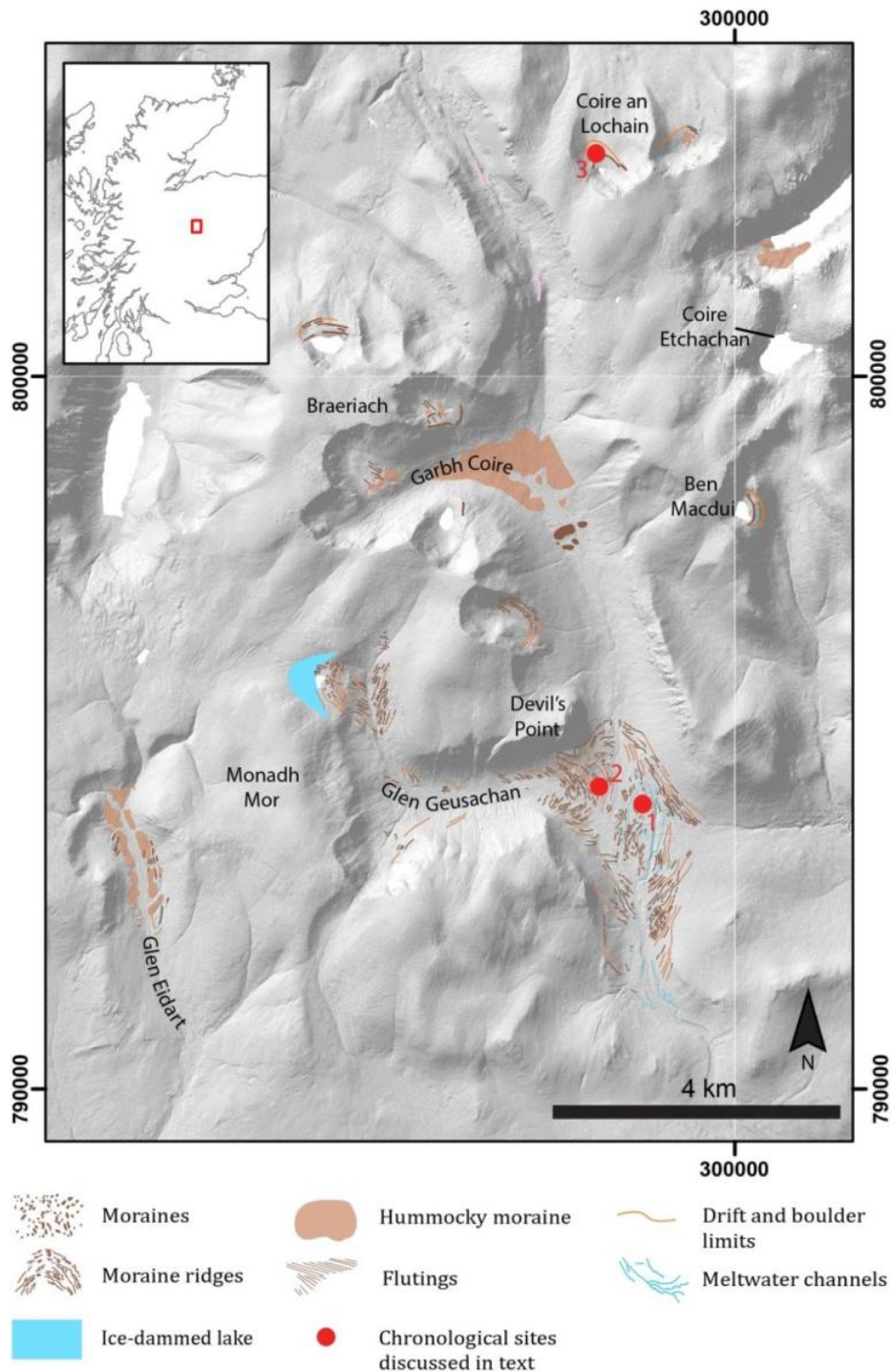


Figure 3.21 Geomorphological mapping of the western Cairngorms (after Sissons, 1979c; Bennett and Glasser, 1991). LLS glaciers were far less extensive in this region owing to the steep west to east reduction in preciptation. Whilst Standell (2014) suggested that plateau glaciers on Monadh Mor, Braeriach and Ben Macdui may have fed the valley and cirque glaciers, this remains uncertain. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

In the most recent glacial reconstruction, Standell (2014) added a further three cirque glaciers on the flanks of Braeriach and a substantial volume of ice on the northeasten slope of Ben Macdui (Figure 3.21). However, he asserted that this represented a minimum reconstruction and that the cirque and valley glaciers could have been fed by plateau glaciers on Monadh Mor, Braeriach, Ben Macdui and Beinn a' Bhuird. As with many locations in Scotland, the cirque glaciers were found to have predominately north to easterly facing aspects, minimising insolation on the glacier surface, and often occurred downwind of areas that would have contributed snow onto the glacier surface (Sissons, 1979c; Kirkbride et al., 2014). Standell (2014) calculated that these topographic factors, when combined with a south to north precipitation gradient across the area, could explain up to 80% of ELA variation. Although Standell's (2014) mapping is the most recent and highest resolution available for this region, this became available after the census date for Bickerdike et al., (2016) and thus mapping by Bennett and Glasser (1991) and Sissons (1979c) was instead used.

The southeast Grampian Mountains (Figures 3.1 and 3.17) lie adjacent to the Cairngorms, but have received far less attention in the published literature. It was proposed that a total of 27 glaciers occupied this area during the LLS, comprising an ice cap on the plateau area southwest of Mount Keen, valley glaciers in Glens Callater, Doll, Clova, Muick, Effock and West Water, and a collection of cirque glaciers (Sissons, 1972; Sissons and Grant, 1972; Sissons and Sutherland, 1976). The extent of the valley glaciers seems relatively well-constrained by abundant hummocky moraine, particularly in the lower reaches of the valleys. Similarly, the majority the cirque glaciers are clearly represented by terminal moraine ridges, which often enclose areas strewn with large boulders or morainic hummocks. The extent of the plateau ice cap is less clear, and Sissons (1972) used the elevation of meltwater channels at the heads of the surrounding valleys (including Glen Mark and Glen Lee) to tentatively reconstruct its dimensions. If this ice configuration is correct, it represents a substantially larger ice mass in the southeastern Grampians than in the Cairngorms, which Sissons (1979c) attributed to the dominant southeasterly winds causing a decrease in snowfall to the northwest.

It is striking that, whilst modelling predicts substantial LLS glaciation in the Cairngorms (far exceeding even Standell's (2014) most extensive reconstruction), it overestimates glacier extent in the western hills, producing ice on the high ground, but fails to produce the large Glen Mark ice cap in the east (Golledge et al., 2008). In light of Standell's (2014) proposal that plateau glaciation may have been more extensive in the Cairngorms, the southeastern Grampians merits further investigation to better constrain glacial

reconstructions for this area, and to determine how it fits with updated understandings of regional trends.

### *3.2.5 Southern Uplands*

LLS glaciation in the southern uplands occurred in the Galloway and Tweedsmuir Hills. The former was characterised by the formation of cirque glaciers at sites that received abundant windblown snow, the volume of which offset the effect of insolation in some cirques. Conversely, the Tweedsmuir Hills are thought to have been occupied by plateau ice which fed substantial valley glaciers (Pearce et al., 2014a, b).

*3.2.5.1 Galloway*

The extent of LLS glaciation in the Galloway Hills is thought to have been restricted to 11 cirques in the area, seven in the Merrick Hills and four in the Rhinns of Kells (Cornish, 1981) (Figure 3.1). Although Geikie (1863) proposed that the area had experienced local glaciation following ice sheet retreat, many of the early accounts (Jolly, 1868; Sissons, 1976) simply describe prominent landforms attributed to this subsequent period of restricted glaciation. End or lateral moraine ridges, typically about 3 m in height and strewn with boulders, mark at least part of the margins of five glaciers. The largest and clearest of these features, the Tauchers moraine complex, comprises a large outer ridge up to 200 m wide and 12 m high, which trends at both ends into lateral moraines and a much narrower inner ridge (Cornish, 1981). These two nested ridges suggest that limited readvances interrupted overall retreat (Ballantyne et al., 2013). Where end moraines are absent, glaciers have been reconstructed from boulder limits and the abrupt termini of hummocky moraine, the patchy coverage of which was mapped inside many of the moraine limits (Cornish, 1981). The disruption of LGM till ridges by the end moraine of the Balminnoch glacier, and the truncation of a till deposit where it meets the hummocky moraines of the Loch Dungeon glacier, were used to infer glacier readvances (Cornish, 1981). The absence of periglacial debris lobes within the limits of these glaciers but their appearance immediately outside the proposed limits supported a LLS age (Cornish, 1981).

The reconstructed ELA for the Tauchers glacier was significantly lower than those for the other glaciers (Cornish, 1981), suggesting that this glacier might date to an earlier phase of glaciation. Subsequently, Ballantyne et al. (2013) dated boulders on the northern end of the moraine ridge associated with this glacier using surface exposure dating, and found mean ages between 12.01±0.78 and 11.91±0.77ka (depending on the scaling scheme used), strongly supporting a LLS age for all the cirque glaciers and potentially indicating that deglaciation occurred before the end of the stadial. Several of the glaciers with particularly low ELAs, including the Tauchers glacier, had potentially large snow-

contributing areas to the southwest. Whilst generally glaciers facing between north and east were larger than those facing other aspects, the snowblow seems to have offset this effect, as observed for the southeast facing glacier on the east slope of Meikle Millyea (Cornish, 1981). Similarly, the largest glacier, the Loch Dungeon glacier, which accounted for almost a third of the total ice in the area, was nourished by snowblow from the large upland area of Meikie Millyea and Milldown to the southwest.

## 3.2.5.2 Tweedsmuir

More extensive LLS glaciation has been proposed for the Tweedsmuir Hills in the eastern Southern Uplands. Early research largely comprised descriptive accounts of the geomorphological evidence (Geikie, 1863; Young, 1864; Brown, 1868). Price (1963) assumed that downwasting of the last ice sheet caused ice to become topographically constrained in valleys where it stagnated before a phase of "local valley glaciation…took place in the high hills" (Price, 1963: p327). The evidence for this reconstruction consisted of well-preserved morainic evidence in Talla, Gameshope, Fruid, Polmood and Manor valleys, coinciding with Young's (1864) descriptions. Price (1963) used this evidence to reconstruct the limits of ice in the valleys but, in the absence of evidence, simply labelled parts of the upland areas as plateau glaciers rather than proposing actual limits. Sissons (1967a) presented similar glacial limits, reconstructing valley glaciers in the main valleys of Winterhope, Talla, Gameshope and the western end of Megget Water but did not present geomorphological mapping for this area. Conversely, May (1981) argued that LLS glaciation had been restricted to three valleys.

The area has recently been systematically mapped by Pearce et al. (2014a) who identified two types of moraines on the valley floors and lower slopes of this area. The first type of moraine occurs in the upper valleys and comprises closely-spaced, sharp-crested chains of ridges, which trend obliquely down-valley. Particularly clear examples of these recessional moraines are found in Talla Valley and at the foot of Loch Skene (Figure 3.22). Most of the valleys in this region lack backwalls and in some instances these moraines extend up onto the plateau, as observed in Fruid Valley where they are associated with lateral meltwater channels, strongly indicative of a plateau icefield style of glaciation (Rea and Evans, 2003; Pearce et al., 2014a). A second type of moraine was found sporadically at lower elevations, these having much more subdued morphology and no apparent pattern. Reconstruction of the extent of the icefield, based on this geomorphological evidence suggested that the Type 1 moraines mark the extent of LLS glaciation, with the Type 2 moraines predating the stadial. The similar configuration of reconstructions based on the empirical evidence and those informed by numerical modelling of the ice surface profiles (Pearce et al., 2014b) inspires confidence in these results.
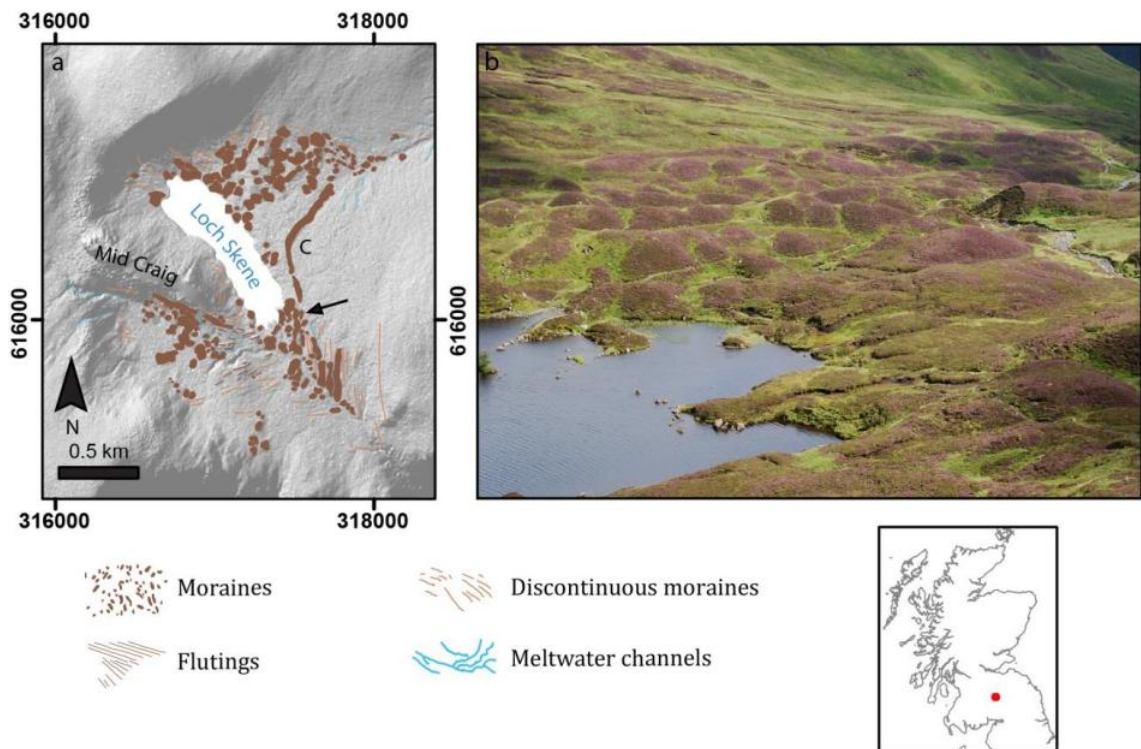
Figure 3.22 a) Geomorphological mapping of the Loch Skene area in the Tweesdmuir Hills (after Pearce et al., 2014a). C: The Causey moraine ridge. Arrow indicates those moraines photographed in Figure 22b. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. b) Recessional moraines at the foot of Loch Skene, taken from Mid Craig looking southeast. Photograph from H.L.Bickerdike.

## 3.3 Geomorphological evidence in northern England

The area of England which was glaciated during the LLS is relatively small and almost entirely lies in the Lake District region in northwest England (Figures 3.1 and 3.23), with a further few cirque glaciers being situated in the Pennines and a possible site in the Cheviots. Like the satellite icefields which flanked the West Highland Glacier Complex, interpretations of the style of glaciation in the Lake District have shifted from an alpine to a plateau icefield style of glaciation. Furthermore, this region supports a spectrum of glacial landform evidence, from sharp-crested hummocky moraine to much more subdued features. Given the paucity of dating in the region, the interpretations of different authors have important implications for glacial reconstructions. An overview of LLS glaciation in the Lake District is also available in Bickerdike et al. (2015).

### 3.3.1 Lake District

The signature of LLS glaciation in the Lake District has been recognised for almost 150 years. Ward (1873) observed that subsequent to ice sheet retreat the majority of higher valleys in the area had been occupied by ice during what he described as a 'second land-glaciation' but noted that the absence of moraines in the wider valleys suggested that ice was relatively restricted. Early studies simply described the glacial landforms (Marr,

1916), or just presented the inferred extent of the ice (Manley, 1959). The first detailed geomorphological mapping of the area was conducted by Sissons (1980a) who reconstructed a series of 64 alpine style valley and cirque glaciers in the area during the LLS (Figure 3.23). He argued that only those glaciers which produced fresh features (e.g. sharp-crested moraines) and clear limits were of LLS age and that features which did not meet these criteria must predate the LLS.

Subsequent remapping of the geomorphology has prompted a radically different interpretation of the evidence. McDougall (1998, 2001), Rea et al. (1998) and Brown et al. (2011, 2013) argued that LLS plateau icefields developed in the upland areas of the central Lake District, the most significant of which was centred on High Raise (Figure 3.23). These icefields were drained by outlet glaciers in the valleys which, in some cases, had similar extents to Sissons' (1980a) reconstructions, but in other areas were more extensive. This ice configuration explains the variations in glacier extent and the ELAs of Sissons (1980a) reconstructed glaciers, which he attributed to a combination of differential snowfall intensity and snowblow across the region (Rea et al., 1998). Ice-flow modelling further supports this plateau icefield interpretation and suggests that the maximum extent of outlet glaciers from this icefield was reached early in the LLS during the coldest conditions and was followed by a more restricted, but longer-lived, final position (Brown et al., 2013). This final position accords well with the location of moraines which have traditionally been attributed to the LLS.
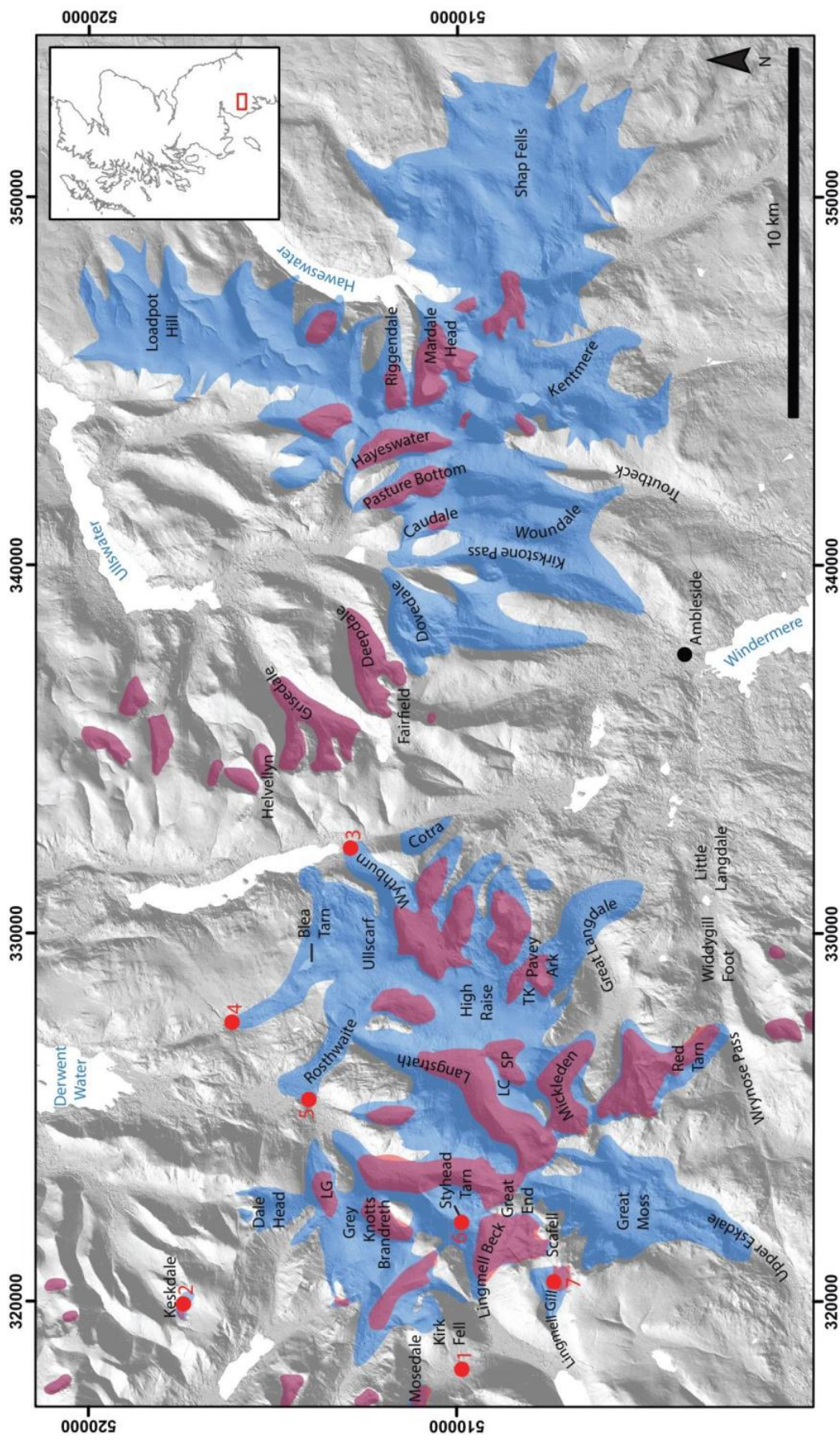
Extent of LLS glaciation proposed by Sissons (1980a)

Extent of LLS glaciation compiled from McDougall (2001, 2013), Wilson (2002, 2011), Hughes et al., (2012) and Brown et al. (2013)

Chronological sites discussed in text

Figure 3.23 The extent of LLS glaciation in the Lake District. Sissons' (1980a) proposal of alpine style glaciers having developed in these hills was challenged (Rea et al., 1998; McDougall, 2001, 2013) in favour of a plateau style of glaciation. Plateau icefields developed on the central and eastern fells, surrounded by cirque glaciers at topographically favourable sites. LG: Little Gatesgarthdale, LC: Langdale Combe, SP: Stake Pass, TK: Thunacar Knott. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

LLS glaciation in the western Lake District comprised 18 small independent glaciers, the extents of which are all marked, at least partially, by end or lateral moraines which range between 1 and 10 m in height and often enclose patches of hummocky moraine (Sissons, 1980a). Particularly clear examples are found in Mosedale (Figure 3.23) where two clear nested arcuate end moraines contrast starkly with the larger more subdued moraine mounds in the lower valley, which are thought to predate the LLS (Brown et al., 2011, 2013). Coring of a peat-filled depression in the lower valley revealed (Figure 3.23, chronological site 1) a Lateglacial stratigraphic and vegetation sequence, indicating that the lower valley was ice-free during the LLS and supporting the assertion that the clear end moraines mark the terminus of the LLS glacier (Evans et al., 2015). Likewise, cosmogenic dating of a boulder on a ridge identified as an end moraine (Sissons, 1980a; Ballantyne and Harris, 1994) of the Keskdale glacier (Figure 3.23, chronological site 2) gave an age of 12.4±1.2 ka, whilst the bedrock lip higher up in the cirque was not exposed until 11.7±1.1 ka (Hughes et al., 2012). These dates not only confirm a LLS age, but also suggest that the glacier reached its maximum position early in the stadial before retreating and stabilising and forming smaller moraines in the upper cirque. This interpretation contrasts with Sissons' (1980a) reconstruction of ice only in the lower cirque, but the presence of mudstone clasts derived from the upper cirque backwall suggests that ice was sourced here and supports the notion of a two-phase LLS (Hughes et al., 2012). However, the number of sites in the Lake District with chronological control, such as these two, is low and most of the glacial limits are dated using morphological criteria.

Glaciation was more widespread in the central Lake District and much of the region's most impressive glacial geomorphology is found in these valleys (Figure 3.24), which Sissons (1980a) believed were occupied by alpine style valley glaciers during the LLS. In light of recent advancements of glaciological theory and remapping of the geomorphology (Rea et al., 1998; McDougall, 1998, 2001; Rea and Evans, 2003), it is now thought that glaciers in the central valleys were fed by several plateau icefields, the largest of which was centred on High Raise, while smaller icefields developed on the summits of Grey Knotts, Brandreth, Kirk Fell and Dale Head (Figure 3.23). The valleys which drained these areas typically contain sequences of recessional moraines. Those in Little Gatesgarthdale form

closely-spaced chains of ridges and hummocks up to 4-5 m high, which trend obliquely down-valley (Figure 3.24). The features gradually become more subdued in their appearance, and Sissons (1980a) acknowledged that the terminus of the LLS glacier could occur anywhere in the zone of uncertainty caused by the gradual transition in the appearance of the moraines. In the absence of a clear limit, McDougall (1998, 2001) assigned a LLS age to all the moraines to reconstruct a larger glacier.

Similar recessional moraines are found in the valleys which radiate out from the summits of High Raise, Ullscarf and Thunacar Knott (Rea et al., 1998; McDougall, 1998, 2001; Brown et al., 2011, 2013). This area preserves some of the most compelling evidence for a plateau rather than alpine style of glaciation. Prominent, sometimes bifurcating, moraine ridges at Langdale Combe and Stake Pass (Figure 3.6d) indicate a complex deglaciation pattern of retreat up onto the plateau rather than to the head of the valley, whilst meltwater channels on the slopes show that ice from Langstrath probably inundated Langdale Combe (McDougall, 2001) (Figure 3.24). On the other side of the ridge a series of ice-marginal moraines at Pavey Ark confirm this pattern of retreat back onto the plateau (Rea et al., 1998). Similarly, at the head of Greenup Gill (Figure 3.24) a chaotic complex of moraines suggests localised ice stagnation, but some features indicate retreat out of the valley and onto the plateau. This evidence points towards a much more extensive plateau icefield which contrasts with Sissons (1980a) reconstruction of valley glaciers interconnected to each other only through low cols. Sub-angular and sub-rounded clasts within these moraines suggest active transportation of debris and that ice in the valleys was probably warm-based, whereas the presence of blockfields on the summits of Thunacar Knott, High Raise and Ullscarf indicates that the ice covering these areas was probably cold-based and non-erosive (Rea et al., 1998). Streamlining of bedrock exposures around the margins of these areas has been inferred to represent the area of transition between these two regimes (McDougall, 2001). Although these features have not been directly dated, coring in the area of Langdale Combe occupied by moraines, revealed that the earliest sediments were of Holocene age, supporting a LLS age for the moraines (Walker, 1965).
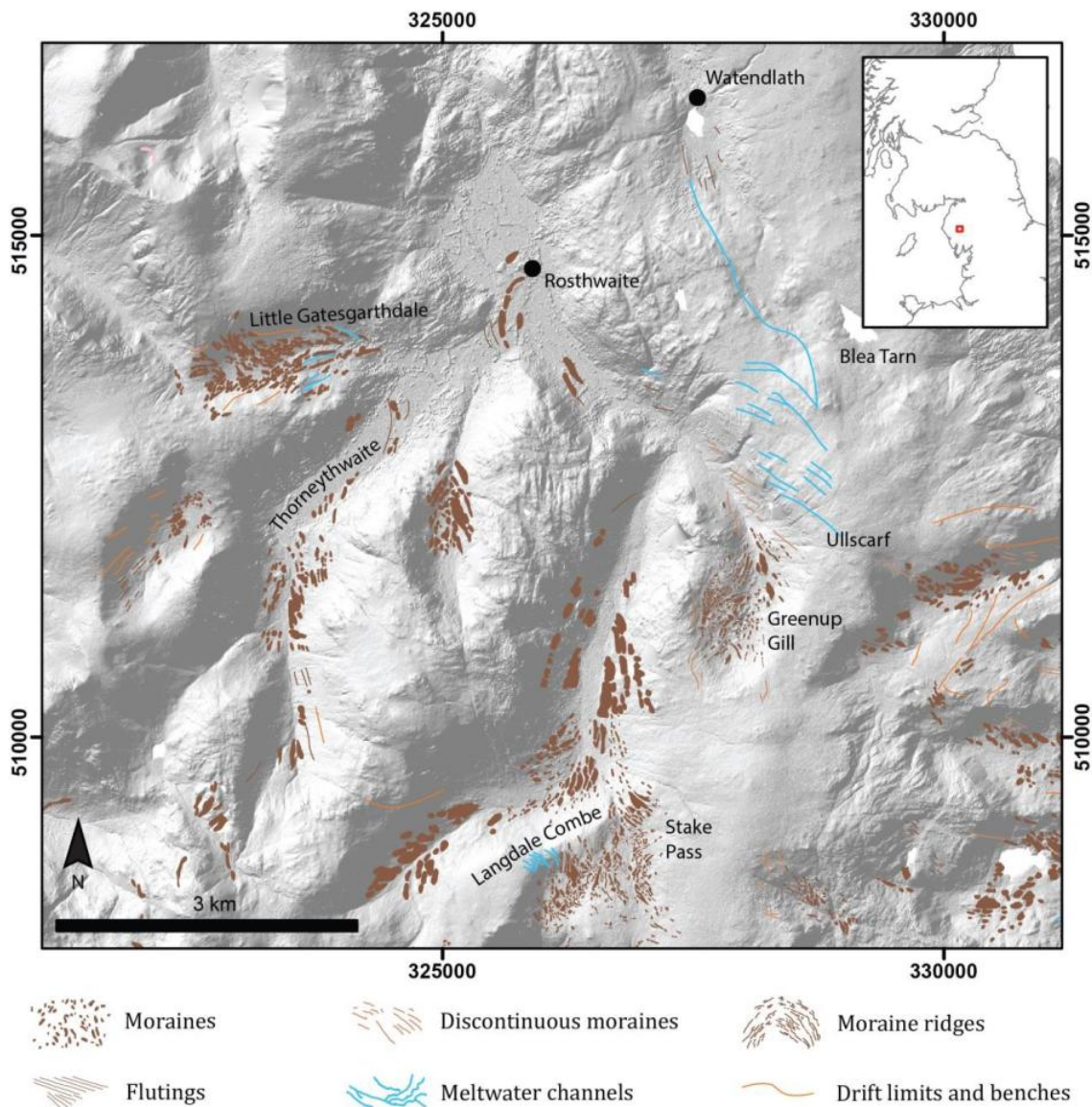
Figure 3.24 Geomorphological mapping of the central Lake District (after Sissons, 1980a; McDougall, 1998, 2001). Sequence of recessional moraines, which are preserved in many of the valleys within this area, show the retreat of ice up onto the plateau at Langdale Combe and Greenup Gill, indicating that a plateau icefield fed these valleys. The termini of the Rosthwaite and Watendlath lobes are shown. Underlying hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Further north, the evidence for the outlet glaciers of the High Raise icefield decreases in clarity. The prominent 3-4 m high moraines in Greenup Gill transition down-valley into a "moundy drift veneer" (McDougall, 2001: p534) which Sissons (1980a) believed was beyond the LLS glacier limit. The actual termini of the Rosthwaite and Wythburn lobes are clearer and have been reconstructed from end moraines, whilst the Watendlath lobe has been inferred from fragments of end moraine and an ice-marginal meltwater channel which can be traced up-valley to the meltwater channel system on the slopes of Ullscarf (Rea et al., 1998; McDougall, 2001) (Figure 3.24). Whether these features are of LLS age remains uncertain (Clark and Wilson, 1994). Pennington (1978) argued that the presence

of Lateglacial sediments at Blea Tarn (Figure 3.24) indicated that the Watendlath lobe could not have covered this area during the LLS, but it has subsequently been proposed that these sediments may have been disturbed by the breakup of lake ice (Pennington, 1996) or preserved beneath a cold-based LLS glacier (McDougall, 2001). Attempts to resolve this debate using cosmogenic isotopes are equivocal and the majority of dates from boulders on the Wythburn, Watendlath and Rosthwaite moraines (Figure 3.23, chronological sites 3, 4 and 5) indicated pre-LGM ages (Wilson et al., 2013). These ages seem implausible given that the last British Ice Sheet completely overran this area and have probably been affected by nuclide inheritance. One date at Rosthwaite and one at Watendlath fell between the LGM and LLS, suggesting that these limits might have been reached during an earlier stage of deglaciation. Considering the potential unreliability of the dates, further work is necessary to resolve this debate.

The age of moraine ridges at Cotra (Figure 3.23) to the east of the icefield have also been debated, with Manley (1959) and Sissons (1980a) arguing that the area was not glaciated during the LLS, but Wilson (2002) and Pennington (1978) arguing that it was. Rather than using dating techniques to determine the age of the glacier that formed these features, Carr and Coleman (2007a) used a glaciological model to calculate that the Cotra glacier would require an implausibly high component of basal motion and thus favoured a pre-LLS age. Wilson (2008) argued that the moraines could not have been formed by this small glacier during a period of general ice wastage but Carr and Coleman (2009) argued that the proposed glacier had an implausibly low ELA for the LLS, and that the features actually represent mass movement; as yet the age and origin of these features remains uncertain.

In the southeastern sector of the icefield, McDougall (1998, 2001) suggested that glaciers in Pavey Ark, Mickleden and Oxendale coalesced to form a large outlet glacier in Great Langdale (Figure 3.23) where McDougall (1998) noted, but did not map, the existence of subdued hummocks on the valley floor. The extent of this glacier relies heavily on a subtle drift limit on the valley side. Although Brown et al. (2013) did not dispute the terminus of the Great Langdale glacier, in their reconstruction, the Great Langdale glacier simply extends from the ice at Pavey Ark and is unconnected to the Mickleden and Oxendale lobes. In their reconstruction, the termini of these latter two glaciers were placed at the maximum of the sequences of inset lateral moraines in these valleys, similar to the positions proposed by Sissons (1980a), and the subdued mounds lower in Great Langdale were thought to predate the LLS. Similarly, Sissons (1980a) and Brown et al. (2013) identified morphological differences between the clear ridges below Red Tarn on the northern slopes of the Wrynose Pass and more subdued features in the lower valley

(Figure 3.23). Numerical modelling supports both of these proposed limits, suggesting ice was not present in either the main Wrynose Pass or western Great Langdale (Brown et al., 2013). A further glacier at Widdygill Foot in Little Langdale (Figure 3.23) has previously been afforded a LLS age (Manley, 1959; Pennington, 1978; Wilson, 2002), but no dating has been undertaken on the moraine ridges. Brown et al. (2013) argued that this site would have been unfavourable for glaciation during the LLS based on its low elevation and south facing accumulation area, an assertion that was supported by numerical modelling in their study. Given the success of this model in identifying the limits of LLS glaciation in other valleys in this area, it seems reasonable that the Widdygill moraines predate the LLS.

Further west, Brown et al. (2011, 2013) confirmed the presence of ice in Lingmell Beck and Lingmell Gill (Figure 3.23) as proposed by Sissons (1980a), but argued that these glaciers were more extensive than previously suggested based on the position of the most prominent moraines in these valleys. In the reconstruction by Brown et al. (2013), the Lingmell Beck glacier is no longer independent as proposed by Sissons (1980a), but is connected to the main icefield to the north of Great End (Figure 3.23). Although this conflicts with the presence of undisturbed Lateglacial deposits found near Styhead Tarn (Figure 3.23, chronological site 6) at the head of the valley (Pennington, 1978), Brown et al. (2013) argued that these could have been preserved beneath cold-based ice and that this configuration is most plausible. Conversely the neighbouring glacier in Lingmell Gill was reconstructed as an independent glacier, and has been dated to the LLS by a minimum exposure age of 12.5±0.8 ka (Ballantyne et al., 2009) (Figure 3.23, chronological site 7). Like Mosedale, Lingmell Beck and Lingmell Gill, moraines are present along the length of Upper Eskdale. Manley (1959) inferred that these were formed by two independent glaciers during the LLS, but Sissons (1980a) and Pennington (1978) argued that they either predate the LLS or were formed by a snowbed situated on the southeastern flanks of Scafell (Sissons, 1980a). However, remapping of the area has suggested that the area was occupied by a large glacier which terminated at an area of prominent moraine ridges near the confluence of Cowcove Beck and the River Esk (Brown et al., 2011, 2013). This glacier actively retreated up along Lingcove Beck, as indicated by a series of prominent recessional moraines, but appears to have undergone stagnation below Great Moss when, as inferred from nested lateral and hummocky moraines, it became decoupled from its accumulation centre on Scafell (Wilson, 2004) (Figure 3.23). Such a retreat pattern has been replicated by numerical modelling (Brown et al., 2013) suggesting this interpretation of the geomorphology is plausible. Although no evidence is preserved in the cirques which fed the Eskdale glacier, the steep gradient of their floors would have inhibited accumulation of glacial debris (Sissons, 1980a; Wilson, 2004), masking the signature of

glaciation. The absence of ice is quite implausible considering the favourable aspect and elevation of these cirques and the presence of LLS ice on the northwestern side of the ridge. Whilst there is no direct dating on the Eskdale moraines, because the last British Ice Sheet moved westwards over the area, the Eskdale moraines could only represent a very late stage of deglaciation (Wilson, 2004). Given the extent of the other outlet glaciers, it seems highly plausible that these features do indeed represent a substantial LLS glacier in the valley.

As with the central Lake District, understanding of the extent of LLS glaciation in the eastern Lake District (Figure 3.23) has shifted radically from an alpine to a plateau style of glaciation. Manley (1959) proposed that 10 glaciers developed in the area east of the Kirkstone Pass, the most extensive of which were four valley glaciers (Hayeswater, Riggendale, Mardale Head and Kentmere) radiating out from Racecourse Hill. Subsequently, Sissons (1980a) reconstructed 11 cirque and valley glaciers in the area (Figure 3.23), supporting Manley's (1959) limits in some areas, but with notable differences in others, including the omission of the large Kentmere glacier and its substitution with two very small niche glaciers. Since Sissons (1980a) inferred a LLS age only for glaciers marked by clear, sharp-crested moraines most discrepancies occur in areas characterised by large subdued mounds with no clear down-valley limits.

A much greater extent of ice has been inferred from recent mapping of the upland area east of the Kirkstone Pass. Like the central Lake District, moraines and drift limits can be traced from lower in the valleys up onto the high ground, including in Hayeswater (Figure 3.25), Kent Valley, Troutbeck and Caudale Beck, indicating that these glaciers were fed by ice on the plateau (McDougall, 2013). At other locations where direct evidence is lacking, plateau-based ice has been inferred from the presence of ice in all surrounding valleys (Figure 3.23). In such instances, assuming alpine style glaciers would require ice to have developed at unfavourable sites such as the south-facing valleys of the Shap Fells (Figure 3.23), a plateau icefield configuration with the valleys acting as outlet glaciers is a more plausible configuration (McDougall, 2013).
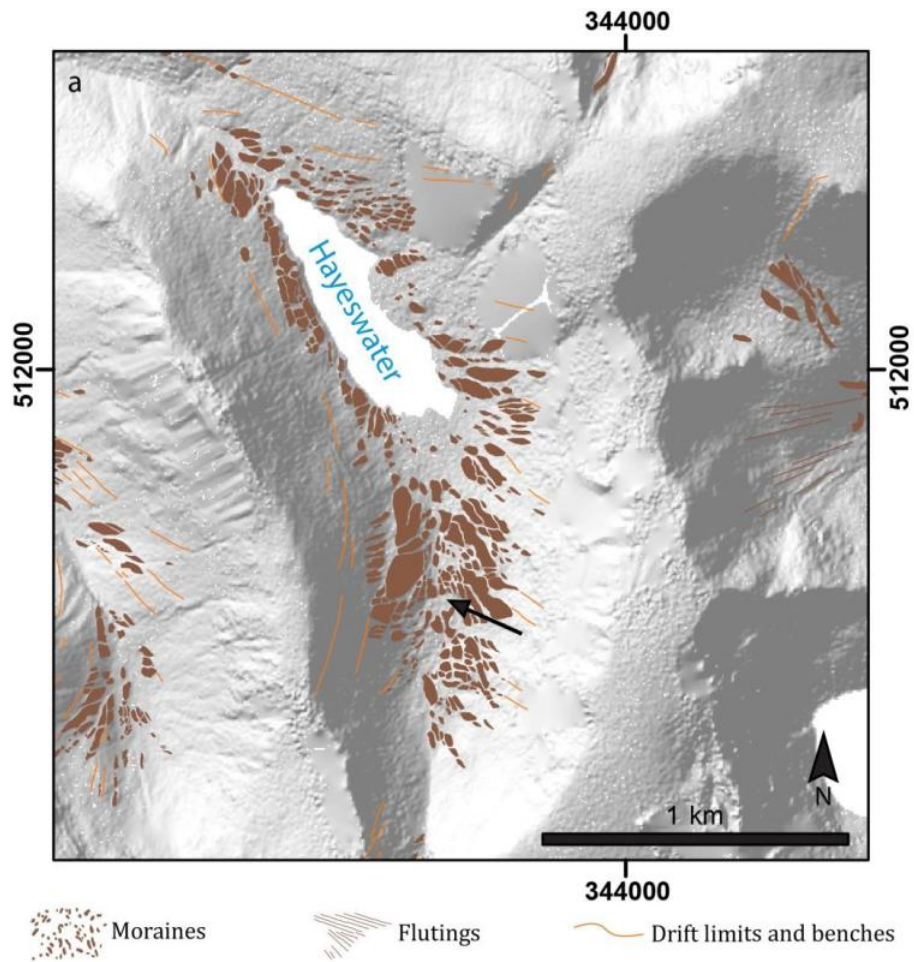
Figure 3.25 a) Sharp-crested recessional moraines at Hayeswater (after McDougall, 2013). At the valley head the moraines show glacier retreat back onto the plateau, refuting the alpine style of glaciation Sissons (1980a) proposed for this site. Figure 3.25b was taken looking south from the point indicated by the arrow. The hill-shaded image derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. b) Hummocky moraines at the head of Hayeswater valley looking north. Photograph from H.L. Bickerdike.

The reliability of using freshness of appearance to determine moraine age has been questioned (Wilson, 2002). McDougall (2013) argued that the valleys occupied by LLS glaciers exhibit one of three types of moraines. Some valleys, typically those believed by Sissons (1980a) to have nourished LLS glaciers such as Hayeswater, Pasture Bottom and Mardale Head, present sharp-crested recessional moraines (Figure 3.25). Other valleys have well-developed but smaller and less clear moraines, as in the Kirkstone Pass and at Caudale Bridge. Finally, in some valleys moraines are generally only 0.5 to 2 m high as observed in Woundale and the Kent Valley. McDougall (2013) inferred that the moraines in all these valleys were of LLS age and reconstructed an icefield that occupied the high ground and surrounding valleys from Loadpot Hill in the north to the terminus of the Kirkstone glacier near Ambleside in the south and the Shap Fells in the east (McDougall, 2013). However, given the absence of absolute dates, the subdued nature of some features and the absence of clear terminal moraines, this reconstruction was speculative.

Located between these two areas of undated reconstructions, much of the Helvellyn Range (Figure 3.23) has not been remapped since Sissons (1980a) proposed that valley glaciers had occupied Deepdale and Grisedale with a further three cirque glaciers at the head of Glenridding and another four in the hills further north. A further, valley glacier in Dovedale (Figure 3.23) was reconstructed by Wilson (2011). The size of a particularly large end moraine below Wolf Crags is at odds with the comparatively small glacier it delimits and is attributed to this glacier remaining stable at this position for a significant length of time while being nourished by snow blown onto the glacier from the plateau area to the southwest (Manley, 1959; Sissons, 1980a). This contrasts with the style of retreat represented in Deepdale by a clear series of recessional moraines and hummocks indicative of active retreat back to its source, which was very likely contiguous with the Grisedale glacier and probably part of an icefield on Fairfield (McDougall et al., 2015a).

The northernmost LLS glaciers in the Lake District occupied cirques in the Skiddaw Range (approximately 10 km north of Derwent Water). Sissons (1980a) reconstructed three small cirque glaciers on the eastern flanks of the Blencathra hills, at Bannerdale Crags, Scale Beck and at Bowscale Tarn. The Bowscale glacier is remarkable for the large end moraine which marks its terminus, which reaches 20 m above the lake on its distal side and up to 40 m above the slope at the cirque mouth, and is littered with boulders (Evans, 1994). Like the Wolf Crags end moraine, the size of this feature is attributed to a small but stable glacier that was nourished by snowblow from the plateau above and formed the ridge from abundant debris left in the cirque during ice sheet deglaciation (Evans, 1994). In the neighbouring Ling Thang, the glacier is represented by two fragments of end

moraine, the incomplete nature of which may be due to erosion by the River Caldew (Clark and Wilson, 2001). Three further small cirque glaciers are proposed to have developed in sheltered hollows at Great Cockup, Nine Gills Comb and between Dale Beck and Yard Steel (Wilson and Clark, 1999). The sites of these three glaciers all have favourable aspects, facing between north and north east and would have received snow from plateau areas to the southwest. Although potential sites of niche glaciers have been identified, such as at Glenderamackin Cove (Evans, 1994), these remain unmapped.

The Lake District is extremely significant in the study of LLS glaciation in Britain as the first region where a plateau style of glaciation was recognised, a style that has since been identified for other LLS icefields in Britain (Boston, 2012a; Boston et al., 2013; Pearce et al., 2014b). Detailed geomorphological mapping for much of this region and numerical modelling (Brown et al., 2013) strongly supports this interpretation for the central Lake District, whilst new mapping of the eastern hills is also in agreement with this style of glaciation. The Helvellyn Range remains largely unmapped since Sissons' (1980a) original study and it remains to be seen what extent and style of glaciers would be proposed if the area were to be fully remapped. Establishing chronological control of glacial features remains problematic in the Lake District. Few sites suitable for radiocarbon dating have been found and cosmogenic isotope dating seems to be particularly susceptible to nuclide inheritance in this region (Ballantyne et al., 2009; Wilson et al., 2013). Given that landform freshness no longer appears to be a reliable indicator of age, some uncertainty will be attached to the limits of LLS glaciation until these challenges are met.

### 3.3.2 Pennines and Cheviots

In contrast to the Lake District, LLS glaciation in the English Pennines (Figure 3.1) has been the focus of considerably less research and was largely ignored until the 1990s. The first account of the LLS in the area was by Rowell and Turner (1952), but their interpretation of ridges as lateral and terminal moraines was challenged by Manley (1959), who reclassified them as snowbed features. Geomorphological mapping by Mitchell (1996) indicated that five cirque glaciers formed in the western Pennines during the LLS, at Great Coum, Swath Fell, Combe Scar, Whernside and Cautley Crags. These were inferred primarily from a combination of terminal and lateral moraines. Similarly, at Cronkley Scar the presence of a 1.3 km long, 32 m high slightly arcuate depositional ridge has been interpreted as a moraine created by a small but stable glacier, much like those at Bowscale and Wolf Crags in the Lake District (Wilson and Clark, 1995). A unifying feature of all six glaciers is their location downwind of large areas of potential snowblow onto the glaciers, which explains their restriction to topographically favourable sites in this glaciologically-marginal region (Mitchell, 1996). Although no direct dates for these

landforms are available, the presence of only Holocene aged sediments from inside the glacier limit at Combe Scar support a LLS age for these glaciers (Gunson, 1966).

In the Cheviots (Figure 3.1), the only site proposed to have nourished a glacier during the LLS is the deep, north-facing Bizzle Cirque where a series of sharp-crested moraine ridges contrast with more subdued forms beyond them (Harrison et al., 2006). Like the Pennines sites, the Bizzle Cirque is situated downwind of a large plateau area and it seems highly likely, given the absence of LLS ice from elsewhere in the Cheviots, that snowblow from this plateau was essential for the development of the Bizzle Cirque glacier. The absence of well–developed periglacial features, including mature talus slopes, frost-shattered bedrock and solifluction lobes, from within the limits of the proposed glacier, but their abundant presence outside, was inferred to show that this area was glaciated during the LLS (Harrison et al., 2006).


## 3.4 Geomorphological evidence in Wales

Unlike Scotland, LLS glaciation in Wales has not received such intensive study and, whilst features associated with this period were first recognised over 150 years ago, it is only recently that a more systematic approach to mapping the landform evidence pertinent to the stadial has been developed. It is generally agreed that, during the LLS, glaciation in Wales was restricted to upland sites, particularly Snowdonia and the Brecon Beacons, where cirque glaciers developed at favourable sites (Carr, 2001; Hughes, 2002, 2009; Coleman and Carr, 2008; Bendle and Glasser, 2012) (Figures 3.26, 3.27 and 3.28). The limited extent of these glaciers means that sequences of moraines are relatively uncommon in Wales, with many glaciers being represented only by terminal ridges lying short distances from the cirque backwalls. Consequently, much of the literature is concerned with distinguishing whether these landforms are moraines formed by glaciers or protalus ramparts, the latter indicating only the existence of semi-permanent snowbeds rather than glaciers (Unwin, 1970; Shakesby and Matthews, 1993; Carr et al., 2007a).

### *3.4.1 Snowdonia and northern Wales*

Evidence of LLS glaciation in Snowdonia (Figure 3.26) consists primarily of well-preserved moraines in the high cirques. These features have long been of interest to those studying Snowdonia's glacial history and from the mid-19th century were used to inform developing glacial theory in Britain (Darwin, 1842). However, it was not until 100 years later that Seddon (1957) mapped the overall distribution of 33 'end moraines' within the area, albeit at a low resolution. Unwin (1970, 1975) subsequently mapped 52 features, again at a low resolution, but split these landforms into two classifications; 11 belonging to an Older

Series (low, degraded features with diffuse outlines and gentle slopes) and 41 to a Younger Series (fresher forms, with higher and steeper slopes and clearer outlines), of which 19 were sub-classified as protalus ramparts. Despite these minor differences, the overall distribution of Unwin's (1970, 1975) Younger Series glaciers showed the same trends of predominantly north-easterly aspect and an eastward increase in altitude, supporting Seddon's (1957) assertion of the importance of insolation, windblown snow and uneven precipitation in influencing glacier formation. Detailed and systematic geomorphological mapping was first conducted by Gray (1982), who mapped evidence for 35 former glaciers (including end moraines, drift and boulder limits and the extent of hummocky moraine) and a further 16 semi-permanent snowbeds, inferred from the presence of protalus ramparts. Recent mapping by Bendle and Glasser (2012) generally agreed with the results of Gray (1982), but included evidence for an additional three glaciers beyond the original study area, based upon the use of digital terrain models to achieve a more detailed representation of the morphology and distribution of moraines (Figure 3.26).

Little absolute dating of landforms has been conducted in Snowdonia and most of the relative dating that has been done has predominantly used stratigraphic and pollen analysis to compare sediments from sites inside and outside the suggested limits. Seddon (1962) found classic Lateglacial tripartite sequences of sediment in two lake basins beyond the cwm moraine limits, at Llyn Dwythwch (2 km west of Figure 3.26) and Nant Ffrancon (Figure 3.26), but only Flandrian sediments, showing a single transition from cold to warm conditions, from inside the limits at Cwm Idwal and Cwm Cynghorion (approximately a kilometre west of Afon Arddu, Figure 3.26). Similarly, Lateglacial sediments were found in lake basins from outside the limits at Capel Curig (Crabtree, 1972) but only Flandrian sediments were found within the moraine limits at Llyn Clyd, Llyn Glas and Melynllyn (Evans and Walker, 1977; Walker, 1978) (Figure 3.26). At Cwm Llydaw and Cwm Cywion (Figure 3.26, chronological sites 1 and 2 respectively) the onset of organic sedimentation was radiocarbon dated to about 10,000 [14]C a BP (Ince, 1981, 1983), indicating a LLS age for these glaciers. Taken together, this body of evidence makes a LLS age for the moraines highly likely. This assertion is further supported by cosmogenic isotope dating of two boulders on the crest of the moraine damming Llyn Idwal (Figures 3.26, chronological site 3, and 3.27), which gave ages of 12.9±2.0 and 11.6±1.3 ka, indicating that the full moraine complex is likely to be of LLS age (Phillips et al., 1994).
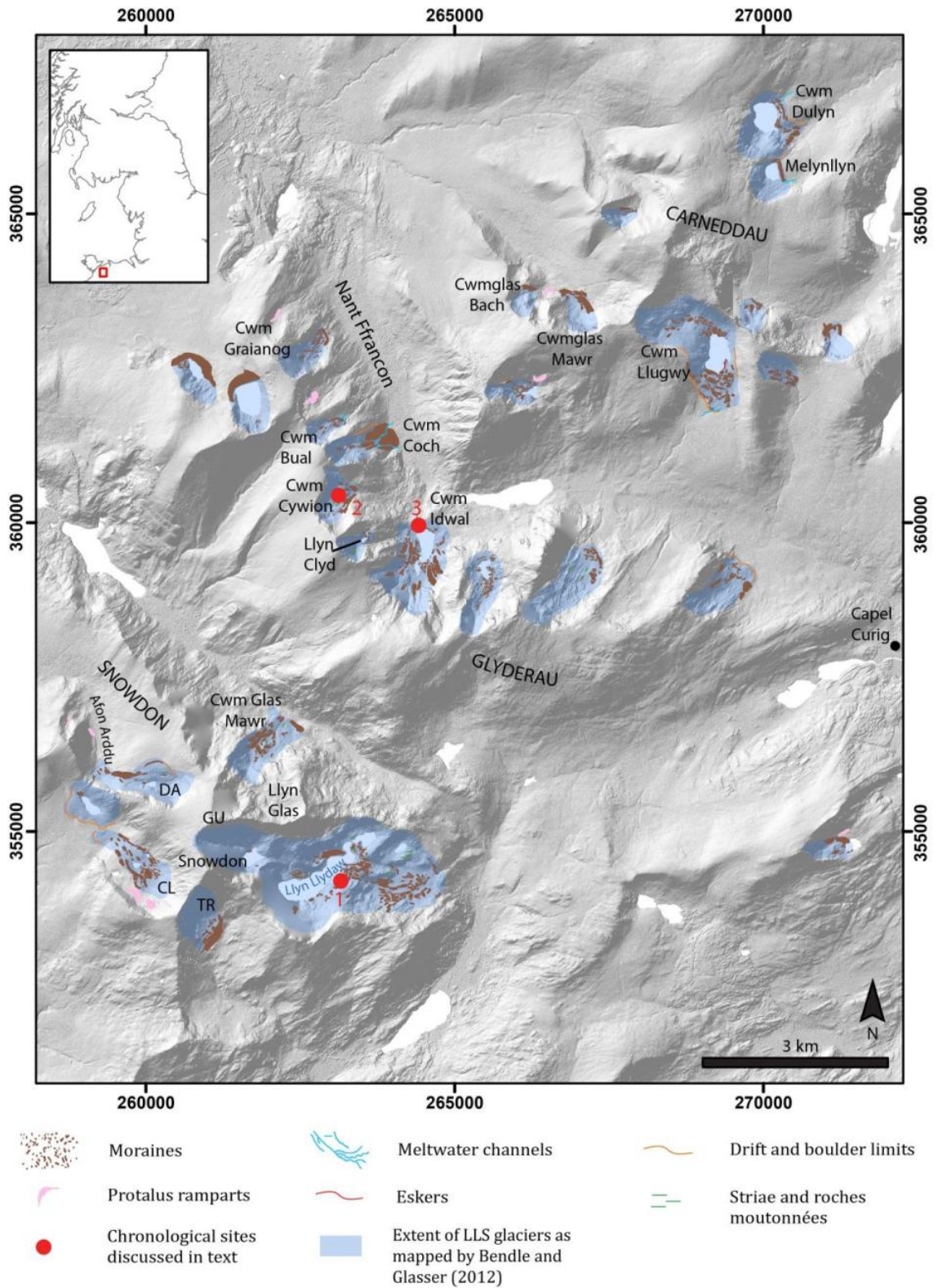
Figure 3.26 The LLS glacial geomorphology of Snowdonia and reconstructed ice extents (after Bendle and Glasser, 2012). DA: Cwm Du'r Arddu, GU: Garnedd Ugain, CL: Cwm Clogwyn, TR: Cwm Tregalan. Underlying hill-shaded image derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The northernmost LLS glaciers in Wales occupy cwms in the Carneddau massif of Snowdonia and are variable in size and aspect, including two of only four glaciers in Snowdonia with area over 1 km$^2$ (Bendle and Glasser, 2012) (Figure 3.26). The largest of these at Cwm Llugwy has a south facing aspect, the extent of which is inferred from recessional, sometimes chaotic moraine ridges within a drift and boulder limit (Gray, 1982). Although these moraines extend into the upper cirque, almost reaching the backwall, the majority of cirques within Snowdonia contain recessional moraines that are concentrated in a belt near the former glacier terminus. In some instances, the terminus is marked by relatively large and distinct terminal moraines, such as the remarkably straight moraine that dams the lake at Melynllyn (Gray, 1982) (Figure 3.26). A further notable feature of several of the Snowdonia cwms is the presence of ice-moulded bedrock in the accumulation areas (as at Cwm Dulyn, Cwmglas Bach and Cwmglas Mawr (Figure 3.26)) and the presence of protalus ramparts at sites that were unfavourable for glacier formation, but that nourished snowbeds during the LLS (Bendle and Glasser, 2012).

The variability in size of the Carneddau glaciers contrasts with the ten small cirque glaciers reconstructed along the northeast face of the Glyderau massif, the outermost moraines of which rarely exceed a kilometre in distance from the cirque backwall (Bendle and Glasser, 2012) (Figure 3.26). The relatively consistent aspect and elevation of these cirques has been attributed to a combination of climatic control and geology. The north-easterly facing glaciers would have been protected from insolation and nourished by windblown snow from the plateau to the southwest whilst the weak mudstone floors of the cirques, combined with resistant igneous side and backwalls would have acted as geological controls on the size and shape of the cirque (Addison, 1997). Of these cirques, Cwm Idwal is perhaps the most famous and intensively researched site in Snowdonia (Seddon, 1962; Phillips et al., 1994; Addison, 1997; Bendle and Glasser, 2012), and the landforms here have proved important in advancing understanding of Britain's glacial history, with Darwin (1842) having recognised their glacial origin from as early as the mid-19th century.

A series of recessional moraines cover the valley floor for approximately a kilometre, the outermost damming Llyn Idwal (Bendle and Glasser, 2012) (Figure 3.27a). The outermost moraines are more subdued and, as such, were previously inferred to belong to an earlier phase of glaciation (Seddon, 1962; Addison, 1997), a premise now refuted by the LLS surface exposure dates collected by Phillips et al. (1994). The moraines in the inner cirque are much more pronounced in appearance, those on the east side of the lake being up to 12 m high and 8-50 m in length, whilst those on the west side are larger, up to 15 m high

and 25-80 m long, with the westernmost feature being a 450 m long, sharp-crested ridge (Graham and Midgley, 2000) (Figure 3.27a, feature X). Unwin (1975) believed that the long ridge is a protalus rampart with moraines lying below, whereas Seddon (1957) argued that the ridges are ice-marginal lateral moraines. Gray (1982) offered a third interpretation of the features as subglacial flutes. More recently, it has been suggested that the ridges are frontal moraines from an Idwal glacier that was partly nourished by an icefall from the overhanging Cwm Cneifion and deflected towards the western side of Cwm Idwal (Addison, 1997) (Figure 3.27a). Based on their morphology and sedimentology, and their similarity with features observed at ice margins in Svalbard, Graham and Midgley (2000) argued that the ridges were formed by sediment entrained in englacial thrusts within a composite Idwal-Cneifion glacier, although they acknowledged the difficulty in substantiating this idea. Whilst it seems unlikely that Cwm Cneifion did not feed the Idwal glacier, the orientation of ridges beneath the backwall, and above the large ridge on the western lake shore, suggest retreat towards the back of the cwm and not exclusively back to Cwm Cneifion.

The geomorphology of the cirques of the Snowdon massif suggests that LLS glaciers in this area were far more variable in their aspect and size than those at Glyderau, with the summits of Snowdon and Garnedd Ugain being almost entirely surrounded by ice readvancing in the cirques (Gray, 1982). These include the Llydaw glacier (Figure 3.27b), the largest LLS glacier reconstructed in the area, which accounted for 4.48 km$^2$ of the cumulative area (20.74 km$^2$) of the 38 Snowdonia glaciers (Bendle and Glasser, 2012). It is believed that this glacier occupied the composite cirque floor staircase, with ice forming in the deep basin of Cwm Glaslyn and feeding an icefall that flowed down into the broader Llydaw Valley, finally terminating at the lip of Cwm Dyli (Addison, 1997; Gray, 1982; Bendle and Glasser, 2012) (Figure 3.27b). The evidence at this site is complex. Glaslyn is dominated by ice-moulded bedrock and glacially transported boulders, with moraines only becoming frequent at the eastern end of Llyn Llydaw, where a large, steep-sided ridge dominates the northern lake shore (Figures 3.6e and 3.27b, feature Y). This feature was interpreted as a recessional moraine formed during deglaciation, when ice in Llydaw re-stabilised after separating from the Cwm Dyli ice further down-valley (Gray, 1982). The size of the feature suggests that the Llydaw glacier remained stable in this position for some time. Analysis of glacial erosional features from this site suggests that the LLS glacier had limited erosive power, having a low sliding velocity, low clast-bed contacts, carrying little englacial debris, and that its flow was topographically constrained (Sharp et al., 1989). In the absence of a distinct terminal moraine in Cwm Dyli, the extent of the Llydaw

glacier has been inferred from the presence of boulder-strewn, recessional lateral moraines (Gray, 1982).
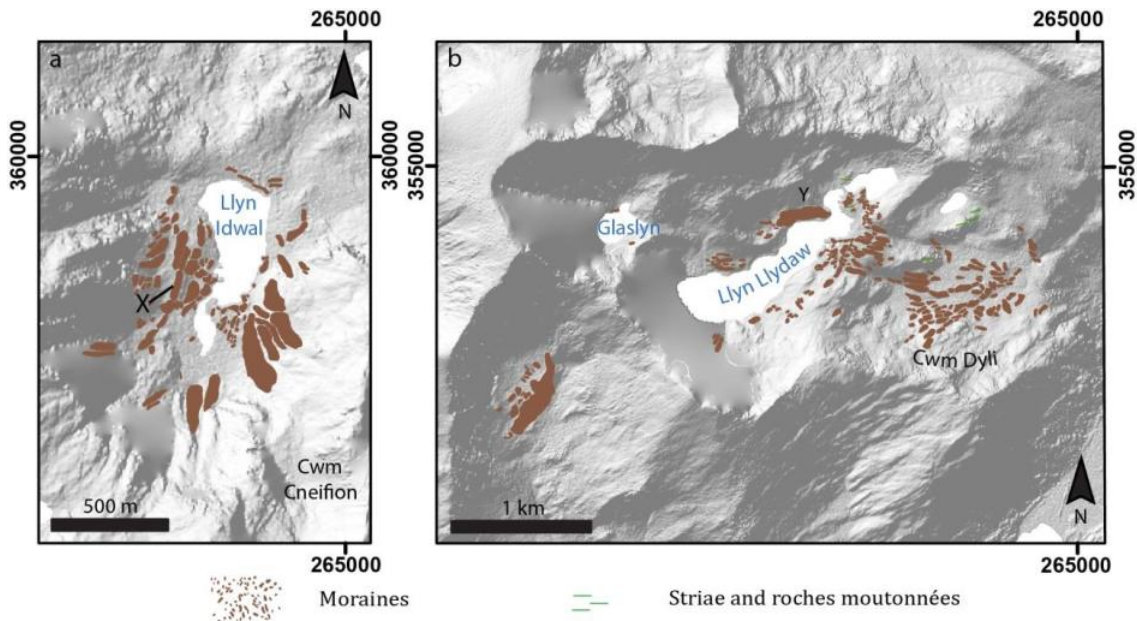


Figure 3.27 Examples of the geomorphology of Snowdonia (after Bendle and Glasser, 2012). a) Recessional moraines in Cwm Idwal.  b) Moraines formed by the Llydaw glacier. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

The remaining glaciers of the Snowdon massif are much smaller in area, the next largest, at Cwm Du'r Arddu, extending only 0.7 km² (Bendle and Glasser, 2012). Evidence in this northwest facing cirque (Figure 3.26) comprises a boulder field which contrasts sharply with peat-covered slopes outside the limits. The margins of this glacier are represented further down-valley by a partly arcuate moraine which becomes a series of straighter ridges running parallel to the slope. Some discrepancy exists between the glacial reconstructions for this site. Bendle and Glasser (2012) mapped some smaller moraines slightly beyond Gray's (1982) outermost limits, indicating that the terminus of the glacier curved north along the Afon Arddu. Further evidence of LLS glaciation, predominately moraines, is found in the larger central cirques around Snowdon (Cwm Clogwyn, Cwm Tregalan and Cwm Glas Mawr) and in the smaller northeast facing cirques that flank the Nant Ffrancon glacial trough (Cwm Graianog, Cwm Bual and Cwm Coch) (Bendle and Glasser, 2012) (Figure 3.26).

Whilst the LLS glaciers of Snowdonia were the most extensive in North Wales, other cirques in the mountainous areas peripheral to Snowdonia were glaciated during the LLS. Research into these glaciers is somewhat limited. Hughes (2002) reconstructed three small cirque glaciers in the Arenig mountains, the limits of two of which (Cwm Gylchedd

and Llyn Arenig Fach) were inferred from drift and boulder limits, and the third (Llyn Arenig Fawr) from a combination of drift limits and a broad end moraine. It should be noted that it was not possible to georeference the mapping for the Cwm Gylchedd site into Bickerdike et al.'s (2016) database with sufficient accuracy and thus these features could not be included. Hummocky moraines are present within the limits of all three glaciers whilst periglacial features are absent, strongly indicating a LLS age, especially when considered alongside the presence of only Flandrian sediment from a basin within the Cwm Gylchedd limits (Hughes, 2002). Furthermore, stratigraphic and pollen analysis of sediments from Llyn Arenig Fach showed an absence of pre-Loch Lomond Stadial sediments, again supporting a LLS age for the phase of glaciation (Lowe, 1993). A further six glaciers in the Aran and Berwyn Mountains have been reconstructed (Hughes, 2002, 2009). The evidence for these glaciers consists predominately of sediment ridges and mounds (interpreted as moraines) formed of clast-rich muddy diamicton with a high proportion of sub-rounded and striated clasts (Hughes, 2009). Of these nine cirque glaciers, the largest (Llyn Arenig Fawr) is only 0.92 km², the next largest being just 0.394 km² (Hughes, 2002). All have aspects between north-northeast and east and many lie in the lee of ridges which would have contributed windblown snow onto the glacier surface, for example at Cwm Dwygo, where snowblow contributed to a much lower ELA than the average for LLS glaciers in North Wales (Hughes, 2009).

Further south, the massif of Cadair Idris is also thought to have been occupied by cirque glaciers during the LLS (Watson, 1960; Lowe, 1993; Ballantyne, 2001; Sahlin and Glasser, 2008). There is some disagreement about whether the northern escarpment was occupied by four small LLS cirque glaciers, forming a series of end moraines, and whether an extensive glacier formed during the LLS in the cirque of Cwm Cau (Lowe, 1993; Sahlin and Glasser, 2008). Lowe (1993) interpreted the boulder-strewn lobe to the west of Cwm Gadair as representing an ice-cored rock glacier that terminated in arcuate ridges. Conversely, Sahlin and Glasser (2008) suggested that the lobe comprises a series of recessional moraines (similar to those produced by the main section of the glacier), inferring that an actively retreating, albeit probably debris-covered, glacier existed at the site. There is also discrepancy concerning the extent of the Cwm Cau glacier. Lowe (1993) proposed that the outermost moraines lie approximately 200 m from the lake, whereas Sahlin and Glasser (2008), whilst presenting no glacier reconstruction, mapped moraines stretching down-valley for a further kilometre. Ballantyne (2001) reported that this much more extensive limit is marked by a clear drift limit and large drift ridge, inside of which significantly less weathering was shown to have occurred and Larsen (1999) suggested that this area was probably not exposed to severe periglacial conditions during the LLS

and thus was glaciated. Thus, there is a lack of dated evidence for the Cadair Idris cirque glaciers, although cores from Llyn Cau and Llyn Gadair were found to have only Holocene sediments and, therefore, the assumption of a LLS age seems reasonable (Lowe, 1993).

### 3.4.2 Brecon Beacons

The Brecon Beacons are of particular interest when studying the LLS in Britain as they form the southernmost location where glaciers formed (Shakesby and Matthews, 1993) (Figures 3.1 and 3.28). Environmental conditions at this location were only very marginally suitable for glacier development (Ellis-Gryffydd, 1977) and, consequently, landform evidence consists mainly of a collection of single depositional ridges at the foot of the predominantly north-facing sandstone escarpments (Figure 3.28), the somewhat ambiguous origins of which have been the focus of research for over a century (Reade, 1894; Lewis, 1970; Shakesby and Matthews, 1996; Carr, 2001). Uncertainties about the mode of formation for these features are compounded by a lack of absolute dating.

Based on the distribution of glacial lineations in the area, Jansson and Glasser (2008) argued that some of the depositional ridge features in the valley heads (namely Fan Hir and Cwm Llwch) were not formed during the LLS niche glaciers, but by regional ice sheet flow that was deflected into the valleys during the LGM. However, where these features have been dated, such as at Craig y Fro and Craig Cerrig-gleisiad, it was consistently found that only Holocene and younger sediment was present within the limits (e.g. Preston et al., 2007; Walker, 2007a, b). Furthermore, Shakesby and Matthews (2009) argued that the morphology of the features at Fan Hir and Cwm Llwch (Figure 3.28a and c) precludes formation by LGM ice deflected up the valleys, and that the valley heads have been modified by glacial action. Lewis (1970) proposed a different chronology, believing that two phases of cirque glaciation had occurred in the Brecon Beacons, similar to Unwin's (1975) Older and Younger Series in Snowdonia, and that during the second, taken to represent the LLS, only three sites (Craig Cerrig-gleisiad, Cwm Llwch and possibly Craig y Fro) nourished glaciers. As with Snowdonia, re-evaluation of many of the sites has since shown that the landforms in the Brecon Beacons relate to a single phase of cirque glaciation (Ellis-Gryffydd, 1977), and thus a more extensive LLS glaciation in this area seems probable.
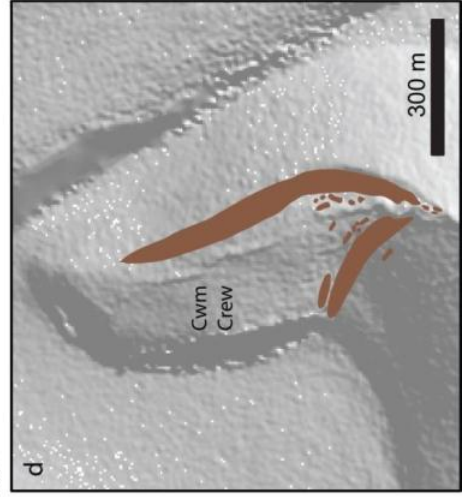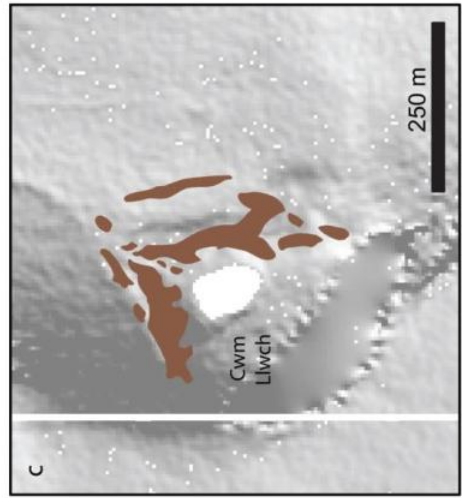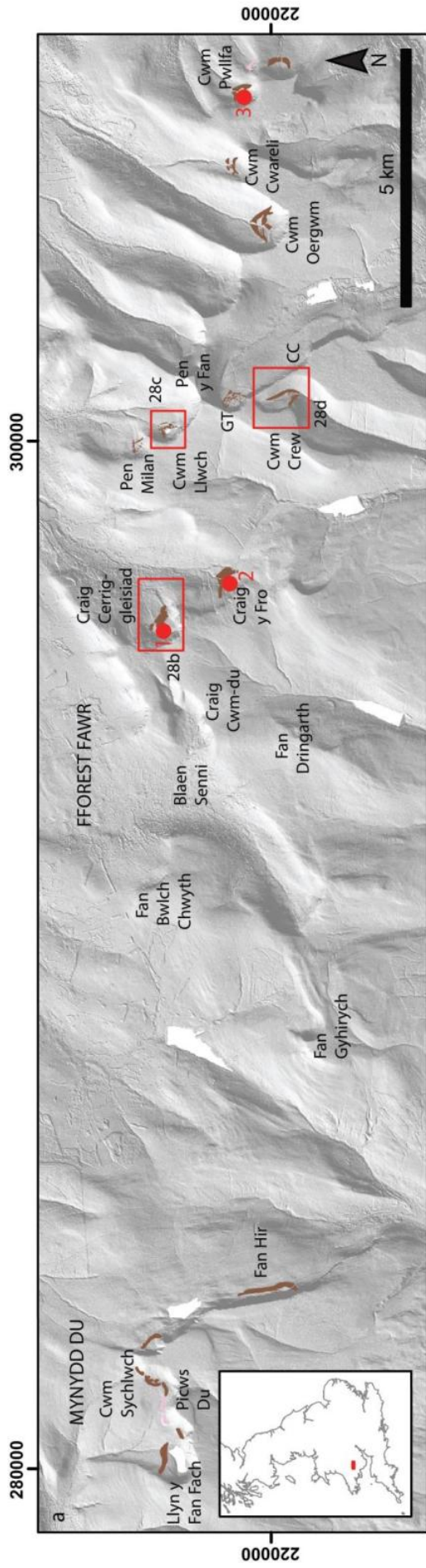
Figure 3.28 a) The glacial geomorphology of the Brecon Beacons (after Shakesby and Matthews, 1993, 2007a, 2007b; Carr, 2001; Shakesby 2007; Carr and Coleman, 2007a; Coleman and Carr, 2008). Legend in Figure 3.26; brown features are moraines, pink features are protalus ramparts and red dots represent chronological sites discussed in the text. The style of glaciation in this marginal region is one of cirque and niche glaciers and permanent snowbeds. GT: Cwm Gwaun Taf, CC: Cwm Cul. b) Landforms at Craig Cerrig-gleisiad, comprising the inner ridges (X) thought to be LLS moraines and the debris tongue (Y) argued to be a landslide deposit. Imagery from Google Earth Pro, imagery date 7/12/2013. c) End moraine ridges at Cwm Llwch. d) The well-defined terminal moraine in Cwm Crew. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Along the foot of the 7 km long, continuous scarp of Mynydd Du (the Black Mountain) (Figure 3.28a) in the western Brecon Beacons, lies a collection of depositional ridges, the origins of which have been widely debated (Ellis-Gryffydd, 1972; Shakesby and Matthews, 1993; Carr and Coleman, 2007a). The most prominent of these comprises a 1.2 km long ridge running parallel to the foot of the east-facing section of the escarpment at Fan Hir (Figure 3.28a). The ridge grades from 2-3 m high mounds at its northern end into a sharp-crested 25 m high ridge and then into a broader feature with peat-filled hollows, before again forming a sharp-crested ridge, which increases in height as it curves towards the scarp (Shakesby and Matthews, 1993). Two much smaller ridges, of no more than 2 m in height, occupy the gully between the ridge and the scarp. Previously thought to be a protalus rampart because of its linearity (Ellis-Gryffydd, 1972), subsequent studies (Shakesby and Matthews, 1993; Carr and Coleman, 2007a) have suggested that a glacial origin is more plausible. A combination of the presence of abraded and striated clasts within the ridge sediment, the curvature of the ridge at its southern end, the inner subsidiary ridges bearing a resemblance to small recessional moraines, all support the interpretation that the Fan Hir feature was formed at the margin of a small, locally nourished glacier, rather than it representing a lateral ice sheet moraine as hypothesised by Jansson and Glasser (2008). The volume of sediment in the ridge is such that an implausibly high rate of erosion would have been necessary for the exposed section of the headwall to produce sufficient debris if the ridge were a protalus rampart. Additionally, the depression behind the ridge provides enough depth of snow to accumulate such that glacier ice would very probably have formed (Shakesby and Matthews, 1993) and glaciological modelling confirmed that a glacier of the inferred dimensions would have been viable under LLS conditions (Carr and Coleman, 2007a; Shakesby et al., 2007).

Difficulties in determining the origin of such ridges has led to uncertainty when interpreting several features along the foot of the north-facing section of the Mynydd Du escarpment (Figure 3.28a). For instance, the easternmost feature in Cwm Sychlwch comprises a small, subdued arcuate ridge which has been interpreted as a pronival

rampart (Ellis-Gryffydd, 1977; Shakesby, 2002), a rock glacier deposit (Shakesby, 2002) and a moraine (Carr et al., 2007a). Ridges have usually been identified as glacial where they contain striated clasts, have an arcuate planform, and are positioned far enough from the backwall that a sufficient depth of snow could have accumulated behind them to form glacier ice (Shakesby and Matthews, 1993; Shakesby, 2007). Glaciological and energy-balance modelling has been used in some studies to confirm whether or not the configuration of ice as indicated by these ridges would be plausible under LLS conditions (Carr et al., 2007a). Based on these characteristics, two ridges below Picws Du are not thought to be of glacial origin (Shakesby, 2007). Indeed, a reconstruction of the two potential Picws Du glaciers was unable to produce credible ice dynamics under LLS conditions (Carr et al., 2007a), refuting a glacial origin. The absence of glaciers from this site is supported by the significantly greater maturity of talus at this site compared to Llyn y Fan Fach further west, with Curry et al. (2007) suggesting that while the latter was glaciated during the LLS, the former remained exposed to severe periglacial conditions. Whilst the easternmost of these is very likely a protalus rampart (Shakesby, 2002; Shakesby, 2007), the other is positioned adversely for snow accumulation and, as such, it has been suggested that the feature is a rock glacier, the formation of which would be facilitated by the absence of thick snow accumulation and the availability of talus (Shakesby, 2002), or a landslide. However, the absence of inner ridges (Shakesby, 2002), and the nature of the ridge in being separate from the infill, conflict with the rock glacier and landslide explanations, respectively (Shakesby, 2007).

Similar difficulties are associated with identifying the origin of depositional ridges in the Fforest Fawr area to the east (Figure 3.28a). This area has received less attention in the literature and mapping suitable for compilation into the GIS compiled by Bickerdike et al. (2016) was only available for two small glaciers. Small glaciers were proposed to have existed at the foot of the scarp at Fan Gyhirych and at three sites (Blaen Senni, Craig Cwm-du and Fan Bwlch Chwyth) in the Senni Valley (Shakesby, 2002). However, despite descriptive accounts of these landforms, there is a lack of detailed and systematic mapping of these features.

The most extensively studied site in the Fforest Fawr area is Craig Cerrig-gleisiad (Figure 3.28b), which provides an insight into the relationship between glacial and paraglacial activity in the Brecon Beacons. The landforms beneath the L-shaped crags have previously been interpreted as belonging to two distinct phases of glaciation within the cirque, although the precise timing of those phases has been disputed (Lewis, 1970; Ellis-Gryffydd, 1972). On the cirque floor, closest to the scarp, is a complex of ridges (between

10 and 20 m high) aligned northwest to southeast, beyond which a 1 km long tongue of deposits extends out of the cirque (Shakesby and Matthews, 1996; Shakesby and Matthews, 2007a) (Figure 3.28b). Despite having distinct ridges around the lateral margins of this feature (a 7 m high single ridge along the north side and a series of up to 10 m high mounds along the southern side) the interior has a subdued relief.

Three potential origins for these features have been proposed. The contrasting morphology, sedimentology, and extent of the two clusters of ridges led Lewis (1970) to believe they represented two phases of cirque glaciation. Whilst variable snowblow from the plateau southwest of the cirque could have produced the observed differences in glacier extent, this would be unusual for a glacier with such a small accumulation area. Nor could the tongue of deposits have been formed during ice sheet wastage, as argued by Ellis-Gryffydd (1972), as its morphology suggests movement away from the scarp. Given that the inner ridges were found to contain significantly more rounded and striated clasts than the tongue deposits, it seems reasonable to assume a glacial origin for the inner ridges, but not for the outer features (Shakesby and Matthews, 1996; Shakesby, 2002). Given its similarity to the landslide at nearby Fan Dringarth (Figure 3.28a) it is likely that the tongue at Craig Cerrig-gleisiad represents a landslide triggered by debuttressing of the scarp during ice sheet deglaciation (Shakesby and Matthews, 1996; Shakesby and Matthews, 2007a). The deposits were then modified in the inner cirque by a small cirque glacier (Shakesby and Matthews, 1996; Shakesby, 2002). Sediment cores from the peat bog enclosed by the moraines indicated that a transition from minerogenic sediments occurred at approximately 12.8 ka cal BP (Walker, 2007a) (Figure 3,28, chronological site 1). Matching of the biostratigraphic horizons at this site with those in Holocene sediments from other locations in Wales, suggested that this date is around 800 years too old, probably due to contamination with older carbon, whilst the pollen sequence supported a LLS age for this limit (Walker, 2007a). The contribution of windblown snow from the plateau to the southwest was almost certainly an important factor in nourishing this glacier and may have accounted for up to 45% of accumulation (Carr and Coleman, 2007b).

At Craig y Fro (Figure 3.28a) the origin of a prominent ridge is uncertain. The inner ridge is up to 10 m high but more subtle ridges are visible beyond it and a glacial origin for this feature is supported by the presence of abraded and striated clasts (Shakesby and Matthews, 2007b). Additionally, the distance from the headwall and the absence of evidence of headwall instability seem to discount pronival rampart or landslide origins respectively (Shakesby and Matthews, 2007b). The cirque is in a favourable location to

have accumulated windblown snow from the neighbouring plateau (Shakesby, 2002; Shakesby and Matthews, 2007b), and sediment cores taken proximal to the ridge revealed the onset of organic sedimentation between 11,247-11,838 cal BP (Walker, 2007b) (Figure 3.28, chronological site 2). Although this evidence supports this feature having been formed by a LLS aged cirque glacier, Carr et al. (2007b) argued that the reconstructed glacier was inconsistent with the parameters (e.g. the role of basal slip) used to verify other possible LLS glaciers in the Brecon Beacons. Furthermore, solar radiation modelling showed that this site received large amounts of solar radiation and would have required at least 60% of accumulating snow to have come from snowblow. Instead Carr et al. (2007b) asserted that the features originate from a paraglacial rockslope failure which produced a wedge of sediment that extends a further 400 m down-valley where it becomes progressively thinner. However, this explanation does not account for the abundance of striated clasts within the ridge, nor the absence of a landslide scar in the headwall, and thus confident interpretation of these features requires further research.

In the eastern Brecon Beacons, depositional ridges have been recognised at about a dozen sites at the heads of the glacial valleys and indentations into the escarpment, which reaches its highest point at Pen y Fan at 886 m OD (Figure 3.28a). The clearest features are found at the head of Cwm Llwch (Figure 3.28c) where a large moraine, up to 18 m high and comprising multiple ridges is suggestive of either a large availability of debris or an especially active glacier (Shakesby, 2002). Glaciological reconstruction suggested that the Cwm Llwch glacier could have formed during the LLS, but would have been fairly slow moving, leading Carr (2001) to infer that a large component of debris fell directly onto the glacier from the valley walls. Other clear features are observed at Cwm Crew (Figure 3.28d) in the form of a sharp-crested ridge, which reaches up to 15 m and contains angular and sub-angular, occasionally striated, rock fragments (Shakesby, 2002). This glacier is exceptional in its unfavourable southerly aspect which would have afforded no protection from insolation. Consequently, the contribution of windblown snow from the high ground to the southwest must have been substantial to offset snow lost to insolation.

This style of a small glacier accumulating in (usually) a northeast facing cirque and forming an arcuate end moraine which in some instances encloses inner fragmented ridges is typical for the Brecon Beacons. At some sites, single moraine ridges are present, whereas other sites have one or two inner ridges, as observed at Cwm Oergwm, where examination of exposures in the ridges led Coleman and Carr (2008) to believe that they represented an outer push moraine and two sub-parallel retreat moraines. Using

glaciological modelling, Carr (2001) and Coleman and Carr (2008) asserted that glaciers reconstructed from these ridges at Cwm Llwch, Pen Milan, Cwm Gwaun Taf and Cwm Crew had plausible ice dynamics (particularly basal shear stresses, ice deformation velocities and the component of glacier flow through basal sliding) and thus could have formed during the LLS. However, application of this technique at Cwm Cwareli suggested that only the inner two of the three arcuate moraines found at this site could have been formed by LLS glaciers, and it was argued that the outermost must have formed under colder conditions than those reconstructed for the LLS, probably during deglaciation of the last British Ice Sheet. Similar moraines were identified at Cwm Cul (Shakesby, 2002), but the reconstructed glacier here was implausibly thin and required almost all movement to be driven by basal motion, leading Carr (2001) to reject a LLS age for the glacier. A further four glaciers along the northern flank of Mynydd Llangattwg (approximately 14 km southeast of Cwm Pwflla) were also deemed to predate the LLS, based on the calculation that an implausibly high proportion of mass-transfer would have been required from basal motion were these glaciers to have existed during the LLS and therefore an older age was suggested (Coleman, 2007; Coleman and Carr, 2007). Sedimentological evidence and radiocarbon dating from Waen Du bog within this area supported this interpretation (Trotman, 1964; Coleman and Parker, 2007). To date, the only absolute dating of these small cirque glaciers was conducted at Cwm Pwllfa (Figure 3.28a, chronological site 3) where a radiocarbon date of 8290±35 [14]C a BP was calculated for basal organics in a core from within the moraine complex (Preston et al., 2007). Although this age is younger than would be expected, the absence of Lateglacial sediments within the core supports a LLS age.

## 3.5 Discussion

### 3.5.1 Assessment of evidence

A key issue in terms of the evidence of LLS glaciation in Britain is the uneven coverage of previously published mapping (Bickerdike et al., 2016). The most conspicuous area is the largely unmapped southwestern Highlands where mapping is limited to four localised sites along the shorelines of Loch Fyne. It is not possible to determine whether this lack of mapping is the result of a general lack of glacial landforms in this region or whether it has simply been overlooked by researchers. It has been suggested that the general reduction in drift along the shores of several of the lochs to the north, including Lochs Nevis, Morar, Ailort, Shiel, Sunart and Linnhe, resulted from very active, potentially streaming, glaciers stripping deposits from the valley sides followed by a rapid retreat that left little depositional evidence (Bennett, 1991). It is conceivable that the same processes might have been responsible for the apparent lack of evidence in the southwest Highlands along

the banks of Lochs Fyne and Long. Therefore, systematic mapping is required to determine whether depositional evidence is indeed absent from these lochs or whether indicators of glacial abrasion, such as the density of striae observed along Loch Hourn and Loch Nevis, are present.

For the eastern outlet glaciers, the evidence compiled from the literature is very patchy aside from detailed geomorphological mapping of the termini of the Lomond, Menteith and Callander lobes (Thompson, 1972; Rose, 1980) and two specific areas at Crianlarich and Glen Lyon (Golledge and Hubbard, 2005; Wilson, 2005). Preliminary inspection of aerial imagery indicated that moraines do appear to be widespread within this area, although future mapping of these features could be inhibited in some areas by a combination of extensive afforestation and blanket peat, both of which obscure glacial landforms. Given the current lack of detailed mapping for much of this region, establishing the extent and dynamics of LLS glaciers in this region is highly speculative and will remain so until this deficiency is addressed.

The identification of offshore moraines using bathymetric surveying has indicated that in some areas LLS glaciers appear to have extended beyond the extent inferred from the onshore evidence (Dix and Duck, 2000; McIntyre et al., 2011). In light of this, systematic surveying of other sites where LLS glaciers terminated beyond the present-day coastline, such as Loch Sligachan on the Isle of Skye, and Lochs Duich, Nevis, Ailort and Sunart on the west coast, is necessary to determine whether the onshore evidence of these glaciers truly reflects their maximum extent. In cases where moraines are observed beyond the maximum onshore limits, features could be dated to confirm whether they represent the LLS and whether these limits were held only briefly or for a longer period during the stadial.

The quality of the source mapping is highly variable. Whilst some researchers have mapped individual moraine mounds and hummocks as defined by the break of slope around the landform (Benn et al., 1992; Ballantyne, 2002a; Lukas and Lukas, 2006a, b), others have mapped moraine ridge crests or, simply areas of 'hummocky moraine'. In the case of the latter it is difficult to assess how well the mapping represents the glacial geomorphology. Since the form of geomorphological features is often used to infer the processes which created them (Kleman et al., 2007), it is important that landforms are accurately represented so that the dynamics of past ice masses are not misinterpreted.

The usage of the term 'hummocky moraine' is particularly problematic as the interpretation of this landform, which is found extensively within the margins of LLS glaciers in Scotland, has changed significantly over time (Benn, 1990). Sissons (1961) argued that hummocky moraine consisted of a combination of till and glaciofluvial sediments and attributed its apparently chaotic arrangement to widespread stagnation of LLS ice. However, others suggested that this landform type was instead associated with active glacial retreat (Thompson, 1972; Hodgson, 1982; Eyles, 1983). Subsequently, it was discovered that large areas of hummocky moraine comprised transverse moraine ridges, chaotic mounds and flutings (Benn, 1992; Benn et al., 1992) and, in many locations, the hummocks form sequences of nested arcuate chains of mounds, believed to be recessional moraines. Where detailed mapping records the position of individual moraine mounds or ridge crests it has been possible for researchers to identify the positions of palaeo-ice fronts and infer the dynamics of these actively retreating glaciers (Bennett and Boulton, 1993a). Where authors have simply mapped the extent of areas of hummocky moraine it is not possible to determine the distribution and morphology of the moraines themselves, nor the dynamics of the glaciers which formed them. This is currently the case for much of the central Highlands, including the southeast Grampians (Sissons, 1972; Sissons and Grant, 1972) and Gaick Plateau (Sissons, 1974), and large areas of Rannoch Moor (Thorp, 1984), which were last mapped when hummocky moraine was associated with widespread ice stagnation. More detailed remapping of the distribution of individual depositional mounds within these areas of hummocky moraine would determine whether they comprise belts of transverse recessional moraines, formed under active retreat, or a more chaotic assemblage, formed during ice stagnation. This would permit a much greater understanding of LLS glacier retreat dynamics across Britain as a whole.

A further challenge in interpreting the landform record and accurately determining the extent of LLS glaciation is the ambiguous origins of some of the geomorphological evidence. Whilst sequences of recessional moraine ridges can be readily identified as glacial in origin, the landform record in marginal locations, where LLS glaciation occurred only in topographically favourable niches, is often limited to single arcuate moraine ridges. These features can appear morphologically similar to protalus ramparts, such as those in the Brecon Beacons (Ellis-Gryffydd, 1977; Shakesby, 2007). In other locations, there has been uncertainty when attempting to determine whether debris accumulations represent glacial features, rock glacier deposits (Lowe, 1993; Shakesby, 2002; Mills and Lukas, 2009) or the products of mass movement (Dawson et al., 1987; Shakesby and Matthews, 1996; Ballantyne, 2002a; Carr and Coleman, 2009). Since these ambiguous features are usually associated with small cirque glaciers, misinterpreting them has a minor effect on

reconstructions of the total extent of LLS glaciation in Britain but is significant at a local scale, because the distribution of ice in these marginal locations is related to the palaeoclimatic conditions during the LLS. Although efforts were made during this review to cross-check landforms against the NEXTMap hillshade data, the similarity in morphology between moraines and other types of debris accumulation, coupled with difficulties regarding the small scale of these features and the 5 m resolution of the NEXTMap data, has inhibited distinguishing the origin of such features. It is hoped that highlighting the ambiguous genesis of these landforms in this review will prompt future targetted research to determine their origins.

Given that geomorphological mapping forms the first step in reconstructing both the dimensions and dynamics of former glaciers, it is crucial that these data are both detailed and accurate. Analysis of retreat dynamics (c.f. Lukas and Benn, 2006) is not possible from low resolution data, such as where areas of moraines, rather than the individual landforms, are recorded. As such, it is only possible to compare the retreat dynamics of glaciers from areas of detailed mapping, meaning that substantial areas (such as much of the central section of the West Highland Glacier Complex) are neglected from in depth study. Although undoubtedly time-consuming, systematic re-examination of areas where mapping is of low resolution or is out-dated will be necessary before a holistic view of retreat dynamics of LLS glaciers throughout Britain can be obtained.

### 3.5.2 Modelling LLS glaciation

Numerical modelling has proved an essential tool in identifying the relationship between glaciers and climate during the LLS in Britain. Hubbard (1999) used a three-dimensional, time-dependant ice flow/mass-balance model, driven by a locally calibrated GRIP temperature time series, to create an ice mass that matched the empirically derived ice limits within 550 years of the onset of cooling. Steep northwards and eastwards reductions in precipitation alongside an overall 20% reduction in overall precipitation from 350 years into the model run were found to be key components to allow the ice mass to reach climatic equilibrium at the reconstructed limits. Golledge et al. (2008) expanded on this work using a modified, higher resolution version of the same model. The optimum Younger Dryas scenario, which closely accorded with the many of the empirical limits, used a 10°C depression of maximum mean annual temperatures to scale the GRIP record. This was coupled with south to north and west to east precipitation reductions of 60% and 80% respectively, with a stepped reduction in overall precipitation from 12.5-12 ka and a low component of basal sliding. The closeness of the modelled output with the empirical limits at a nation-wide scale and the small parameter space within which this match occurs suggest that these conditions could closely represent those during the LLS.

However, a closer inspection shows discrepancies between the modelled and empirical limits in several locations which probably arise where local conditions facilitated or inhibited the formation of LLS glaciers. For example, the modelled Lomond, Menteith and Callander lobes fall far short of the empirical limits; Golledge et al. (2008) suggested that this mismatch could be the result of local factors, such as where the basal drag was reduced by a deformable sediment bed or water body, allowing the glacier to advance further, which the model does not account for. Incorporating bed characteristics into future numerical models could both contribute to more accurate representation of glacier extents but also allow inferences about the controls on ice flow dynamics to be made. Similarly, as highlighted by Golledge et al. (2008), their modelling assumes LLS glaciers developed from ice-free conditions. However, recent mapping of offshore moraines around the northwest coast of Scotland, coupled with surface exposure dates, suggests that ice caps existed in the northwest Highlands throughout the Lateglacial Interstadial and into the LLS (Bradwell et al., 2008). If the LLS glaciers grew from pre-existing ice at the beginning of the stadial, this could explain why Golledge et al.'s (2008) initially ice-free model under-predicts the extent of LLS glaciation in the northwest Highlands and other sites. Further work to determine whether ice was inherited in other locations is needed and, if this is found to be the case, such initial conditions must be incorporated into numerical models if an accurate representation of LLS glaciation is to be obtained.

Likewise, although the overall configuration of the 'best fit' simulation of Golledge et al.'s (2008) model for the Beinn Dearg ice mass (Figure 3.11) was similar to the empirical limits reconstructed by Finlayson et al. (2011), at a local scale there was a degree of mismatch with the model overestimating the extent of glaciers in the west and underestimating their extent in the east. Finlayson et al. (2011) suggested that this disparity could relate to the model not accounting for redistribution of windblown snow which would explain the eastward decline in ELAs across the ice cap in the empirical reconstruction. In numerous regions, snow redistribution onto glaciers lying in the lee of plateau surfaces by predominantly southwesterly winds has been identified as a control on glacier accumulation (Sissons, 1980b; Cornish, 1981; Mitchell, 1996; Carr, 2001; Harrison et al., 2006). In light of this, future modelling, particularly at a local scale, should incorporate an approach that reflects the influence of snow-contributing area.

Numerical modelling has been used to create dynamic reconstructions of glacier extent throughout the LLS. A time-dependent, two-dimensional ice flow model was used by Brown et al. (2013) to create a dynamic reconstruction of the icefield in the central Lake District. The model, which was forced by an ELA record scaled from the GRIP record,

predicted that the Younger Dryas glaciation was characterised by three advance phases in the Lake District: "an initial maximum extent, a middle minor advance or stillstand, and a pronounced but less extensive final advance" (Brown et al., 2013: p1022). The terminal positions of glaciers during the second and third phases closely matched the distribution of prominent latero-frontal moraines mapped by Brown et al. (2013). The ability of such models to create dynamic glacial reconstructions which capture the changes in glacier extent throughout the stadial, as driven by climate, provides perhaps a more realistic insight into glacier behaviour than the static maximum reconstructions commonly produced from empirical data. Application of this style of numerical modelling to other LLS ice masses could be used to aid comparison of the possible dynamics of glaciers across Britain. Furthermore, such modelling could be used to assess whether the apparent pattern of moraine formation during the less extensive second and third advance phases is replicated at other locations.

### 3.5.3 Style of glaciation

From a review of the literature, it is clear how advances in glaciological theory and mapping technology have influenced the prevailing understanding of LLS glaciation in Britain. Charlesworth's (1955) account of the extent and retreat patterns of LLS ice in Scotland represented a significant study within this body of literature but, given the lack of field evidence included in that study, it is unsurprising that much of this work has subsequently been revised. The first systematic mapping of many locations, much of which was conducted by Sissons (1974, 1977a) from aerial photographs, gave rise to the understanding that the LLS was characterised in most areas by alpine style valley and cirque glaciers. Following the proposal that LLS glaciers in the central Lake District were likely to have been nourished by a plateau icefield (McDougall, 1998; Rea et al., 1998) many of the upland areas believed to have only been occupied by valley and cirque glaciers have been reinterpreted as having supported icefields or ice caps (Ballantyne, 1989; Benn et al., 1992; Rea et al., 1998; McDougall, 2001, 2013; Lukas and Bradwell, 2010; Finlayson et al., 2011; Brown et al., 2011, 2013). Such reinterpretation has implications for glacial reconstructions and for ELAs calculated from them; incorrectly assuming cirque occupation only will produce an artificially lower ELA than if the mass of ice on the plateau was accounted for (McDougall, 2001). As several studies have assessed regional trends in ELAs (e.g. Ballantyne, 2007a) or used them to infer palaeoclimatic conditions, such as temperature and precipitation (e.g. Ballantyne, 2002a, 2007a, b; Benn and Ballantyne, 2005), accurately calculating ELAs and, therefore, correctly interpreting the style of glaciation of a particular area is essential.

The increased availability of high resolution remotely sensed data has facilitated mapping in previously under-researched locations, such as the Monadhliath (Boston, 2012a). Cumulatively, the shift towards icefield styles of glaciation and mapping of new areas indicates that a much larger area of Britain was glaciated during the LLS than previously believed. These limits accord with those derived from numerical modelling (Golledge et al., 2008), supporting the long-held assumption that glaciation in the central and eastern Grampians was restricted by the reduced precipitation levels in these regions (Sissons and Sutherland, 1976; Golledge, 2010).

Whilst published modelling has not accommodated the contribution of windblown snow onto the glacier surface, this factor was critically important in ice accumulation and accounts for the presence of cirque glaciers in topographically favourable locations. This applied especially to marginal areas where, although the conditions were generally unsuitable for glaciation, glaciers formed downwind of large potential snow-contributing areas. Although most of these small glaciers were north to east facing, as in the Pennines (Mitchell, 1996), in some cases the effect of snowblow was probably sufficient to offset the increased insolation received by more southerly facing cirques, for example the cirque just southeast of Meikle Millyea in the Southern Uplands (Cornish, 1981) or Cwm Crew in the Brecon Beacons (Carr, 2001).

### 3.5.4 Extent of LLS glaciation

A key contribution of this review is a new assessment of the maximum extent of glaciation during the LLS made at a regional scale (Figure 3.29). This builds upon earlier studies (e.g. Sissons, 1967b; Golledge, 2010) concentrated on Scotland by also incorporating glaciers in England and Wales. Empirically based glacial reconstructions derived from geomorphological mapping were compiled from the published literature and were georeferenced into the database. In all instances, the maximum possible extent of glaciation was used; for example, where areas of cold-based ice had been proposed, these were incorporated into the reconstruction (as in Lukas and Bradwell, 2010). As with the geomorphological mapping, preference for inclusion was given to the most recent reconstructions, but it was necessary to amend these in specific locations where they did not match the most recent geomorphological mapping. For example, the termini of the glaciers that occupied Loch Ainort on the Isle of Skye (Figure 3.3) and Loch Hourn, on the west coast of Scotland (Figure 3.2), were extended to include moraines mapped during bathymetric scanning of these sites (Dix and Duck, 2000; McIntyre et al., 2011).
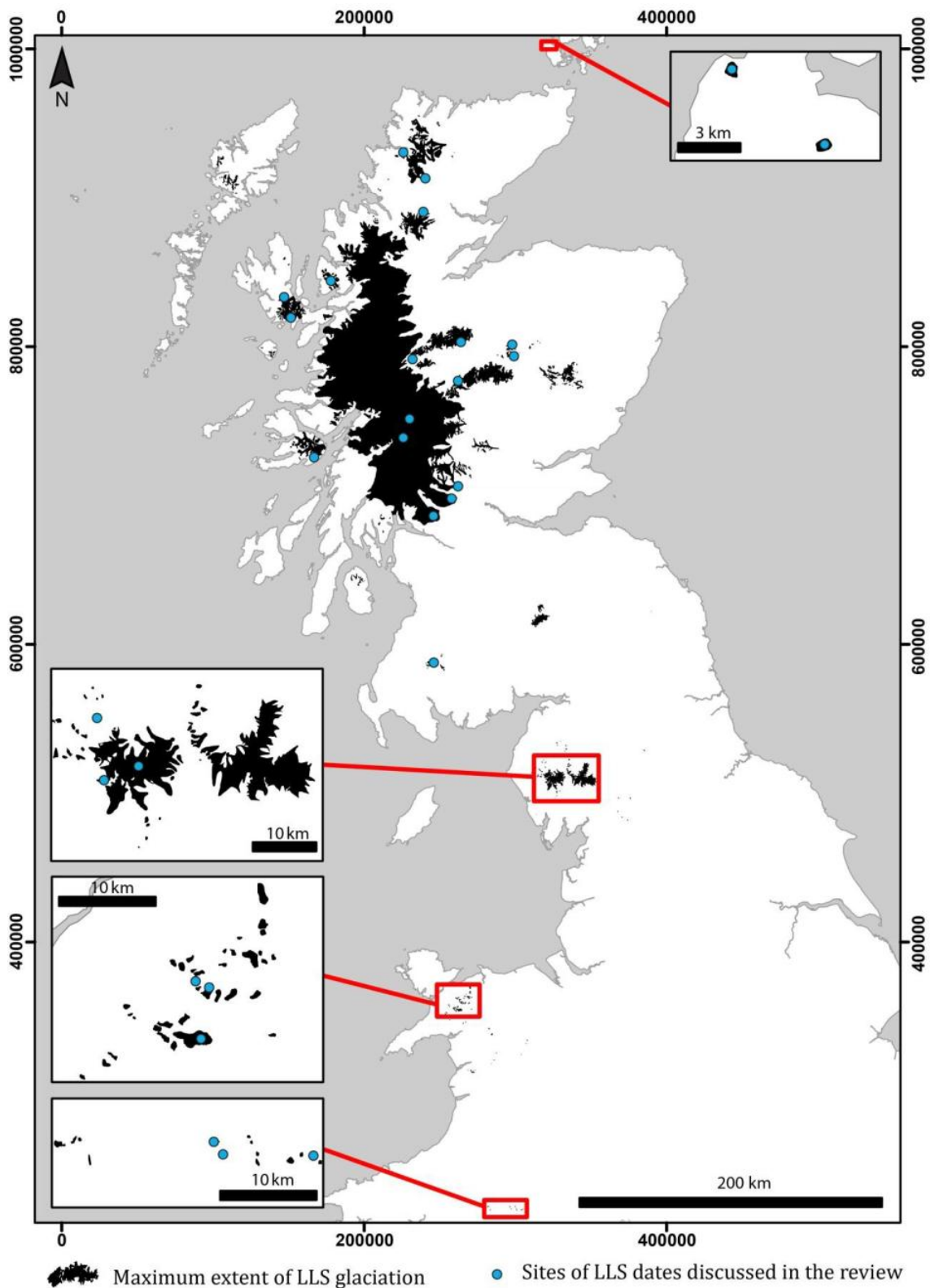
Figure 3.29 Approximate reconstruction of the maximum extent of LLS glaciation in Britain. The West Highland Glacier Complex is shown flanked by satellite icefields, while in marginal areas, glaciation is limited to topographically favourable valleys and cirques. Britain coastline reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence).

In areas where geomorphological evidence is absent, the resolution of this reconstruction is, by necessity of its scale, reasonably coarse and areas such as the northern Tweedsmuir

Hills (Figures 3.1 and 3.29) and the southwest sector of the main icefield are speculative and will require refinement once the necessary geomorphological mapping of these sites has been conducted. Likewise, although the vertical extent of the satellite icefields is generally well-constrained (Ballantyne, 2002a; Lukas and Bradwell, 2010; Finlayson et al., 2011) and thus has been included, this is not the case for the West Highland Glacier Complex. Geomorphological features reflecting the thickness of these glaciers (such as trimlines) are either absent from mapping (as in the northern sector) or are no longer thought to represent the maximum elevation of the ice surface and thus identifying the vertical extent of the main icefield has not been attempted for this study. Consequently, using probable estimates of the ice thickness to identify the most likely configuration of ice and nunataks in this area is a natural next step in improving this reconstruction. The reconstruction shapefile, available as supplementary material with this paper, provides a framework for future research. The shapefile allows rapid calculation of statistics associated with LLS glaciation, such as the total area of ice, at a scale not previously attempted. This facilitates comparison of the extent and characteristics of ice at different locations. Usage of the shapefile in conjunction with digital terrain models (DTMs) can be used to assess the relationship between underlying topography and glacier development. For example, slope models created from DTM data can be used to measure the potential snow-contributing area onto a neighbouring glacier, allowing the relationship between this factor and glacier extent to be quantified for different locations. This provides an insight into the potential controls on glacier development. The shapefile also provides a target maximum extent for numerical modelling experiments. Whilst modelling was used to identify a narrow parameter space of palaeoclimatic and basal conditions that matched well with the extent of glaciation in Scotland (Golledge et al., 2008), this shapefile provides limits that could be used to determine whether the same parameters fit as accurately with the LLS glaciers reconstructed in England and Wales, where glaciation was more restricted.

It is crucial to note that the certainty with which the glacier limits presented on Figure 3.29 have been identified varies greatly across Britain. As this review of the published literature has highlighted, although the extent of LLS glaciation can be confidently identified in some regions, in many areas this is not the case. Figure 3.30 represents a preliminary attempt to qualitatively identify which limits are well-constrained, both by geomorphology and chronology. Chronological sites discussed in the review have been included on the map to highlight the relative paucity of dates on, or within, the LLS limits and to illustrate how many of the LLS glacier limits have not been robustly dated. Each site has been labelled as follows: "LLS" where chronological evidence supports an "LLS" age;

"LD" where dating has indicated formation during the retreat of the last (Late Devensian) British-Irish Ice Sheet; or "?" where the date can indicate either a LLS or earlier age, depending on the choice of production rate used. Dates associated with the build-up and retreat of the British-Irish Ice Sheet, as compiled by Hughes et al. (2011), are represented by black dots.

From Figure 3.30, it is clear that sites where the extent of the LLS glaciers is well-constrained by both geomorphology and absolute dating are comparatively rare across Britain. Examples of these well-constrained limits include the termini of the Lomond and Menteith glaciers (indicated by solid black lines). In other areas, the extent of ice has been reconstructed from high resolution geomorphological mapping but chronological control on the limits may either be indirect, such as for sections of the Mull icefield, or has been contradicted by subsequent work (e.g. Gleann Chaorainn in the Monadhliath) (purple dashed lines). Indeed, high resolution mapping has facilitated reconstruction of many of the satellite icefields, yet chronological control has either been interpolated to the whole icefield from a single outlet glacier (such as for the Ben Hee, Monadhliath and Beinn Dearg icefields), or may be completely absent (as is the case on Lewis and Harris in the Outer Hebrides) (blue dashed lines).

The extents of the glaciers which comprised the West Highland Glacier Complex are moderately well-constrained in both the northern and central sectors (indicated by the green dashed lines). However, since it has been suggested that glaciers which flowed beyond the present-day coastline may have extended further than suggested by the onshore evidence (Dix and Duck, 2000; McIntyre et al., 2011), the limits of the glaciers along the western coast to the north of the Great Glen are here viewed as speculative (indicated by the orange dashed lines). Likewise, the limits of glaciation on the Gaick Plateau and in the Cairngorms and southeastern Grampians have been the subject of considerable uncertainty in the literature and require further study. This is also the case for the plateau icefield in the eastern Lake District, where the geomorphology indicating the limits of glaciation is at times ambiguous and where no chronological constraint is available. The least constrained limits are found in the southern sector of the West Highland Glacier Complex, which, with the exception of the termini of the Lomond, Menteith and Teith glaciers, is almost entirely unmapped and undated (represented by the red dashed line). By detailing the degree of confidence in the extent and chronology of the LLS glaciers, Figure 3.30 highlights regions where further work is required. It also provides a constraint for those modelling the extent of LLS ice by indicating which limits models need to fit closely and which are more speculative.
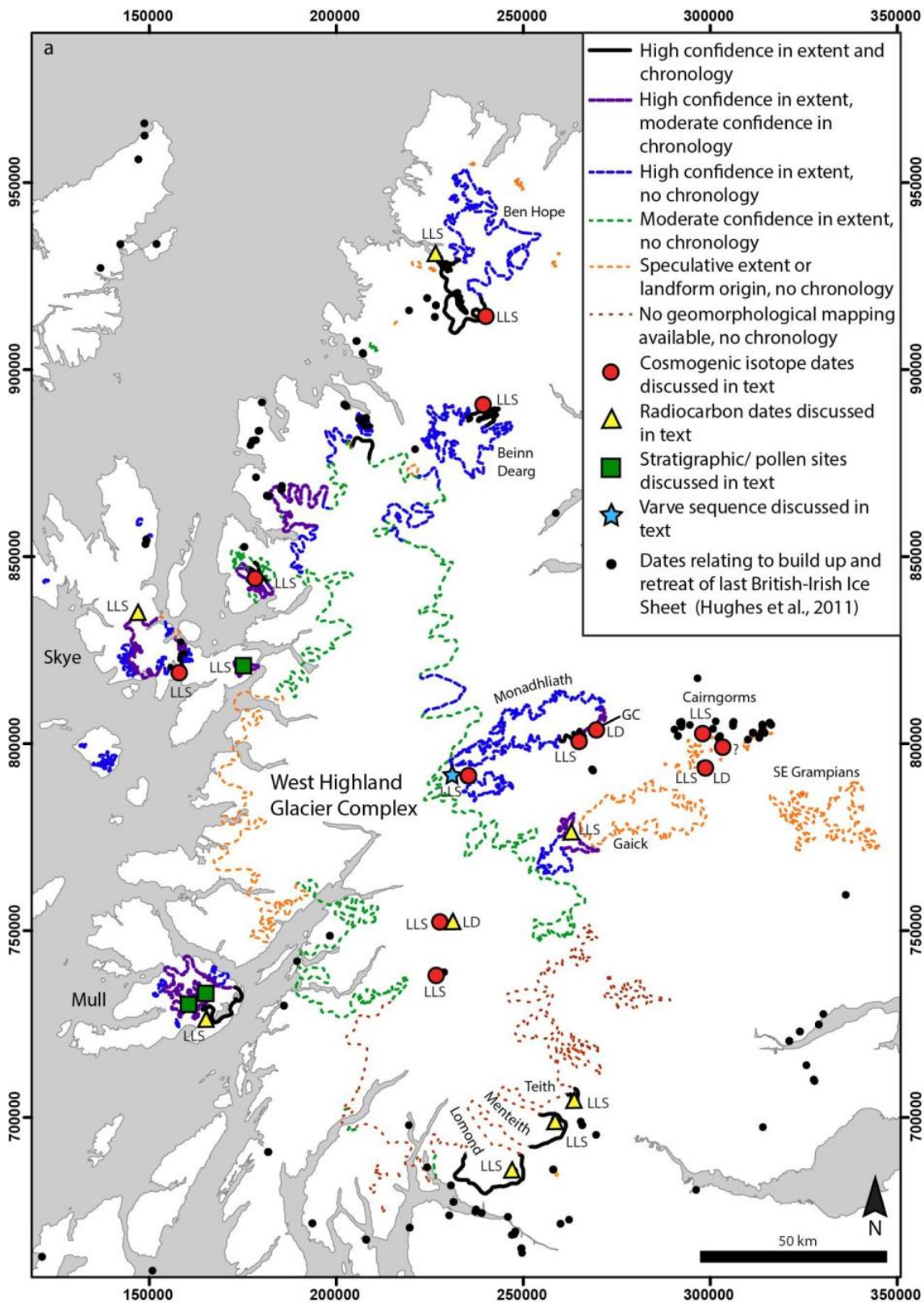
Figure 3.30 Qualitative assessment of the level of confidence in the extent and chronological constraint of LLS glaciers in a) Scotland, b) the Lake District, c) the Isle of Arran, d) Snowdonia, e) Tweesdmuir, and f) the Outer Hebrides. Across Britain are few sites where both the extent and age of the ice limit is well-constrained. Reconstructions of numerous ice masses, including several in the eastern Scottish Highlands and eastern Lake District, remain speculative and a significant section of the West Highland Glacier Complex remains unmapped. GC: Gleann Chaorainn. Britain coastline reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence).
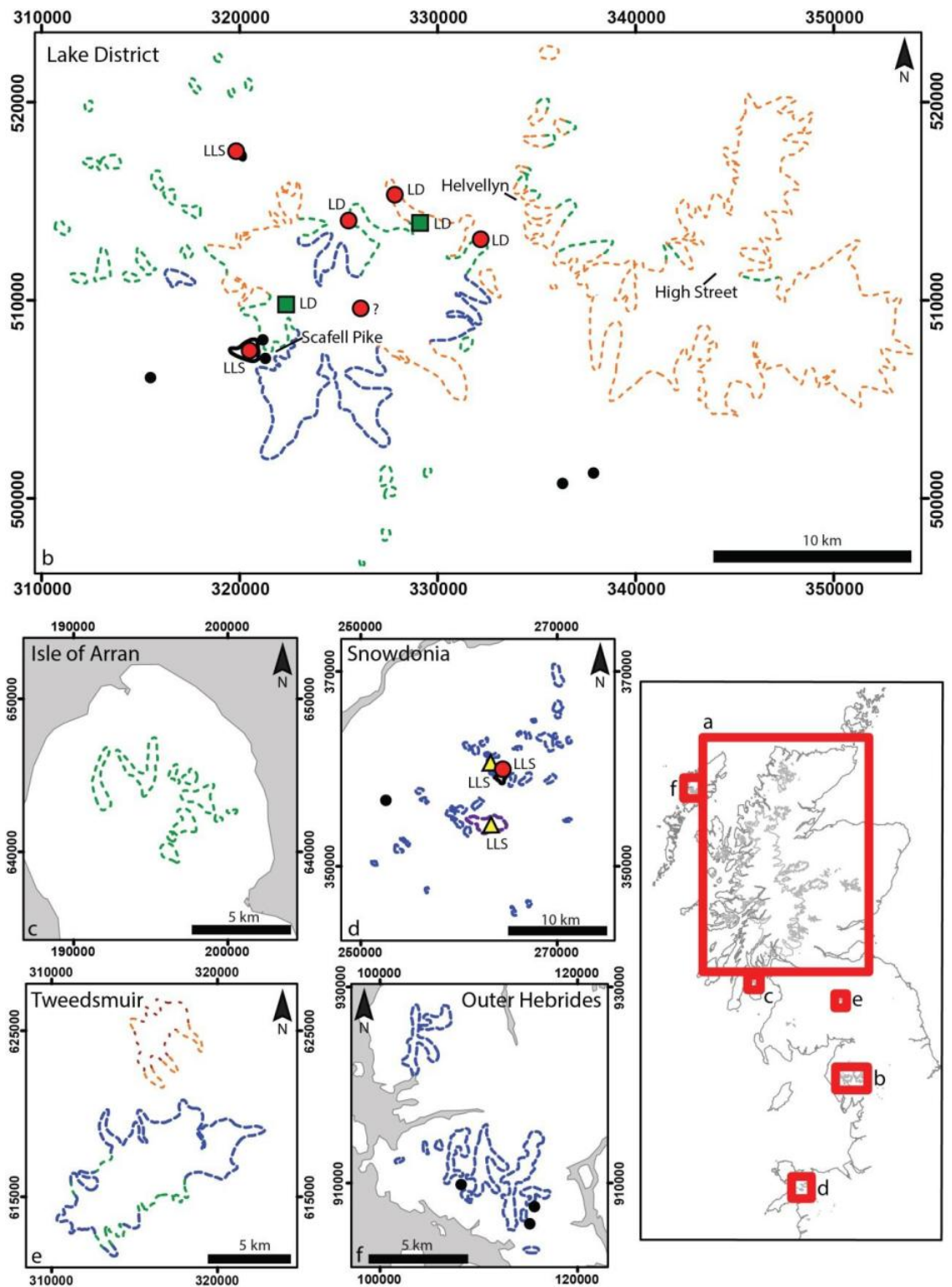
### 3.5.5 Timing and dating LLS glaciation

A great deal of the uncertainty surrounding the extent and nature of LLS glaciation in Britain is the result of the currently limited number of absolute dates for glacial features associated with this period, as illustrated by Figure 3.30. Consequently, the age of many landforms has been extrapolated from what dates exist or inferred from other evidence.

During the 1970s a LLS age was commonly inferred for glacier limits based on the presence of hummocky moraine within the limits, which was believed to be diagnostic of this period (Sissons, 1974). However, the presence of Late Devensian sediments from within the area of hummocky moraine at Loch Builg (Clapperton et al., 1975) and the existence of similar landforms beyond the apparent LLS limit at Strollamus on the Isle of Skye (Benn, 1990) (Figure 3.3), indicated that hummocky moraine was not exclusively the product of LLS glaciation, and indeed there is no sound glaciological reason to expect it to be. Consequently, even where hummocky moraine is present within glacial limits, more recent studies cite multiple lines of evidence to support a LLS age. Many authors, for example, have observed a mutually exclusive relationship between the landforms created by locally nourished glaciers and relict periglacial features, including frost-weathered bedrock, solifluction lobes and mature talus slopes (Sissons, 1974; Ballantyne and Wain-Hobson, 1980; Ballantyne, 1989, 2007a; Benn and Ballantyne, 2005; Finlayson, 2006). Given the absence of periglacial features within the reconstructed glacier limits, but their presence immediately outside of them, it is reasonable to assume that the areas inside the limits were protected by glacier ice during the last period of severe cold conditions and thus a LLS age is inferred for these limits. Similarly, in some locations, Lateglacial shorelines were restricted to outside these limits whereas only lower, Flandrian shorelines, were present within them, suggesting that the older shorelines formed during ice sheet deglaciation were destroyed during a later phase of glacial readvance, i.e. the LLS (Ballantyne, 1989, 2002a). Likewise, in several locations it has been observed that whilst both extensive river terrace sequences and large glaciofluvial features, such as kames or eskers, may be present in the lower valleys, these are absent within the inferred glacier limits (Sissons, 1974; Lukas, 2006; Boston et al., 2015). These lines of evidence were formalised by Lukas (2006) into a series of morphostratigraphic principles to identify landform assemblages of specific age. This approach stresses the importance of using multiple lines of evidence when inferring a LLS age for landforms. These criteria have been applied at a variety of locations including the northwest highlands (Lukas and Lukas, 2006a, b), Monadhliath and Creag Meagaidh mountains (Finlayson, 2006; Boston, 2012a) and Tweedsmuir hills (Pearce et al., 2014a) and provide a more robust method of determining the extent of LLS glaciers than over-relying on a single landform type such as hummocky moraine.

A LLS age for glacial features has commonly been inferred from stratigraphic evidence (Seddon, 1962; Pennington, 1977; Benn et al., 1992). Sediment cores from basins beyond the limits of LLS glaciation show a classic Lateglacial tripartite sediment sequence, whereas inside the limits only Flandrian organic sediments are present. The age of these

sediments has been supported by radiocarbon dating and pollen analysis (Pennington, 1977; Walker et al., 1988; Benn et al., 1992). However, this technique can only be applied to locations where organic material is preserved and thus has been largely unsuitable for areas such as the northwest Highlands where bedrock remains close to the surface (Lukas and Bradwell, 2010). In such instances, cosmogenic isotope dating has provided a possible alternative technique to determine the age of glacial landforms and has been used at a variety of locations (Bradwell, 2006; Ballantyne et al., 2007; Finlayson and Bradwell, 2007; Golledge et al., 2008; Small et al., 2012; Wilson et al., 2013; Small and Fabel, 2016). However, this approach is not without problems, particularly those associated with the uncertainty regarding the production rate of isotopes used in the calculation. Ballantyne (2012) used a locally derived production rate to recalibrate 33 previously published dates and found that this led to ages 130-980 years older than previously calculated. Likewise, Standell (2014) could not determine whether Coire Etchachan (Figure 3.21) was last glaciated before or during the LLS as both results were possible depending on the production rate used. These uncertainties are especially pronounced given the relatively brief 1200 year total duration of the LLS.

Determining the age of glacial landforms is of particular importance when reconstructing the extent of the LLS ice cover and numerous cases exist where changing age attributes have radically altered the palaeoglaciological reconstruction. For example, reinterpretation of the moraines and terraces at Achnasheen (Figure 3.12) as having formed during the LLS rather than during an earlier phase of glaciation led to the reconstruction of a much more extensive icefield in the northwest Highlands (Bennett and Boulton, 1993a); this has been compounded by the fact that subsequent numerical modelling suggests a more restricted ice extent was appropriate for this area (Golledge et al., 2008). Likewise, in the Lake District, the extent of the northern sector of the icefield has been inferred from a few end moraines at Rosthwaite and Watendlath which, if they are of LLS age, imply a substantially greater ice extent than if they predate the stadial (McDougall, 2001). Where conflicting interpretations of geomorphological evidence exist, systematic dating of particular landforms can be used to determine which reconstruction is most realistic. For example, cosmogenic isotope dating was used by Golledge et al. (2007) and McCormack et al. (2011) to indicate that LLS glaciers were more vertically extensive than previously proposed for Rannoch Moor and the Applecross Peninsula respectively.

These various dating techniques have been applied in a piecemeal fashion to the LLS glacial landforms. Studies now commonly use multiple lines of geomorphological evidence

when inferring the age of undated landforms, rather than relying on hummocky moraine or morphological characteristics such as moraine 'freshness' (e.g. Gray and Brooks, 1972; Sissons, 1974, 1979c, 1980a) being diagnostic of a LLS age. Morphostratigraphy has proved a useful technique in assessing the age of glacial limits as marked by particular landform assemblages. This is especially true when such limits are locally subject to absolute dating, allowing this age to be extrapolated to adjacent valleys displaying the same landform assemblages. However, there remain several locations where the age of particular landforms remains poorly constrained, particularly where absolute dating is lacking.

As a result of the previously unsystematic and spatially inconsistent approach to dating LLS glacier features, assignation of landforms to the stadial is problematic and uncertainties relating to age are far more common than those regarding landform genesis. As illustrated by Figure 3.30, the limit of LLS glaciation has been robustly dated in very few locations. Even where dating has been attempted, the reliability of chronological evidence may still remain controversial, as is the case in the Lake District (Wilson et al., 2013) and Gleann Chaorainn in the Monadhliath (Gheorghiu and Fabel, 2013; Boston et al., 2015). In both of these examples, dating of features on, or immediately inside, the empirical LLS limit returned pre-LLS ages that were subsequently challenged on the basis of probable nuclide inheritance (e.g Boston et al., 2015). Likewise, it is not possible to reconcile the conflicting radiocarbon (Bromley et al., 2014) and cosmogenic (Small and Fabel, 2016) dates from Rannoch Moor. Even in areas where chronological control is thought to be reliable (e.g. Bradwell, 2006), interpolating a single date to an entire icefield should be done with caution, especially in areas where the glacier limits are not clearly traceable from one valley to the next. Thus, in Figure 3.30, only sections of the limit in the immediate vicinity of dates are deemed to be chronologically constrained.

Whilst the stratigraphic and chronological evidence from Croftamie and Callander (as discussed in Section 3.2.3.3) makes it very difficult to refute that at least some glaciers underwent a significant readvance during the LLS, such sites are exceptional. The majority of LLS chronological evidence, including both cosmogenic isotope dates and basal radiocarbon dates on organic sediments, relates to the onset of ice-free conditions. As such, it is not possible to determine that these dates relate to the retreat of glaciers from a distinct readvance period (i.e. the LLS) rather than deglaciation of the last British-Irish Ice Sheet. Clear contrasts between the landform signatures in and outside proposed LLS limits (for example sharper-crested moraines inside and large, more subdued moraines outside), have been cited as evidence that the former features were created during a distinct

readavance period (e.g. Lukas, 2006; Boston, 2012a; Pearce et al., 2014a), rather than as part of a continuous process of ice sheet retreat. However, as discussed above, the validity of using morphological appearance as a proxy for age has been questioned (Wilson, 2002). Cosmogenic isotope dating and bathymetric surveying of moraines in the northwest Scottish Highlands has indicated that, at least in this region, ice sheet deglaciation was punctuated by major readvances (such as the Wester Ross Readvance) but also by much smaller oscillations of the ice margins (Bradwell et al., 2008). The existence of these other oscillations, coupled with the fact that substantial ice masses may have survived throughout the Lateglacial Interstadial (Bradwell et al., 2008), complicates assignation of landforms to the LLS, as geomorphological evidence relating to the LLS glaciers is overprinted onto those formed during earlier readvances. In the absence of robust absolute dating, it is difficult to assign these landforms to the LLS with a high degree of confidence (as illustrated by Figure 3.30) and, furthermore, to definitively prove that the LLS represented a period of significant glacier advance, rather than a smaller oscillation during ice sheet retreat.

Given the limitations of the current body of chronological evidence associated with the LLS glaciers, further targetted research is required before the total extent of glaciation during the stadial can be inferred reliably. There are a number of locations where collection of robust absolute dates could help to distinguish between very different hypotheses of the extent of LLS glaciers. For example, cosmogenic isotope dating of the outer moraines in valleys flanking the Gaick Plateau could be used to indicate whether these limits were reached during the LLS. If these limits do indeed represent the maximum extent of LLS glaciation, this would support the configuration of ice proposed by Sissons (1974, 1980b), that of valley glaciers nourished by ice accumulating on the plateau. Conversely, if these limits were found to predate the LLS, this would support the model of restricted glaciation of a few cirques favoured by Merritt (2004b). Likewise, the presence of LLS ice in the neighbouring Drumochter Hills has been inferred from the absence of mature periglacial features within the limits proposed by Benn and Ballantyne (2005) and dating of basal sediments to 10.5 ka cal BP (Walker, 1975). Although it does seem likely that this location was glaciated during the LLS, absolute dating would help to determine whether the geomorphologically well-constrained limits reconstructed by Benn and Ballantyne (2005) do represent an icefield of LLS age, or whether these glacial landforms were formed during ice sheet retreat, as favoured by Merritt et al. (2004a, b).

The lack of chronological evidence in the Lake District poses probably the most significant barrier to establishing the extent of LLS glaciers in this region. Absolute dating of glacier

limits to the LLS has only been successfully undertaken at two locations, Lingmell Gill (Ballantyne et al., 2009) and Keskdale (Hughes et al. 2012). Both of these dates relate to relatively small cirque glaciers and, at present, there is no chronological evidence that glaciation during the LLS was not simply restricted to small cirque and valley glaciers. Dating of landforms at the limits of the icefields reconstructed by McDougall (1998, 2001, 2013) would help to distinguish whether the LLS glaciers did form extensive plateau icefields or whether glaciation was more restricted, as suggested by Sissons (1980a). However, given that several cosmogenic isotope dates within this region seem to have been impacted by nuclide inheritance, and thus produced implausible ages, it is possible that alternative dating techniques may be required. Additionally, chronological control (either in the form of cosmogenic isotope dates or basal radiocarbon dates) for sites where authors have been unable to reach a consensus regarding the age of landforms, as at Cotra and Widdygill Foot, would help to determine whether the landforms in these areas were formed during the LLS or by an earlier phase of glaciation.

On the Isle of Arran, there is an apparent discrepancy between the empirical limits of LLS glaciation (Ballantyne, 2007b) and those generated by, otherwise broadly accurate, numerical modelling (Golledge et al., 2008). Although the extent of glaciation is relatively well-constrained by high resolution geomorphological mapping, chronological control on these limits is absent (Figure 3.30). Dating of the more subdued moraines in the lower valleys, compared to the fresher features which lie further up-valley, could help to test between the restricted glaciers proposed by Ballantyne (2007b) and the more extensive icefield supported by Gemmell (1973) and would act as a valuable chronological constraint in an area that currently lacks any LLS dates.

Chronological control is also rather limited within the limits of the West Highland Glacier Complex. Assignation of the moraines at Achnasheen to the LLS resulted in the reconstruction of a substantially more extensive icefield (Bennett, 1991; Bennett and Boulton, 1993a), which has since been reproduced in much of the published literature. However, the evidence for a LLS age was limited to the fact that the Achnasheen moraines appeared to form a continuous sequence with moraines attributed to the LLS further west (Bennett and Boulton, 1993a) and was supported by the absence of Lateglacial sediments within the limits at Achnasheen (Sissons, 1982). Given the importance of this site in determining the configuration of the northern sector of the West Highland Glacier Complex, establishing chronological control on the limits at Achnasheen would represent a significant step in constraining the overall extent of LLS glaciers within this region. Likewise, recent bathymetric surveying (McIntyre et al., 2011) has suggested that LLS

glaciers may have extended further into the lochs along the west coast of Scotland than previously inferred from the onshore evidence (Bennett and Boulton, 1993a). Further surveying is required to determine the outermost position of moraines within these lochs, followed by dating to ensure that these features represent the LLS glacier limits, rather than those of an earlier glacial phase.

In addition to differentiating between LLS and older features, determining the timing of advances within the short-lived LLS, and thus the synchroneity of glacier behaviour, has proved problematic and requires high resolution absolute dating. The pattern of deglaciation during the LLS, as suggested from the growing number of absolute dates on these landforms, is extremely complex, with glaciers in some regions apparently reaching their maximum extent during the mid-stadial while others peaked much later. Radiocarbon dates collected by Bromley et al. (2014) which appear to show that Rannoch Moor, at the centre of the main icefield, was ice-free by 12,262±85 cal yr BP, further complicate this picture of retreat, although the accuracy of these dates has been challenged by Small and Fabel (2016). Only through a systematic and comprehensive program of absolute dating, such as that currently being undertaken for the BRITICE-CHRONO project, will it be possible to determine the timing and synchroneity of LLS glaciation. Most of the existing dating has been conducted near the termini of the LLS glaciers (e.g. Palmer et al., 2010; MacLeod et al., 2011) and thus, comparatively little is known about the behaviour of the ice once retreat was underway. Dating of sites along transects of the major retreat corridors, such as the Lomond basin and Loch Linnhe, and further dating to clarify the timing of deglaciation of Rannoch Moor, would enable assessment of the rate of retreat. Rates could be compared between transects to investigate the timing, speed and nature of retreat across the main icefield as a whole. This would act as a useful constraint of numerical models of the extent and retreat of LLS glaciers. Although the errors associated with most dating techniques are presently too large to permit such detailed analysis, one can expect that technological advances in the dating methodologies will in the future produce sufficiently precise and reliable dates. Cross-matching sufficiently high resolution dates with existing palaeoclimate records for the LLS (e.g. Brooks and Birks, 2000) could be used to assess to what degree climate controlled LLS glacier retreat and whether significant warming periods within the stadial produced a corresponding increase in the rate of glacier retreat. Given the rapid nature of LLS climate change, establishing how this affected the extent and dynamics of LLS glaciers could be used as a potential analogue for modern climate-glacier relationships.

## 3.6 Conclusions

- This paper has reviewed the literature on the evidence for LLS glaciation in Britain, which was recently compiled into a map and GIS database (Bickerdike et al., 2016).

- From this compilation of evidence, the uneven coverage of the mapping becomes clear. Some areas have been mapped extensively whilst others remain largely unmapped. Future work must seek to address such shortfalls in order to allow accurate reconstruction of the extent of LLS glaciation in all areas. Particular areas to target are the southwest Scottish Highlands and the eastern outlet glaciers between Loch Rannoch and Loch Lomond. Bathymetric surveying of sea lochs where LLS glaciers terminated beyond the current coastline, such as Lochs Sligachan, Duich, Nevis, Ainort and Sunart, will be essential in firmly constraining the maximum extents of LLS glaciers at these sites.

- The quality of geomorphological mapping in the literature is highly variable across Britain. Whilst recent studies have produced high resolution data (e.g. Ballantyne, 2002a; Lukas and Lukas, 2006a, b; Bendle and Glasser, 2012; Boston, 2012a) large regions have yet to receive such attention and this lack of detail limits opportunity for further analysis of retreat dynamics. Furthermore, caution should be exercised when using older mapping, which may have been conducted using outdated techniques and glaciological theories, or when examining areas where the glacial origin of features remains uncertain. Remapping of these areas, making use of improved modern aerial imagery, would expand the coverage of reliable, high resolution geomorphological data, offering improved scope for detailed advanced analysis. The specific areas to target include the Gaick Plateau and the southeast Grampian mountains in the central Scottish Highlands, and large areas of Rannoch Moor.

- Numerical modelling has been instrumental in testing the relationship between LLS glaciation and climate. Thermomechanical modelling, forced by a locally scaled GRIP temperature series, has been able to closely replicate the extent of LLS glaciation as inferred from empirically based limits when steep precipitation gradients are imposed on the model. However, local conditions, which were critical to glacier formation particularly in marginal areas, were not included. These factors include the contribution of windblown snow onto glaciers and the presence of pre-existing ice masses from which the LLS glaciers could grow at the onset of LLS cooling. This has led to discrepancies between modelled and empirical LLS glacier limits and promotes uncertainty as to whether LLS ages have been assigned incorrectly to some outlets, for example the Callander lobe (Golledge et al., 2008). Whilst modelling is a powerful tool for determining the extent of LLS glaciers in

areas of uncertainty, these important local factors should be incorporated into future simulations to more realistically represent the controls on glacier formation during the LLS.

- Various styles of glaciation are represented by the geomorphological evidence. These range from the icefield and ice cap that covered much of the western Highlands, to satellite icefields in surrounding upland areas, to valley and cirque glaciers, usually restricted to topographically favourable sites, in marginal locations. Following the proposal that a plateau icefield occupied the hills of the central Lake District during the LLS, there has been a general shift away from alpine to icefield styles of glaciation, as supported by the geomorphological evidence and observations on modern upland ice masses.

- A reconstruction of the lateral extent of the LLS glaciers at their maximum positions has been compiled from empirically based reconstructions from the published literature and the confidence regarding these limits is shown on an accompanying figure. These figures are intended as a tool to guide future work and to refine the reconstruction. The shapefile provides a useful tool for other researchers to investigate controls on LLS glacier formation at a larger scale than previously possible.

- Dating LLS features remains a significant challenge to understanding the extent and dynamics of these glaciers. Although relative dating has become more reliable and multiple lines of evidence are being employed before assigning an age to features, there continues to be a paucity of absolute dates, inhibiting our understanding of the timing and synchroneity of glaciation during the LLS. Whilst the number of absolute ages continues to grow, implementation of a comprehensive dating program will be necessary in order to address these uncertainties. In particular, dating transects along the retreat pathways of major LLS glaciers, such as the Lomond Lobe, could provide a far greater insight into the rate and timing of LLS deglaciation.

# Chapter 4

# Glacial landsystems, retreat dynamics and regional controls on Loch Lomond Stadial glaciation in Britain

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## Abstract

Glacial geomorphology relating to the Loch Lomond Stadial (Younger Dryas) in Britain is used to construct five glacial landsystem models. These landsystems lie on a continuum of increasing ice thickness and decreasing topographic control and typify the principal styles of glaciation during the stadial. The landsystems comprise: the cirque/niche glacier landsystem, the alpine icefield landsystem, the lowland piedmont lobe landsystem, the plateau icefield landsystem, and the ice cap landsystem. The geomorphological features common to each landsystem are identified and process-form relationships are discussed. Spatial patterns in the distribution of these landsystems are then examined: geomorphology representing the ice cap landsystem is present only at the centre of the West Highland Glacier Complex which was flanked primarily by satellite alpine and plateau icefields. The cirque/niche glacier landsystem was present predominantly in areas which experienced conditions only marginally favourable for glacier development at peripheral sites. Three styles of glacier retreat are recorded by the geomorphology: active, two-phase and uninterrupted retreat. Of these, active retreat appears to be most widespread within the LLS limits. These retreat styles reflect a combination of climatic and topographic conditions, coupled with local factors influencing the preservation of landforms from which retreat dynamics can be inferred. Likewise, the distribution of landsystems was influenced by an interplay between topography and climate, with glacier formation being facilitated in locations where topographic conditions aided in the accumulation of snow. The pattern also supports the existence of previously recognised northward and eastward precipitation gradients across Britain during the stadial.

## 4.1 Introduction

The Younger Dryas Stadial, between 12.9 and 11.7 ka, was characterised by an abrupt return to cooler conditions across much of the Northern Hemisphere, following the Last Glacial Maximum (LGM) (de Menocal et al., 2000; Nakagawa et al., 2003; Genty et al., 2006; Golledge, 2010). Although the global net cooling of the event was a moderate 0.6°C (Shakun and Carlson, 2010), the cooling experienced, particularly in regions flanking the North Atlantic Ocean, was sufficient to cause glaciers to readvance or regrow in North America (Lowell et al., 1999), Iceland (Ingólfsson et al., 2010), Scandinavia (Andersen et al., 1995; Mangerud et al., 2011) and Britain (Sissons, 1974; Golledge and Hubbard, 2005; Golledge, 2010; Lukas and Bradwell, 2010; Ballantyne, 2012). In Britain, the period, known locally as the Loch Lomond Stadial (LLS), was characterised by glacier regrowth in the form of a large icefield (over 9000 km$^2$), which covered much of the western Highlands of Scotland, and other satellite icefields, valley and cirque glaciers in surrounding upland areas of Scotland, England and Wales (Golledge, 2010). The LLS glaciers, which were predominantly land-terminating, produced a wealth of glacial geomorphology which has attracted the attention of researchers for over a century (e.g. Darwin, 1842; Jolly, 1868). Much of this research took the form of case studies of geomorphological mapping for specific areas, with authors focusing their efforts on reconstructing the extent and dynamics of these glaciers at local scales (Sissons, 1980a; Ballantyne, 2002a; Lukas and Benn, 2006). This fragmented and spatially inconsistent body of research long inhibited our understanding of the extent and dynamics of LLS glaciers at a regional scale.

Recently, the published literature on the glacial geomorphology of the LLS in Britain was compiled into a GIS database and glacial map (Bickerdike et al., 2016). This database has, for the first time, permitted the geomorphology of the total extent of the LLS glaciers to be assessed at scales varying from within individual valleys to across whole icefields. Furthermore, it has facilitated comparison of the geomorphology between different regions and has allowed patterns in the style of glaciation and the nature of retreat of these glaciers to be identified at a previously unavailable regional scale.

This paper uses the landsystems concept to identify the styles and processes represented by the LLS glacial geomorphology. Landsystems are areas of common terrain attributes, including topography, soil and vegetation, which are different to those of surrounding areas (Evans, 2003a). Landsystems themselves represent the top tier of a classification hierarchy of glacial geomorphology. Individual landforms, such as drumlins, moraines or eskers, are classed as 'elements', and sequences of elements (e.g. recessional moraines or

drumlin fields) are termed 'units'. Thus, landsystems comprise a series of genetically linked units (Evans, 2003a).

Landsystem attributes are influenced by the underlying geology, climate and past erosional and depositional processes and thereby act as process-form models by which the genetic relationships between landforms and sediments can be determined both at local and regional scales. Benn and Evans (2010) and Evans (2003a) have argued that landforms and sediment assemblages are, in part, influenced by a continuum of glaciation styles and dynamics which are themselves influenced by the relative importance of topography and thermal regime. Consequently, a wide range of glacial landsystems have been identified, ranging from polar-continental (Fitzsimons, 2003) and active temperate glacier margins (Evans, 2003b), to plateau icefields (Rea and Evans, 2003), to glaciated valleys (Benn et al., 2003) and surging glaciers (Evans and Rea, 2003). Landsystem type is partly governed by the supply and turnover of both ice and debris, which is itself influenced by climate (Benn et al., 2003). Therefore, by mapping preserved landsystems, and using these in combination with process-form models and modern analogues, it is possible to make inferences about spatial and temporal variations in climate at the time of landform formation (Benn et al., 2003; Evans, 2013). In this context, it should be noted that the timing of the maximum extent of LLS glaciation is controversial and, whilst this paper presents a reconstruction of the maximum extent of the LLS ice masses, chronological control on their termini is extremely limited. Absolute dates suggest that some LLS glaciers reached their maximum extent during the mid-stadial (e.g. Benn et al., 1992; Ballantyne, 2012) or earlier (Bromley et al., 2014) whilst others continued advancing until the end of the stadial (e.g. Palmer et al., 2010; MacLeod et al., 2011; Small and Fabel, 2016).

Landsystem models have rarely been applied to LLS landforms and deposits in Great Britain. Hence, the aims of this study were to identify the type of glacial landsystems and retreat styles represented by the LLS geomorphology and to map the distribution of landsystems at a regional scale in order to determine possible controls on the nature of LLS glaciation across Britain. Specifically, this employs the geomorphology compiled by Bickerdike et al. (2016) to create a series of glacial landsystem models that incorporate the LLS landform signatures and uses these to identify the process-form relationships associated with each of these landsystems. The landsystem models are then used to create a first-order classification of the LLS ice masses and to map the distribution of each landsystem type across Britain. This facilitates the identification of principal glacier

retreat styles which characterised LLS deglaciation and the most significant controls on LLS glacier extent, style and dynamics.


## 4.2 Methods and approach

LLS glacial landsystems were identified from the glacial map compiled by Bickerdike et al. (2016). The map was examined for recurring patterns of genetically linked landform units which were then used to identify five landsystem models. In order to map the distribution of each landsystem type, empirically based reconstructions of the maximum extent of LLS glaciation (e.g. Ballantyne, 2002a; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Boston et al., 2013, 2015) were compiled from the literature into ArcGIS and were digitised as polygon features. This provided a limit to LLS geomorphology, inside of which the landsystems could be identified. The landsystem models were then applied to a first-order classification of each individual ice mass, based on both the landforms present within the limits of each polygon and their topographic setting. For instance, sequences of recessional moraines, ice-marginal meltwater channels and ice-smoothed bedrock were commonly found together in areas that were classified as plateau icefields (McDougall, 2001; Boston et al., 2015).


Typically, a similar combination of landform units is found in each of the landsystems identified. This is potentially a result of the relatively small scale of the LLS glaciers. Large scale glacial landsystems, such as ice sheet or ice stream landsystems, typically occur over hundreds of kilometres (Clark and Stokes, 2003; Golledge, 2007), and thus were beyond the scale of the relatively restricted LLS glaciers. The narrow range of landsystems observed probably also reflects the relatively narrow range of topographic and climatic conditions across Britain during the stadial, in comparison to those found on a global scale. Consequently, not only is the presence of landform units instrumental in classifying the landsystem of each ice mass, but also their distribution in relation to the underlying topography. For example, sequences of recessional moraines are found in four of the LLS landsystem models proposed here, but their distribution in each is different. In the cirque/niche glacier landsystem, recessional moraines are restricted to the floors of cirques and seldom extend beyond the mouth of the cirque itself (e.g. Sissons, 1977a; Carr, 2001; Bendle and Glasser, 2012). In the alpine icefield landsystem, they are found extensively within the valleys (e.g. Ballantyne, 2002a, 2007a, b; Lukas and Lukas, 2006 a, b), whilst in the plateau icefield landsystem recessional moraines extend from the valleys onto the upland plateau surface (e.g. McDougall, 2001, 2013; Boston, 2012a; Boston et al., 2015). Likewise, recessional moraines are found on the high ground of the ice cap landsystem but, unlike those in the plateau icefield landsystem, these features are

discordant with the underlying topography (Golledge, 2007). Thus, it is not only the presence or absence of landform units that informs the landsystem classification of LLS geomorphology, but the distribution of landforms with regard to the glacial limits and underlying relief.

Each ice mass was classified as one of the five landsystem types, depending on which style of glaciation was best represented by the landform assemblages mapped within its limits. This methodology required some generalisation of local scale variations between different areas of the same ice mass but, due to the extensive and complex nature of the LLS ice masses, subdivision of ice masses into individual drainage basins was unfeasible. It was assumed that the reconstructions accurately represent the maximum extent of LLS glaciation, although given the paucity of absolute dates on LLS landforms, some features may predate the LLS. However, given the probable asynchronous nature of the LLS glaciers (Ballantyne, 2012), it should not be inferred that the glacial configuration or landsystem distribution occurred in a single phase during the LLS. Furthermore, the landsystem types developed time-transgressively both within and between catchments, so that the largest ice masses retreated and thinned to become more topographically constrained throughout deglaciation, transitioning from ice caps to icefields, and then to individual cirque glaciers before the ice ultimately disappeared. In most instances, the clearest geomorphological evidence comprises sequences of recessional moraines on the floors and lower slopes of valleys and is thought to relate to the retreat of the LLS glaciers (Bennett and Boulton, 1993a; Ballantyne, 2002a, 2007a; Lukas and Benn, 2006).

## 4.3 Results
Figure 4.1 provides an example of the LLS glacial geomorphology compiled by Bickerdike et al. (2016) and serves as an overview of the types of geomorphological evidence that have informed the continuum of landsystems identified in this paper and illustrated in Figure 4.2. The five landsystems incorporate over 90% of all landform signatures. Occasional landform examples which did not precisely accord with the proposed models are few in number and reflect the influence of localised depositional and erosional processes on the geomorphology. Such signatures are classified by Evans (2013) as either intrazonal or azonal.

The map is dominated by the glacial geomorphology of the West Highland Glacier Complex, which primarily comprises sequences of recessional moraines and areas of hummocky moraine. In the sector to the north of the Great Glen, the orientation of these moraine sequences indicates strong topographic constraint on the retreat of the ice up the

valleys to upland source areas (Bennett, 1991; Bennett and Boulton, 1993a, b). The moraines continue into cirques and up to the flanks of mountains that would have provided favourable source areas, but appear absent from the mountain summits themselves. Conversely, in the vicinity of Rannoch Moor, Glen Lyon and the valleys near the proposed centre of ice accumulation, there are topographically discordant features, including moraines, roches moutonnées and asymmetric till deposits (Golledge and Hubbard, 2005; Golledge, 2007). The geomorphology of the southeastern sector of the ice mass remains largely unmapped and, as such, this area could not reliably be classified as any particular landsystem. To the west, the main ice mass was flanked by a series of satellite icefields on the Western Isles of the Outer Hebrides, Skye, Rùm, Mull and Arran. These icefields comprise interconnected networks of valley glaciers, generally characterised by sequences of recessional hummocky moraines within the valleys (Ballantyne and Wain-Hobson, 1980; Benn, 1990, 1992; Benn et al., 1992; Ballantyne, 2002a, 2007a, b). The vertical limits of these glaciers are often marked by clear drift limits or trimlines separating areas of ice erosion and/or LLS drift from areas with widespread evidence of periglaciation. Peripheral to the main valley glaciers, smaller cirque glaciers existed, as marked by short sequences of recessional moraines, drift or boulder limits (e.g. Sissons, 1977a). Additional icefields developed to the south and east of the ice mass, but the geomorphological signature of many of these differs from those on the islands in that moraines and ice-moulded features extend onto the upland areas themselves, which sometimes host sequences of ice-marginal meltwater channels (Sissons, 1974; McDougall, 2001, 2013; Boston et al., 2013, 2015). In peripheral areas, glaciation was restricted to cirques or topographic hollows at the base of escarpments.
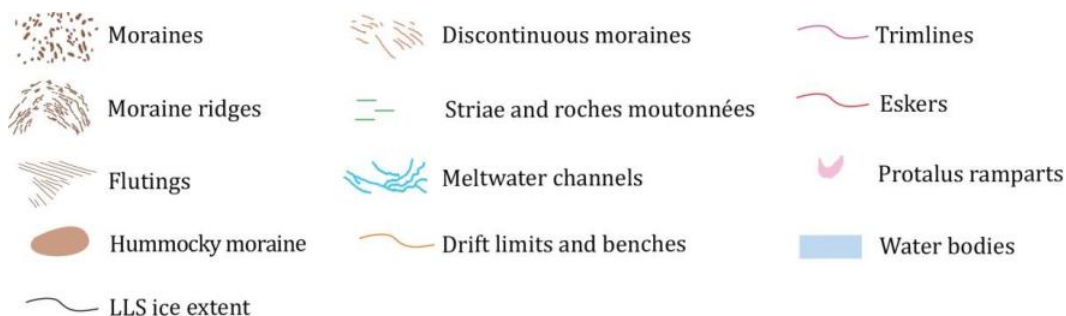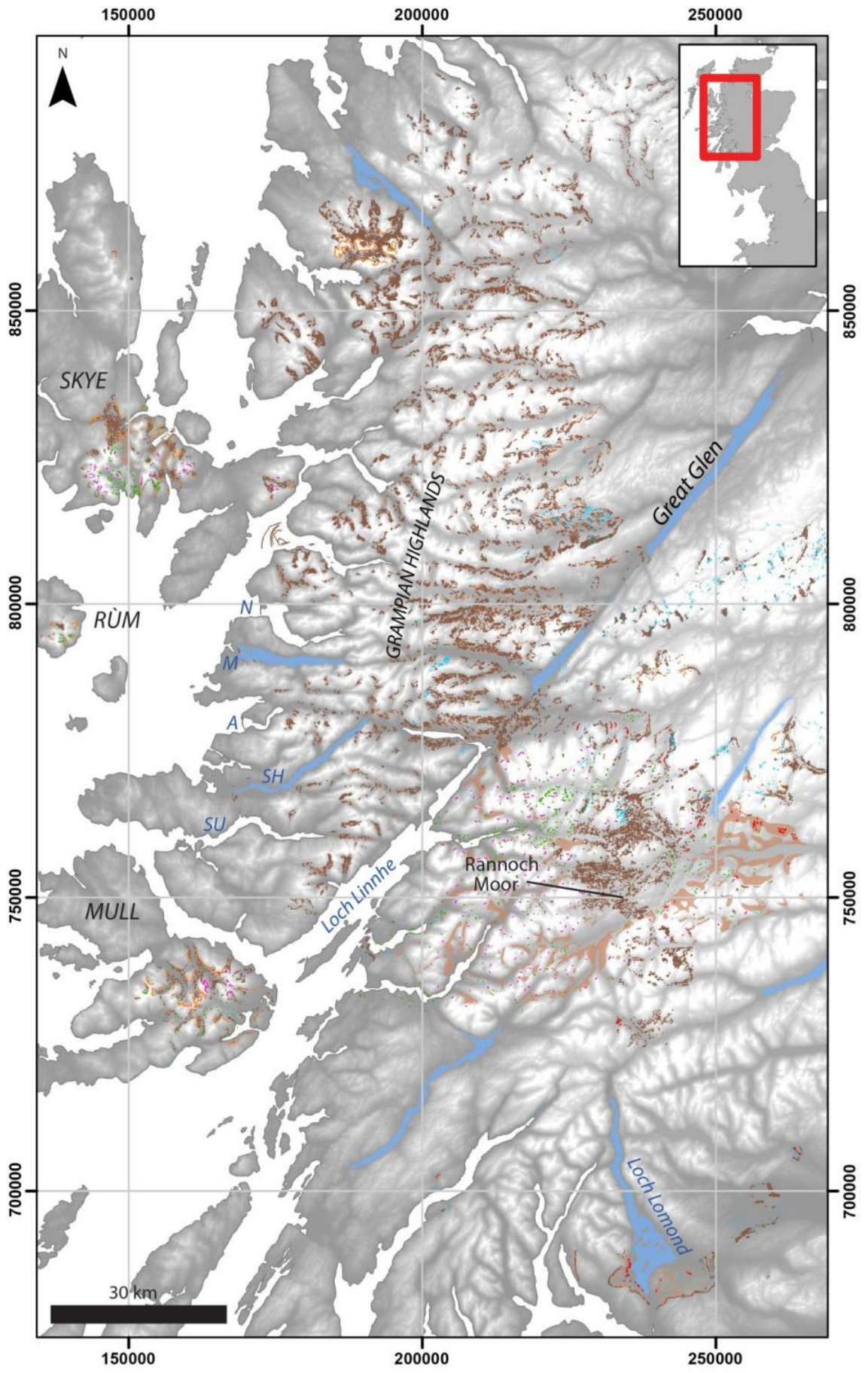
| | | |
|---|---|---|
| Moraines | Discontinuous moraines | Trimlines |
| Moraine ridges | Striae and roches moutonnées | Eskers |
| Flutings | Meltwater channels | Protalus ramparts |
| Hummocky moraine | Drift limits and benches | Water bodies |
| LLS ice extent | | |

| ICE CAP | PLATEAU ICEFIELD | LOWLAND PIEDMONT LOBE | ALPINE ICEFIELD | CIRQUE/ NICHE GLACIERS |
|---|---|---|---|---|
| Glacier size and thickness | | | | Topographic constraint |

**ICE CAP**

- topographically discordant flow.
- varied topography including alpine-style mountains and topographic basins.
- a few nunataks with frost-shattered bedrock on their summits.
- widespread ice-smoothed bedrock including on the high cols and plateau surfaces.
- roches moutonnées on valley floor and lower slopes which transition to whalebacks at higher elevations.
- LGM moraines overridden by LLS flutings, including on the plateau surface.
- asymmetric till deposition within valleys, particularly thick accumulations at the foot of reverse slopes or in topographic hollows.
- sharp-crested moraines along valley floors.
- e.g. Glen Lyon and the surrounding valleys, Scotland.

**PLATEAU ICEFIELD**

- topographically concordant flow.
- upland gently undulating plateau of rounded summits, flanked, and sometimes incised, by valleys.
- some nunataks with blockfield developed on their summits.
- ice-moulded bedrock, roches moutonnées and moraines at the heads of valleys.
- substantial networks of ice-marginal meltwater channels, particularly on the plateau.
- sequences of (recessional) hummocky moraines in the valleys.
- some valleys with an absence of evidence.
- e.g. Monadhliath mountains, Scotland.

**LOWLAND PIEDMONT LOBE**

- topographically concordant flow.
- open and gently undulating, low relief terrain at the foot of glacial valleys.
- mountain ridges and summits exposed above glacier surface to periglacial conditions.
- drift-mantled slopes in middle reaches of valley, though evidence often obscured by afforestation.
- prominent terminal moraine ridges which may enclose areas of chaotic hummocky moraine, kames and kame terraces and eskers.
- evidence of proglacial glacitectonism from the advance of LLS ice over pre-existing deltaic sediments.
- thrust-block moraines and hill-hole pairs, possibly indicating glacier surging.
- e.g. the Lomond valley, Scotland

**ALPINE ICEFIELD**

- topographically concordant flow.
- alpine-style mountains separated by cols and steep-sided glacial valleys. Well-developed cirques. Some limited plateau areas.
- mountain summits and ridges exposed to periglacial conditions as inferred from blockfields and solifluction features.
- ice-smoothed cols restricted to the icefield centre or low elevations.
- vertical extent of ice marked by trimlines or drift limits.
- transition along valleys from erosional zone in upper valleys (roches moutonnées, striae, flutings) to depositional zone with sequences of recessional moraines and thick till.
- limited evidence of meltwater routing.
- e.g. Isle of Mull, Scotland.

**CIRQUE/ NICHE GLACIERS**

- topographically concordant flow.
- cirques in alpine mountain settings, or the flanks of escarpments, marginal to icefield sites.
- blockfields, frost-weathered debris and solifluction features on summits. Mature talus slopes beyond glacier limits.
- glacier development usually restricted to topographically favourable sites with low insolation and/or snow-contributing areas upwind of them.
- single, arcuate terminal moraine ridges often enclosing an area of hummocky drift, or short sequences of recessional moraines.
- protalus ramparts marking the extent of permanent snowbeds.
- mounds and ridges formed from rockslope failures, which may have a similar appearance to moraine ridges.
- e.g. Snowdonia, Wales.

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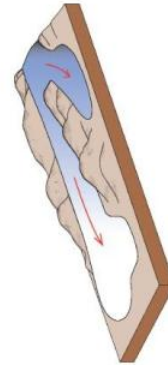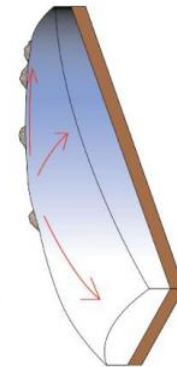
Figure 4.1 Example of the LLS glacial geomorphology compiled by Bickerdike et al. (2016). The map is dominated by moraines (brown), particularly the extensive sequences of recessional moraines that are widespread throughout the northern sector of the West Highland Glacier Complex. Satellite icefields formed on the isles of Skye and Mull and on upland areas peripheral to the main ice mass on the Grampian Highlands. British coastline and present-day waterbodies reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence). GB SRTM Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA.

Figure 4.2 The landsystem continuum represented by LLS glacial geomorphology. The glaciers decrease in size and thickness from the ice cap to the cirque/niche glacier system and become increasingly topographically constrained, as demonstrated by the conceptual flow diagrams. Key characteristics of each landsystem are noted alongside a location where each landsystem has been observed in Britain. Sections 4.3.1 to 4.3.5 describe the evidence for each landsystem.

### 4.3.1 The cirque/niche landsystem

The cirque/niche landsystem (Figure 4.3) is primarily found in locations that were peripheral to the main centres of glaciation during the LLS. Glaciers in such areas were generally present in cirques or topographic hollows, particularly those that faced between north and east (Evans, 1994, 2015a). The maximum extent of glaciers which comprise this landsystem are typically demarcated by sequences of recessional moraine ridges (1 on Figure 4.3), which sometimes extend back into what would have been the source areas of the glacier, but which usually form a belt of moraines near the terminus, or comprise single prominent arcuate terminal moraine ridges. These moraines frequently enclose areas of hummocky drift mounds (2) with no obvious pattern apparent in their distribution within the glacier limits. Protalus ramparts (3), ridges of angular debris, may be present along the foot of escarpments or on steep valleys sides, whilst debris accumulations formed by paraglacial rockslope failures (4) may also be present on the backwalls of slopes and in the valleys below. Relict periglacial features, such as frost-weathered detritus and blockfields (5) can be found on mountain and escarpment summits.

The build-up and movement of ice at these sites was controlled by the underlying topography, with ice flowing from the backwalls of cirques towards the glacier margins. Recessional moraines marking the extent of these glaciers indicate that, in most instances, retreat was active, at least initially, with glaciers remaining near to climatic equilibrium (see Section 4.4.1.1). Conversely, at sites where single arcuate terminal moraines are found, the glacier probably remained at its maximum extent for a sustained period, before undergoing uninterrupted retreat later in the stadial. Cirques in which recessional moraines stretch to the backwall are less common, although they do occur in some areas, such as Cwm Idwal in Snowdonia, Wales (Bendle and Glasser, 2012). However, it is

difficult to determine whether the relative scarcity of cirque/niche glacier sites where moraine sequences extend to the backwall results from the dominance of an uninterrupted style retreat in the later stages of deglaciation, or simply reflects that debris could not accumulate on the steeply dipping backwall slopes of the cirque.

At sites where snow accumulation was insufficiently thick for glacier development, perennial snowbeds formed, over which debris from the cirque backwall or escarpment slid or rolled to produce protalus ramparts. In several locations, such as Fan Hir in the Brecon Beacons, Wales, difficulty in differentiating between single terminal moraines and protalus ramparts arises from their similar morphology (Shakesby and Matthews, 1993; Carr and Coleman, 2007a). Similar difficulties have been encountered when differentiating between glacial deposits and debris ridges formed by paraglacial rockslope failures, particularly when these occur within cirques (Carr et al., 2007b; Mills and Lukas, 2009). Such rockslope failures in Britain have been attributed, primarily, to enhanced seismic activity during periods of accelerated glacio-isostatic uplift as the crust responded to removal of the Late Devensian Ice Sheet (Ballantyne et al., 2014).

Relict periglacial features are frequently found as components of the cirque/niche glacier landsystem, with frost-weathered detritus and blockfields being present on many summits and mature talus slopes mantling the valleys sides. Such features have been interpreted to be indicative of areas that remained exposed to severe cold conditions during the stadial, rather than being protected beneath glacier ice. Consequently, their mutually exclusive relationship with glacial landforms has aided in the reconstruction of the limits of LLS glaciation in numerous studies (Sissons, 1974; Benn and Ballantyne, 2005; Finlayson, 2006). In some instances, relict periglacial features have been found inside the limits of LLS glaciation where these features have been protected beneath cold-based ice (e.g. Boston et al., 2015). However, given the evidence in many locations for the restriction of ice in the cirque glacier landsystem to topographic hollows (e.g. Carr, 2001; Shakesby, 2007), it seems unlikely that such areas of cold-based ice would have been present in these locations. Therefore, the use of periglacial features to identify the LLS glacier limits in these localities would appear reliable.

The cirque/niche landsystem is perhaps best exemplified in the mountains of Snowdonia, Wales. In this area, a series of 38 LLS cirque glaciers have been reconstructed in topographically favourable sites on the flanks of the mountains (Bendle and Glasser, 2012), 18 of which are shown in Figure 4.3b. The geomorphological evidence for these glaciers comprises primarily moraines, ranging from large individual terminal moraines,

as exhibited at Marchlyn Bach and Mawr, to sequences of recessional moraines which extend to the cirque backwalls, as at Cwm Idwal. At other sites, such as Cwm Tryfan, recessional moraines form a belt near the former glacier terminus but are not present in the middle and upper cirque areas, which are dominated by ice-moulded bedrock. There is a trend in the aspect of these glaciers, with all but two facing between north and east. Beyond the margins of LLS glaciation, protalus ramparts have been identified, either high on the walls of cirques or on the convex flanks of mountains, indicating that perennial snowbeds developed at these sites. Although not mapped in this study, mature scree and talus are widespread on the upper valley slopes but are truncated by drift deposits within the limits of LLS glaciation (Bendle and Glasser, 2012). Likewise, frost-weathered debris is common on the mountain summits, indicating that these were ice-free during the stadial, or were covered by cold-based ice.

a
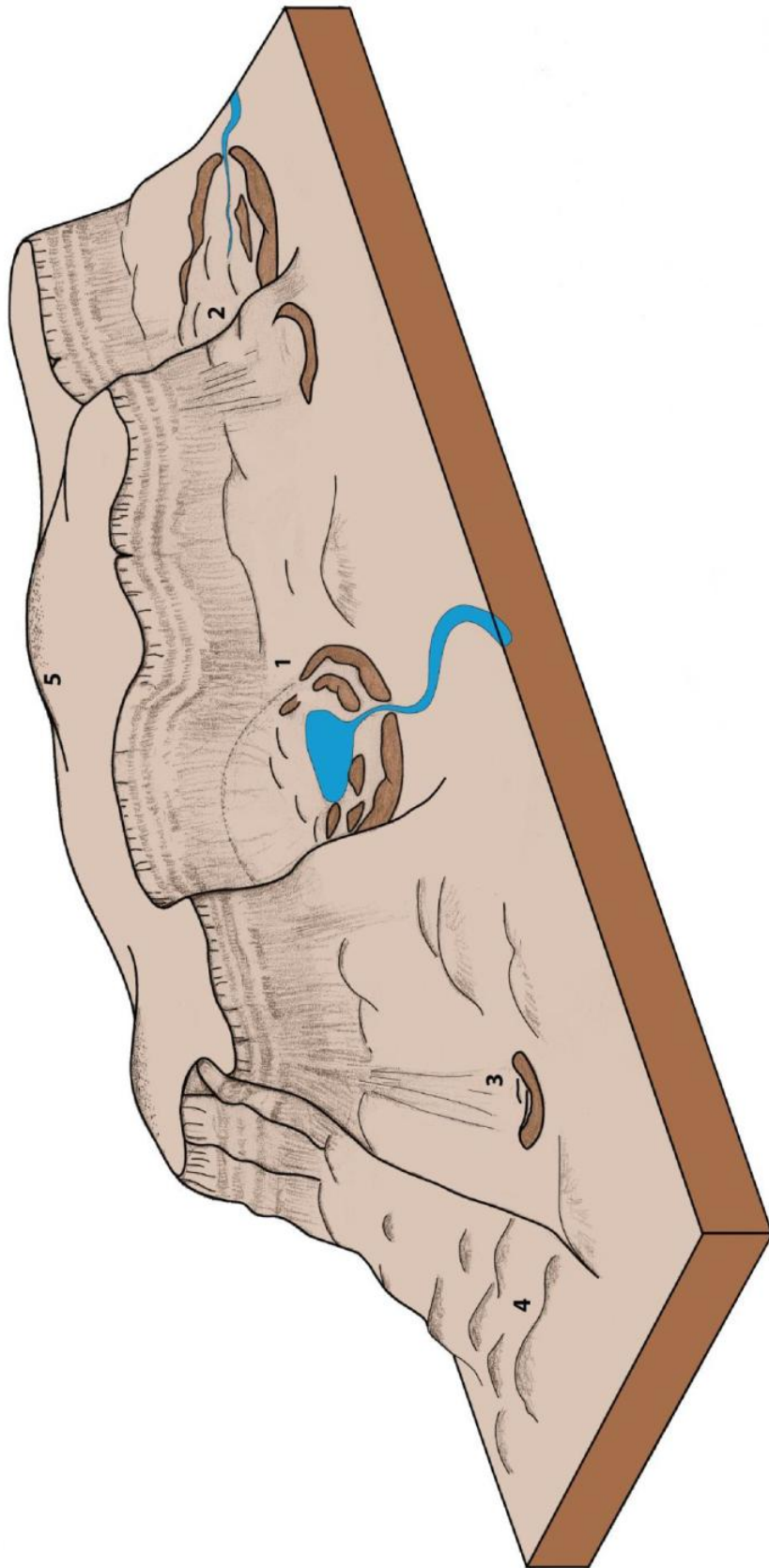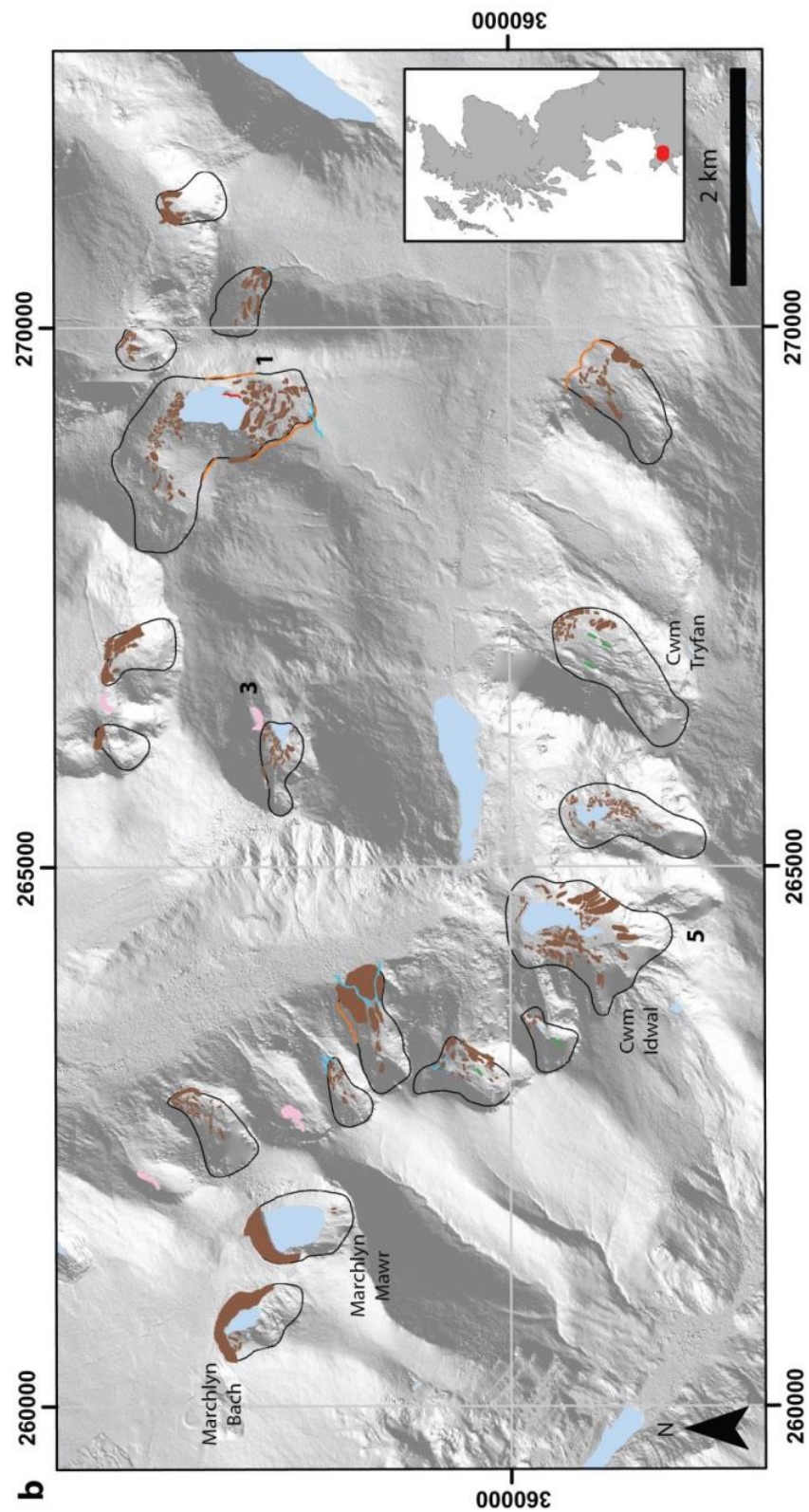
Figure 4.3 The cirque/niche glacier landsystem. a) Conceptual model showing the main characteristics of the cirque/niche glacier landsystem, which is found predominantly in areas where conditions were only just above the threshold required for glaciation. 1. Recessional moraines, 2. Hummocky drift, 3. Protalus rampart, 4. Rockslope failure debris, 5. Blockfield. b) Example of the LLS cirque glacier landsystem in Snowdonia, Wales (after Bendle and Glasser, 2012). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

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### 4.3.2 The alpine icefield landsystem

The alpine icefield landsystem (Figure 4.4) represents one end of the mountain icefield landsystem continuum, the other being the plateau icefield landsystem. It is widespread throughout upland Britain in association with LLS glaciation. Areas characterised by the alpine icefield landsystem typically comprise a series of steep-sided, glacial valleys, sometimes connected by cols at their heads, which are separated by arêtes and spurs. Deep, well-developed cirques flank these mountain ridges, whilst plateau surfaces are limited or absent. The alpine icefield landsystem is characterised by sequences of recessional moraines (1 on Figure 4.4) arranged in arcuate chains, trending obliquely across the valley floors and lower slopes. Small patches of more chaotically arranged morainic mounds (2) can be observed within these moraine sequences but are not widespread and, likewise, eskers (3) are occasionally present on the valley floors. Medial moraines (4) may mark the confluence of two neighbouring glaciers over a topographic spur and reflect concentrations of supraglacial debris in the ice, but these features are relatively rare (Bennett, 1991). In some areas, particularly cirque-headed valleys, recessional moraines are restricted to a depositional zone near the glacier terminus (5) whilst further up-valley a transition occurs to thin till, sometimes moulded into flutings (6). Nearer the valley heads, erosional features, including roches moutonnées (7), dominate. Ice-smoothed bedrock is frequently present in the upper valleys, including over many of the high cols (8), and in many places the maximum elevation of glacial erosion is often marked by clear trimlines (9) between the zone of ice moulding on the lower valley sides and periglacial features above.

During the LLS, numerous alpine mountain areas were occupied by interconnected networks of glaciers which developed in the valleys and cirques, with both glacier morphology and debris transport pathways being highly influenced by the underlying topography (Benn et al., 2003). The relief of alpine topography represented in Britain is relatively low compared that of other alpine mountain regions, such as the Himalaya or Alps, with the valley floors usually only a few hundred metres lower than the mountain summits. Thus, the contribution of supraglacial debris from steep slopes onto these glaciers is comparatively less important (Benn and Lukas, 2006).

The restricted patches of chaotic moraine ridges probably indicate that localised stagnation occurred at the margins of actively retreating glaciers (c.f. incremental stagnation of Eyles, 1979, 1983; Bennett and Evans, 2012), possibly due to the decay of sediment covered ice cores (Benn, 1992). Flutings further up-valley are often found in areas where wet-based ice would have accelerated down steep sections of the underlying

topography, efficiently reworking the subglacial sediment to form these elongate features parallel to ice flow direction (Benn, 1990; Benn and Lukas, 2006). A zone of erosion is generally found at the valley heads, comprising ice-moulded bedrock and roches moutonnées, supporting the assertion that the LLS glaciers were warm-based in their source areas. Likewise, evidence for glacial erosion across the lower cols has been cited as evidence that these were overridden in several instances, allowing individual valley glaciers to coalesce to form an icefield configuration rather than just a series of independent valley glaciers. Trimlines between this ice-scoured terrain and periglacial features (e.g. blockfields, solifluction features) above have aided in reconstruction of the accumulation areas of glaciers in alpine icefields, particularly where these transition into drift limits in the zone of deposition (e.g. Ballantyne, 2002a, 2007a, b; Lukas and Lukas, 2006a, b). It is unfortunate that much of the present geomorphological mapping of the West Highland Glacier Complex has focused on the depositional evidence (Bennett, 1991; Bennett and Boulton, 1993a, b) whilst erosional features remain largely unmapped, making it difficult to reconstruct the vertical limits of much of the icefield.
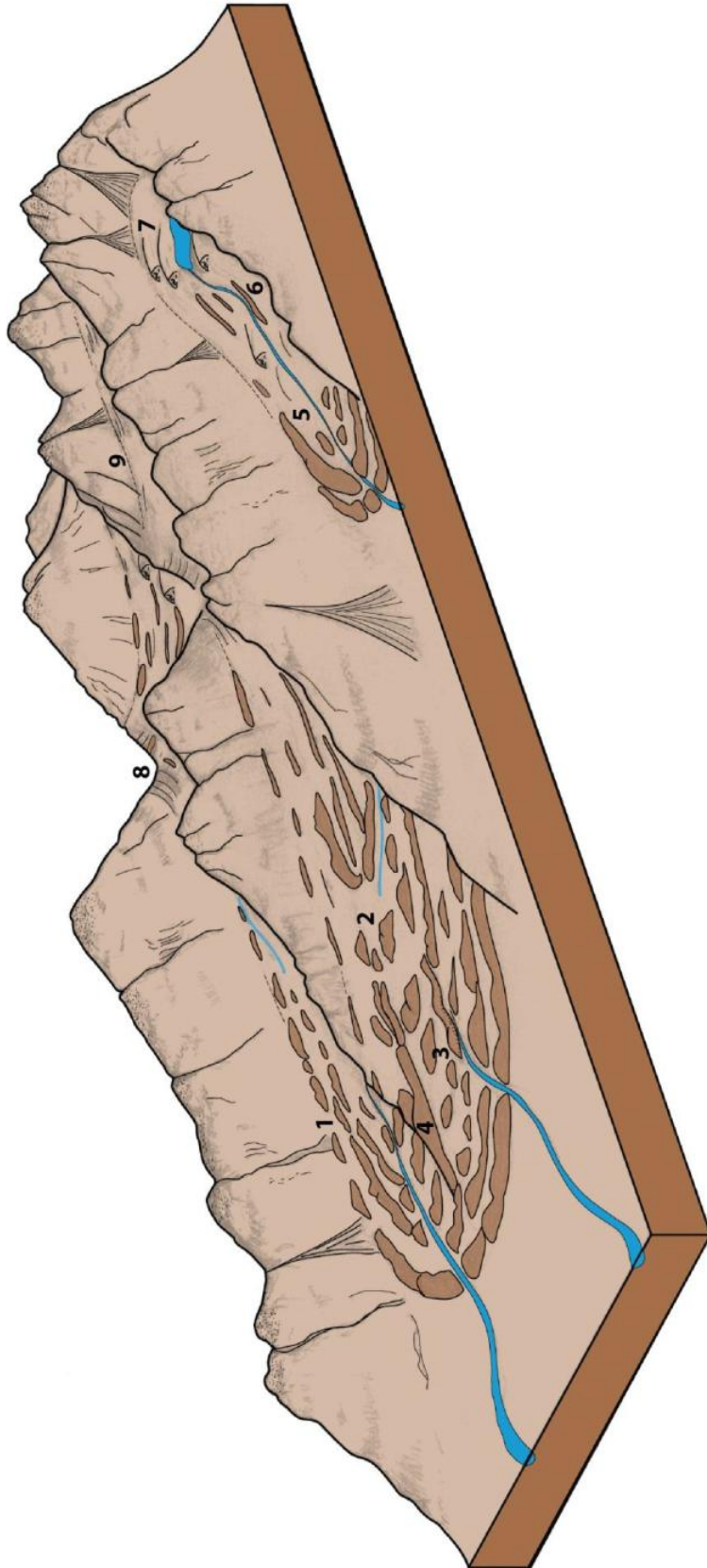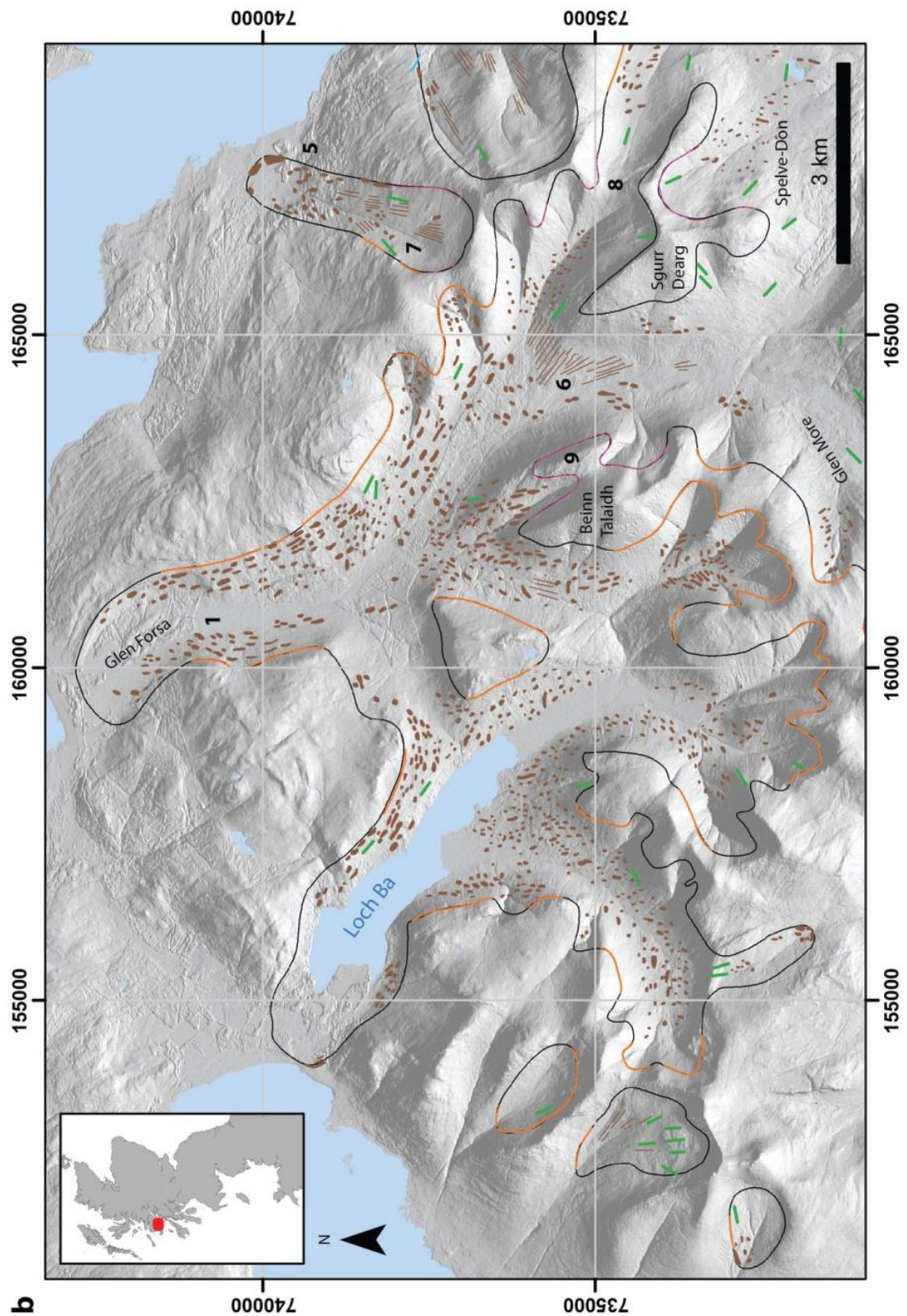
Figure 4.4 The alpine icefield landsystem. a) Conceptual model of the alpine icefield landsystem, found in areas of steep alpine topography where interconnected networks of valley glaciers develop. The extent of glaciation is highly influenced by topography with most mountain summits remaining exposed above the ice. 1. Recessional moraines, 2. Chaotic moraine mounds and ridges, 3. Esker, 4. Medial moraine, 5. Restricted belt of end moraines, 6. Flutings, 7. Roches moutonnées, 8. Ice-smoothed bedrock, 9. Periglacial trimline.  b) Example of the LLS alpine icefield landsystem on the Isle of Mull, Scotland (after Ballantyne, 2002a). Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

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The Isle of Mull provides a good example of the geomorphology of the alpine icefield landsystem, as shown in Figure 4.4b. The extents of the Ba and Forsa glacier lobes are marked by a near-continuous spread of clear recessional hummocky moraines from the terminus to the upper cirques, indicating that these glaciers remained close to climatic equilibrium throughout deglaciation (Ballantyne, 2002a). Conversely, much of the area occupied by the Spelve-Don glacier is free of recessional moraines, suggesting that retreat of this lobe was largely uninterrupted until it reached the high cirques where the presence of recessional moraines suggest retreat in equilibrium with climatic conditions was re-established (Ballantyne, 2002a). The presence of flutings is clearly visible in the cirques which fed the Forsa glacier, formed by wet-based erosive ice flow over the drift. In numerous valleys, thick mantles of drift within the LLS glacier limits contrast with drift-free slopes above, allowing the vertical extent of ice to be established. Evidence of ice-scouring is present in the cols which connect the source areas of the Forsa, Glen More and Spelve-Don outlet glaciers, but it is clear from drift limits and the trimlines which mark the surface of the ice on the slopes of Beinn Talaidh and Sgurr Dearg that most of the mountain ridges and summits remained above the ice surface.

### 4.3.3 The lowland piedmont lobe landsystem

The mountain icefield landsystem is best represented by the glacial geomorphology at the termini of the Lomond and Menteith glaciers. Although there is insufficient geomorphological mapping for the upper valleys of these glaciers, their termini are well marked in the geomorphological record and represent examples of lowland piedmont lobes and their landsystem signature. In contrast to the majority of outlet glaciers from the West Highland Glacier Complex, the Lomond and Menteith lobes appear not to have produced extensive sequences of recessional moraines (1 on Figure 4.5). Instead, their termini are, at least partially, marked by proglacial glacitectonic landforms (including thrust-block moraines and hill-hole pairs, 2) which appear to be otherwise largely absent from the LLS landform record, with the possible exception of the Spelve-Don lobe on the Isle of Mull (Benn and Evans, 1993). The area inside the prominent moraine ridge which marks the maximum limit of the LLS Lomond glacier (3), is characterised by chaotic hummocky moraine (4), kames and kame terraces (5) and eskers (6).

During the LLS, the Lomond glacier advanced and blocked drainage down the Endrick and Blane valleys, forming proglacial Lake Blane (Evans and Rose, 2003a). Sediment exposures at Drumbeg quarry indicate that the Lomond glacier retreated from proglacial Lake Blane, forming an ice-contact Gilbert-type delta before undergoing a significant readvance into the delta (Benn and Evans, 1996; Phillips et al., 2002, 2003) (Figure 4.5). During this readvance, the deltaic sediments were folded and thrust into a series of composite ridges

(arcuate suites of sub-parallel ridges, formed from upthrust sediment, with intervening depressions), which were then overridden by the Lomond glacier, leading to glacitectonic disturbance of the upper deltaic sediments and the superimposition of a glacitectonite and subglacial till above them (Benn and Evans, 1996; Evans and Rose, 2003b). The outer limit of the Menteith lobe is marked by a large moraine ridge, immediately inside of which sits the Lake of Menteith. These features have been interpreted as a hill-hole pair, where the Menteith glacier excavated marine clays, sands and gravels to form the lake basin and thrust and compressed these sediments to form the neighbouring hill (Wilson, 2005; Evans and Wilson, 2006b). Both hill-hole pairs and composite ridges can be associated with surging glaciers, but neither feature is individually diagnostic of this landsystem (Evans and Rea, 1999) because they have also been linked to the dislocation of permafrost by polar and sub-polar glaciers (Evans and England, 1991; Evans and Rose, 2003b). However, crevasse-fill ridges, which Smith (1993) suggested were present within the Menteith glacier limits, are associated with surging glaciers, as are chaotic hummocky moraine (3) and kames (4) which form during *in situ* stagnation of ice between surges (Evans and Rea, 1999); these landforms are present within the limits of the Lomond glacier (Rose, 1980, 1981) and hence it is possible that the Lomond and Menteith piedmont lobes were prone to surging during the LLS (Evans and Wilson, 2006b).
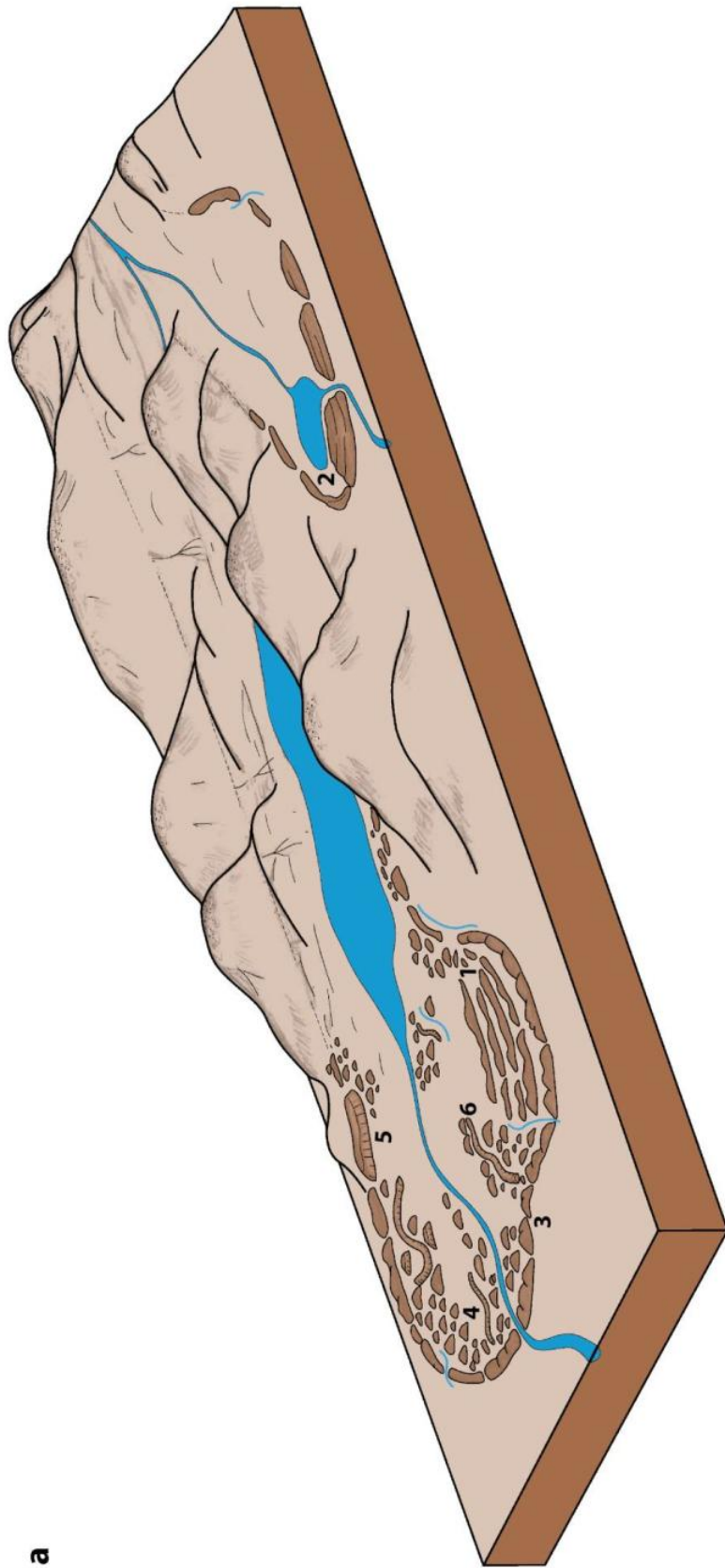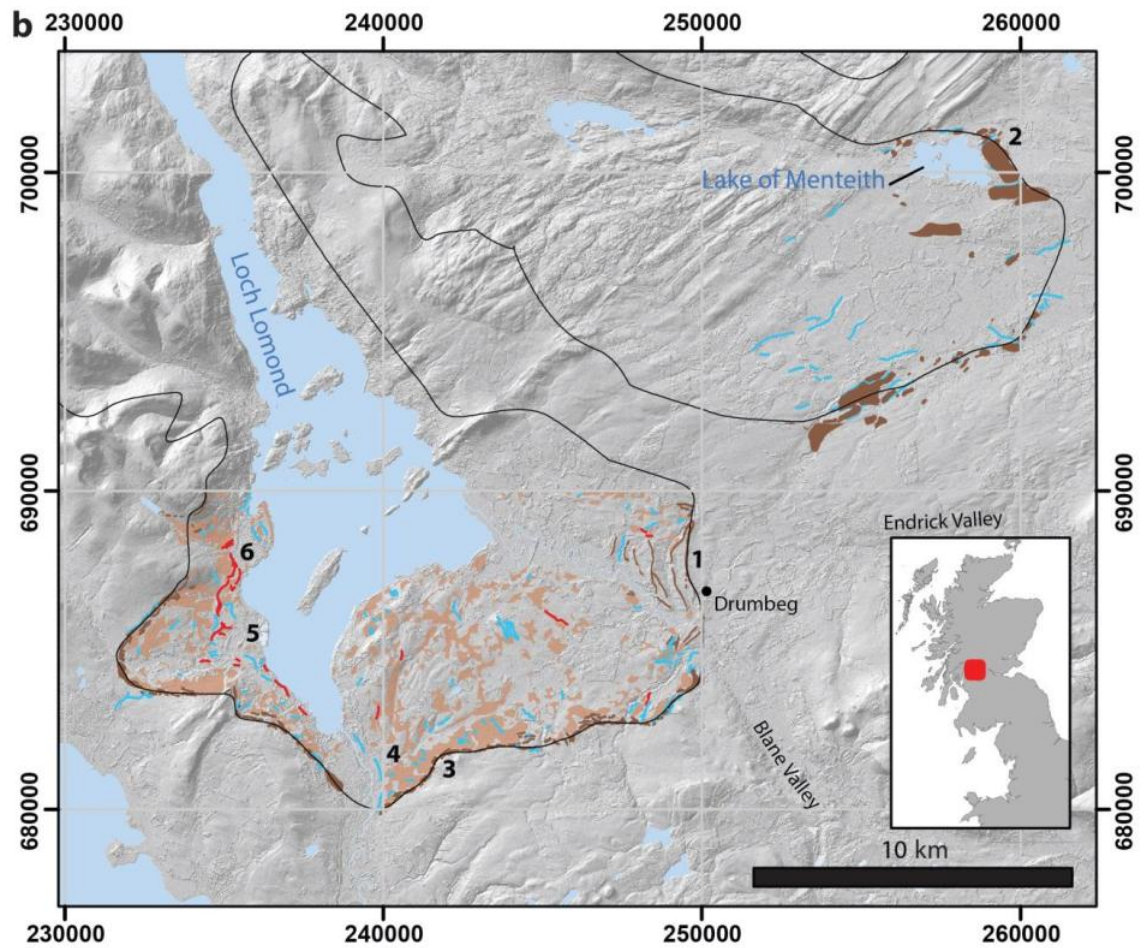
Figure 4.5 The lowland piedmont lobe landsystem. a) Conceptual model of the lowland piedmont lobe landsystem. 1. Recessional moraines, 2. Hill-hole pair, 3. Terminal moraine ridge, 4. Chaotic hummocky moraine, 5. Kame terraces, 6. Eskers. b) Example of the LLS lowland piedmont lobe landsystem, the Lomond and Menteith Valleys, Scotland (after Rose, 1980, 1981; Wilson, 2005). Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Observation Data Centre. Key as in Figure 4.1.

### 4.3.4 The plateau icefield landsystem

Plateau icefield landsystems (Figure 4.6) fall at the opposite end of the mountain icefield landsystem continuum to the alpine icefield landsystem, and form more extensive ice masses, that are less constrained by topography. The topography of these regions is dominated by gently undulating, upland plateaux of rounded mountain summits, flanked, and sometimes dissected, by deep glacial valleys. The heads of these valleys may have steep backwalls or may rise gently so that the valley transitions into the plateau surface. The outlet valleys of the plateau icefield landsystem are similar in their geomorphology to those of the alpine icefield. Sequences of recessional moraines (1 on Figure 4.6) can be found on the floors and many of the lower valley sides. In some instances, these moraines can be traced back up onto the plateau surface at the valley heads (2), particularly where backwalls are absent. Ice-moulded bedrock and roches moutonnées may also be present at the valley heads, descending from the plateau surface into the valleys. Much of the evidence on the plateau surface itself can be subtle, particularly in locations where the plateau surface supports a thick covering of peat. Ice-marginal meltwater channels run between moraine ridges in the valleys but can extend onto the plateau, where they may form extensive networks (4). Periglacial features such as blockfields and solifluction lobes (5) may be present on the higher summits of the plateau.

Accurate mapping of this geomorphology has been instrumental in a shift in the paradigm of the style of LLS glaciation in numerous regions. Earlier studies favoured a cirque or valley style of glaciation for all the satellite icefields which flanked the West Highland Glacier Complex, such as in the English Lake District (Sissons, 1980a) and on the Beinn Dearg massif (Sissons, 1977a). However, the presence of moraines at the valley heads, leading onto the plateaux prompted remapping of the geomorphological evidence and it has subsequently been proposed that many valley glaciers were, in fact, outlet glaciers fed by glaciated plateau areas (Rea et al., 1998; McDougall, 1998, 2001; Finlayson et al. 2011). In some cases, plateau ice provides the only possible accumulation area for glaciers in the surrounding valleys, the extents of which are well-constrained by geomorphology (e.g. Rea et al., 1998; Rea and Evans, 2003, 2007; McDougall, 2013). Such a configuration is further supported by down-valley aligned ice-moulding and roches moutonnées at the edge of the plateau where glaciers accelerated as they flowed down steeper slopes.

In light of this evidence, ice within the valleys was probably warm-based on account of the widespread evidence for glacial transport and deposition of debris (Benn and Lukas, 2006). In contrast, the presence of ice-marginal meltwater channels on the plateau surface, as observed on many of the plateau surfaces in the Gaick (Sissons, 1974) and

Monadhliath (Boston, 2012a), represents the drainage of meltwater to the margins of the glacier, suggesting that subglacial drainage was inhibited by basal ice being frozen to the underlying bed at least some of the melt season (Ó Cofaigh et al., 2003; Rea and Evans, 2003). Where nested sequences of these meltwater channels are found, these mark successive retreat positions (Dyke, 1993; Ó Cofaigh et al., 1999; Atkins and Dickinson, 2007). The predominance of cold-based, non-erosive ice on the plateau is supported by the presence of periglacial features such as blockfield and solifluction lobes, which are present on the higher summits of the plateau, including areas that, according to empirically based reconstructions (e.g. Boston et al., 2015), were overridden by ice during the LLS. However, the presence of recessional moraines in topographic hollows on the plateau implies that pockets of warm-based ice existed in favourable sites. Thus, the geomorphological signature of plateau icefields suggests a complex thermal regime, characterised by relatively thin, cold-based ice on the plateau surfaces themselves but thicker, warm-based ice in the surrounding areas (Boston et al., 2015).

The LLS glacial geomorphology of the Monadhliath Mountains (Figure 4.6b) is characteristic of a plateau icefield. Clear sequences of recessional hummocky moraines are present on the valley floors and lower slopes and continue up onto the plateau surface, such as in Coire nam Beith, Coire Fionndrigh and in the upper Findhorn Valley (Boston, 2012a). Sometimes, these moraines are found in association with ice-marginal meltwater channels (Boston, 2012a), suggesting that these lobes were either warm-based or polythermal, whereas in other valleys the presence of ice-marginal meltwater channels alone, indicates that LLS ice was cold-based (Boston et al., 2013, 2015). Areas of blockfield and solifluction lobes are widespread on the plateau, inside the reconstructed limits of LLS glaciation, but are thought to have been protected beneath cold-based, non-erosive ice, rather than exposed as nunataks above the ice surface (Boston et al. 2015).
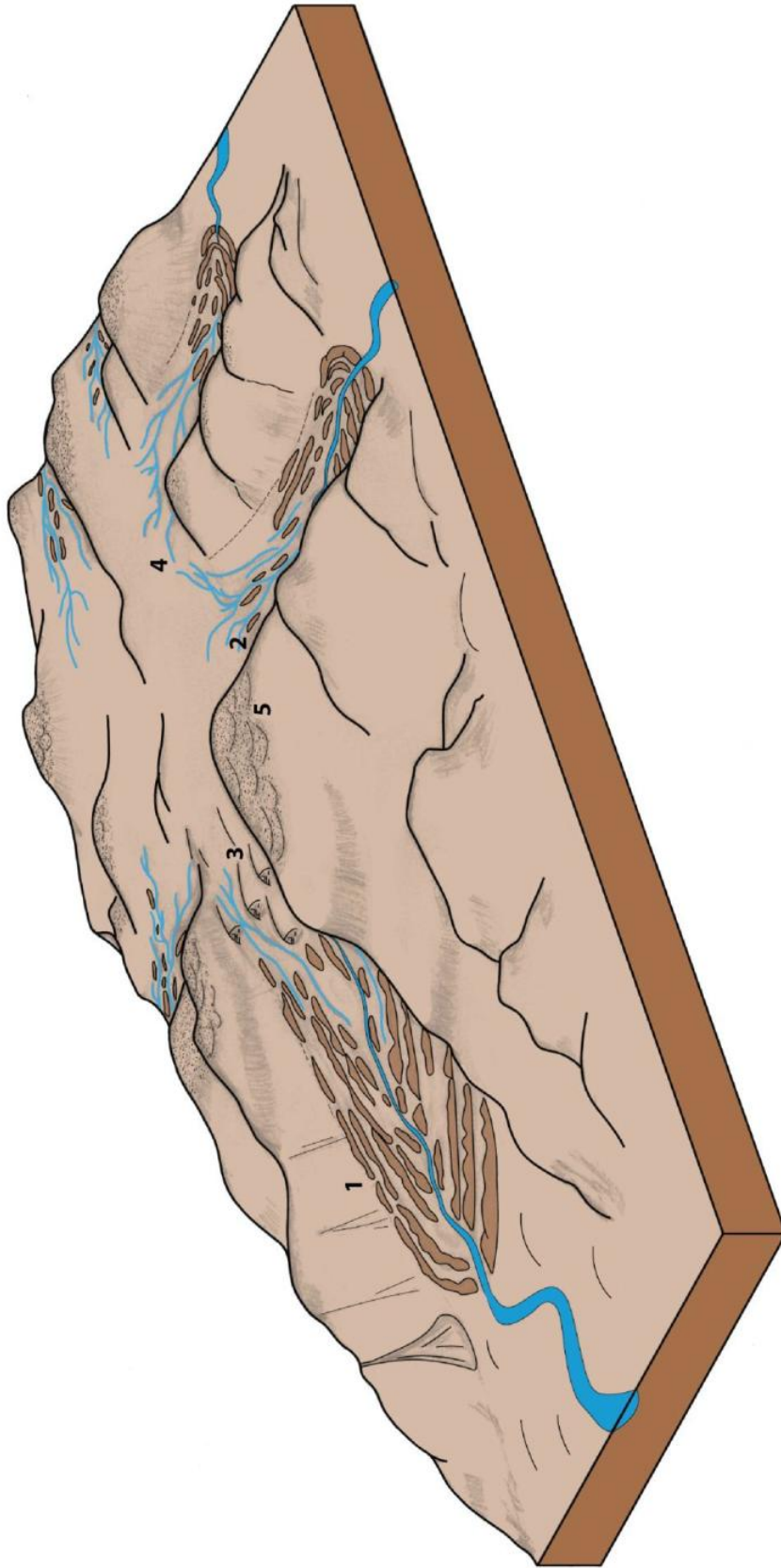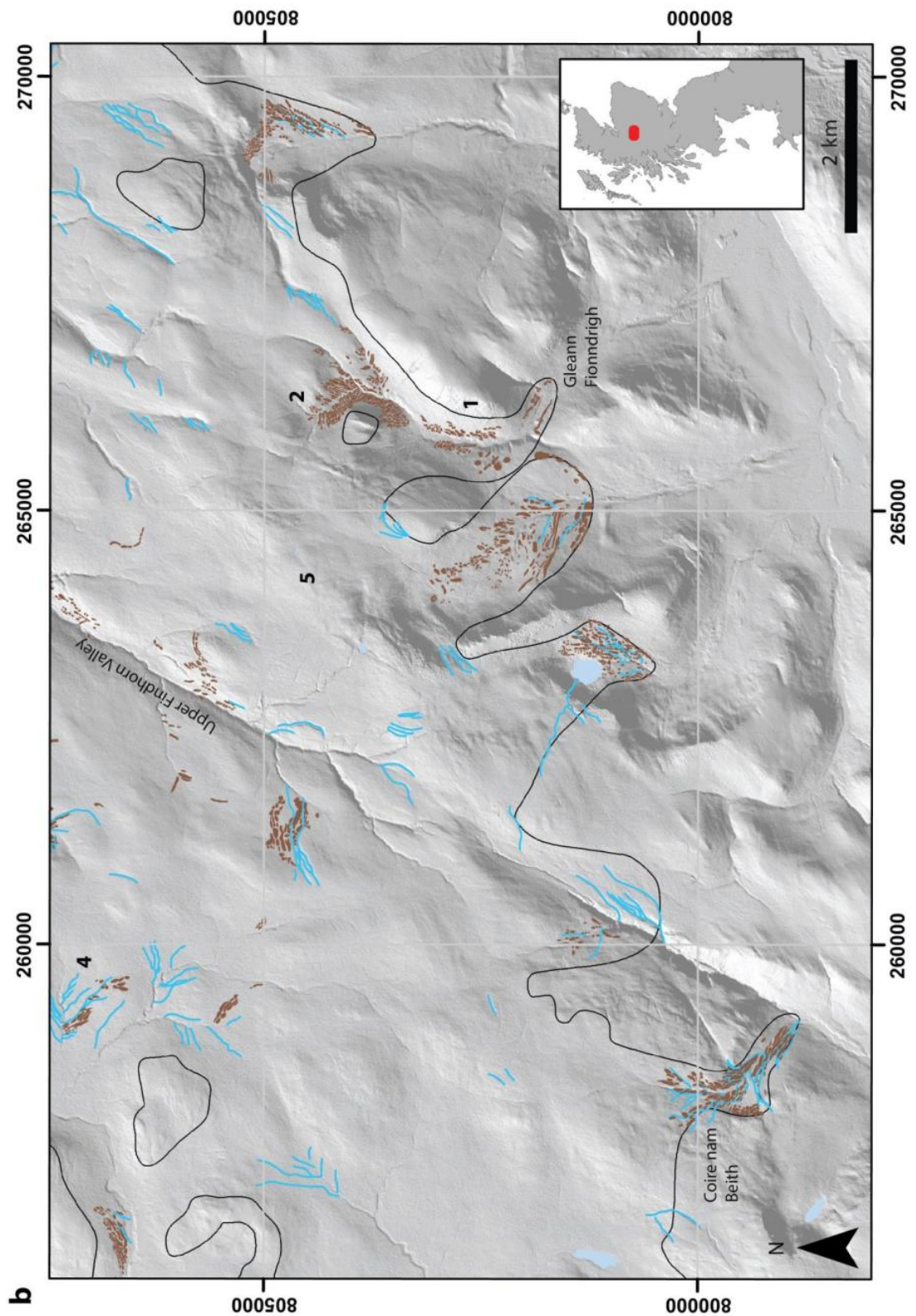
Figure 4.6 The plateau icefield landsystem. a) Conceptual model of the plateau icefield landsystem, characteristic of upland areas with gently undulating summits, on which ice accumulated. The plateau was drained by outlet glaciers in the surrounding valleys. 1. Recessional moraines in the valley, 2. Recessional moraines at the valley head which rise onto the plateau, 3. Ice-moulding and roches moutonées at the valley head, 4. Ice-marginal meltwater channels, 5. Periglacial blockfields and solifluction lobes. b) Example of the LLS plateau icefield landsystem in the Monadhliath Mountains (after Boston, 2012a). Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

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### 4.3.5 The ice cap landsystem

Unlike the aforementioned landsystems, many of the elements of the ice cap landsystem are discordant with the underlying topography, as detailed by Golledge (2007). Broad, smooth-topped moraines which are aligned transverse to ice flow can be found throughout much of the area (Figure 4.7), including on the high plateaux, and show modification of their surfaces with superimposed lineations (1 on Figure 4.7). Moraines are also found on many of the valley floors (2), sometimes misaligned with the surrounding topography, and thick accumulations of till are asymmetric and deposited on valley sides, particularly where ice flowed obliquely across (3) or up (4) the valleys. Evidence of ice-scouring is widespread; roches moutonnées (5) are common on the lower valley sides and a transition occurs to whalebacks at higher elevations (6), but ice-moulded bedrock is present even in the high cols (7). A transition from ice-moulded to frost-weathered debris (8) is apparent on the high summits, although Golledge (2006) argued that this transition occurs more gradually rather than a sharp trimline boundary. Periglacial features, such as frost-shattered regolith (9) and solifluction terraces, are present on the summits of these hills.

Golledge (2007) argued that many of the landforms which feature in the ice cap landsystem are inherited from the Late Devensian Ice Sheet and had been largely preserved beneath a minimally erosive LLS ice cap. For example, it was suggested that the broad moraines were deposited during a previous phase of glaciation, most likely the Late Devensian, before the lineations were superimposed onto them during the LLS (Golledge, 2006). Generally, till cover may be patchy throughout the area, but localised exposures show the preservation of thick till sequences, particularly where the ice flow was not controlled by the underlying topography, such as where it flowed against reverse slopes or obliquely across valleys (Golledge, 2007). Like the moraine evidence, this supports the presence of a LLS ice cap where flow direction was governed by the ice surface slope, rather than by the underlying relief. The evidence for ice-moulding over even the high cols again confirms that the areas covered by a considerable thickness of ice during the LLS would either have continued to modify ice-moulded surfaces that were abraded by the Late Devensian Ice Sheet, or would have at least protected them from modification by severe periglacial conditions during the LLS.

Exemplar geomorphology of the ice cap landsystem has been reported for the area between Rannoch Moor and Glen Falloch (Golledge, 2007). An asymmetric distribution of deposits is common in numerous valleys. For example, moraines are present only on the southern valley slopes and hills of Glen Lyon (Figure 4.7b), whilst the northern slopes host

only gullied drift (Golledge and Hubbard, 2005). Similarly, the moraines in Cononish Glen do not run symmetrically across the valley and those in neighbouring Glen Auchreoch show ice-contact slopes facing down-valley, which is impossible to reconcile with glacier flow being controlled by topography. A further clear example occurs at Lairig an Lochain, (Figure 4.7c) where the ice-marginal positions inferred from the moraine evidence are aligned almost parallel to the valley axis, perpendicular to what would be expected for a topographically constrained glacier. Transverse moraines overprinted by lineations are particularly well displayed on the upland area south of Ben Lui and Ben Oss (Figure 4.7d). Thick till accumulations, can be observed where the ice was forced to flow obliquely across valleys or upslope, as occurs in the cirque south of Loch Lyon's head. Streamlining of the high cols between the mountain summits is displayed in the upland area northwest of Loch Lyon, above Lairig nan Lochain and between Ben Lui and Ben Oss (Golledge, 2007), requiring a substantial thickness of ice overriding these sites. The sum of this evidence strongly supports an ice cap, rather than icefield landsystem for this area (Golledge and Hubbard, 2005; Golledge, 2007).
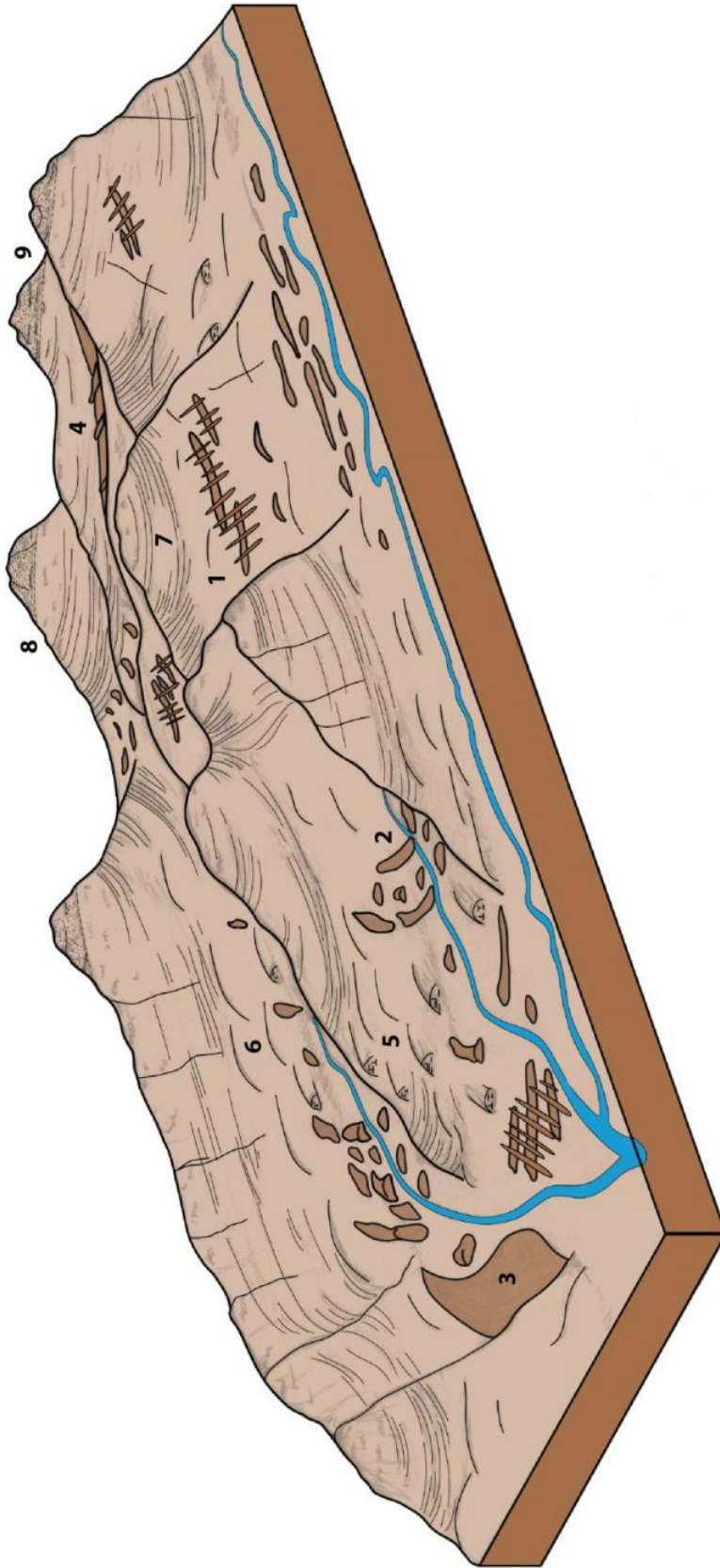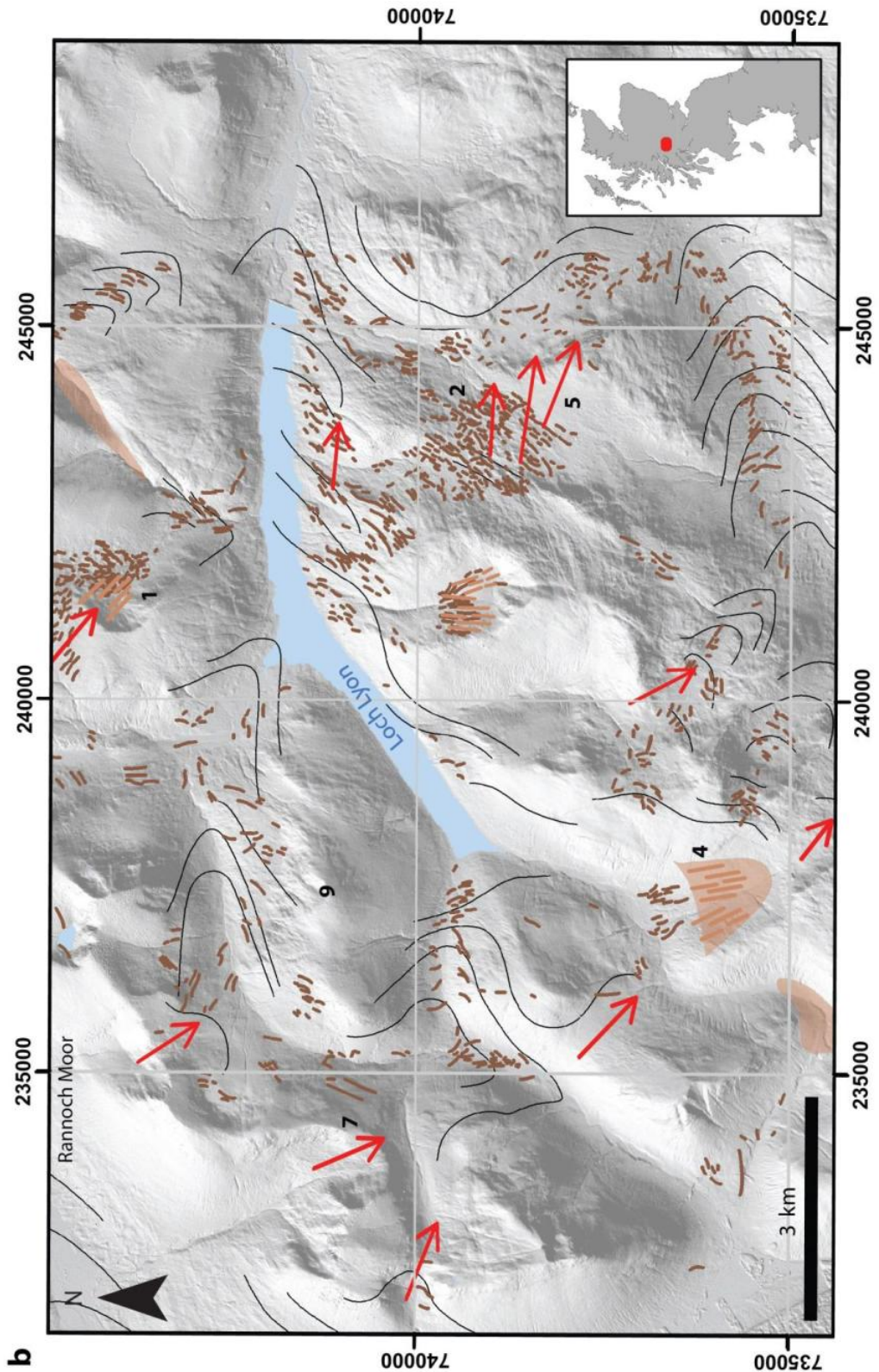
Figure 4.7 The ice cap landsystem. a) Conceptual model of the ice cap landsystem, found at the centre of the main LLS ice mass. 1. Pre-LLS moraines overridden by LLS-aged lineations, 2. Moraines in valleys aligned discordant to topography, 3 and 4. Thick accumulations of preserved till. 5. Roches moutonées, 6. Whalebacks, 7. Ice-moulded bedrock in the high cols, 8. Transition from ice-moulded bedrock to frost-weathered debris, 9. Solifluction features and frost-weathered regolith. b, c and d) Examples of the LLS ice cap landsystem in the vicinity of Rannoch Moor/Glen Lyon, Lairig and Lochain and Ben Lui-Ben Oss respectively (after Golledge, 2007). Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

**Moraines** · · · ·

**Lineations** ⫽⫽

**Thick till accumulations** ▬

**Interpreted ice margins** ))))

**Water bodies** ▬

**Ice flow direction indicators (streamlined cols, striae, roches moutonnées and erratic carry)** →
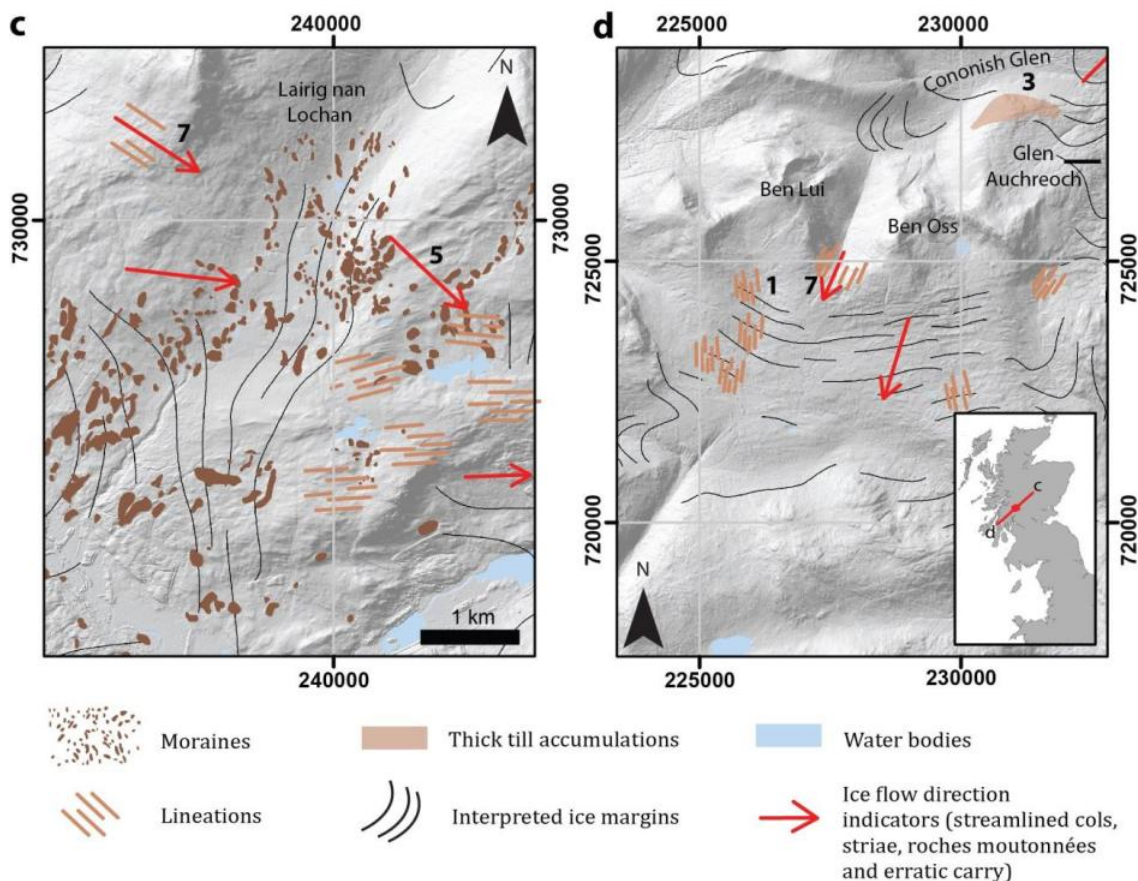
### 4.3.6 Spatial patterns in the distribution of landsystems

The glaciers which comprise the cirque/niche landsystem were highly topographically constrained, and thereby constituted discrete ice masses. This landsystem, (shown in red on Figure 4.8) is found primarily around the peripheries of other larger ice masses, at favourable topographic sites where conditions were not conducive for the development of valley glaciers. Suitable locations include topographic hollows or sites in the lee of escarpments, which were protected from insolation or nourished by additional windblown snow. For example, icefields on many of the Scottish islands, including the Outer Hebrides, Skye and Mull, are flanked by smaller cirque glaciers. This landsystem becomes progressively more common with distance away from the centre of glaciation in the West Highland Glacier Complex. Glaciation was exclusively of this type in the uplands of Galloway in Scotland, the Pennines in northern England, and both Snowdonia and the Brecon Beacons in Wales. Likewise, glaciation on the Orkney Isles, Scotland, comprised only two small cirque glaciers during the stadial.

The alpine icefield landsystem was widespread in Scotland during the LLS, and covers the greatest combined area (Figure 4.8). Clear examples of this landsystem are represented by the satellite icefields which formed in upland areas around the flanks of the West Highland Glacier Complex (Golledge, 2010). Substantial alpine style icefields occupied the upland

areas of the isles of Skye and Mull with networks of valley glaciers fed from accumulation areas in cirques and upper valleys (Benn et al., 1992; Ballantyne, 2002a). The steep nature of the underlying topography inhibited accumulation of ice on the summits themselves and ice drained away from prominent topographic barriers. Smaller alpine icefields were nourished on the isles of Lewis, Harris, Rùm and Arran, some of the outlets of which extended beyond the current coastline. On the mainland, an icefield formed on the Applecross Peninsula. Although previous studies have referred to the glaciation of plateau surfaces in this area during the LLS (Jones, 1998), reconstructions by some authors (Robinson, 1977, 1987; McCormack, 2011) indicated that the valley glaciers were connected over a col but that the majority of the plateau areas were exposed beyond the glacier limits. To the north of the West Highland Glacier Complex, a substantial alpine icefield developed in the mountains of Sutherland (Lukas and Bradwell, 2010), the valleys of which are dominated by extensive sequences of hummocky moraines (Benn and Lukas, 2006; Lukas and Benn, 2006). The same landsystem is present in the West Drumochter Hills.

The alpine icefield landsystem is also represented by the geomorphology of the northern sector of the West Highland Glacier Complex, between Loch na Sealga and the Great Glen (Figure 4.8a). The terminology used for the ice mass in this area has alternated between 'icefield' (Bennett and Boulton, 1993a; Ballantyne, 2012) and 'ice cap' (Bennett, 1991; Bennett and Boulton, 1993b). This is largely because the vertical extent of these glaciers is poorly constrained. The majority of field mapping to date has concentrated on the geomorphology in the valleys (Bennett, 1991; Bennett and Boulton, 1993a, b) and little to no evidence has been presented from which former ice thickness can be inferred. However, although the majority of the evidence was formed during deglaciation, the signature of retreat in these valleys contrasts starkly with that from the sector of the ice mass south of the Great Glen. The majority of valleys in the northern sector are dominated by sequences of recessional hummocky moraines formed during deglaciation, indicative of active retreat which continued until the late stages of deglaciation, with glaciers retreating back towards decay centres on the northern and northeastern faces of mountain ridges (Bennett and Boulton, 1993a). This retreat pattern indicates a strong topographic influence on the retreat directions of these glaciers, suggesting that an alpine icefield landsystem most accurately reflects the style of LLS glaciation in this area. Furthermore, numerical modelling by Golledge et al. (2008) produced an ice mass which closely matched the lateral extent of LLS glaciation, as indicated by empirical evidence, the northern sector of which appears to have contained more nunataks than the southern sector, again supporting an alpine icefield classification for the northern sector.
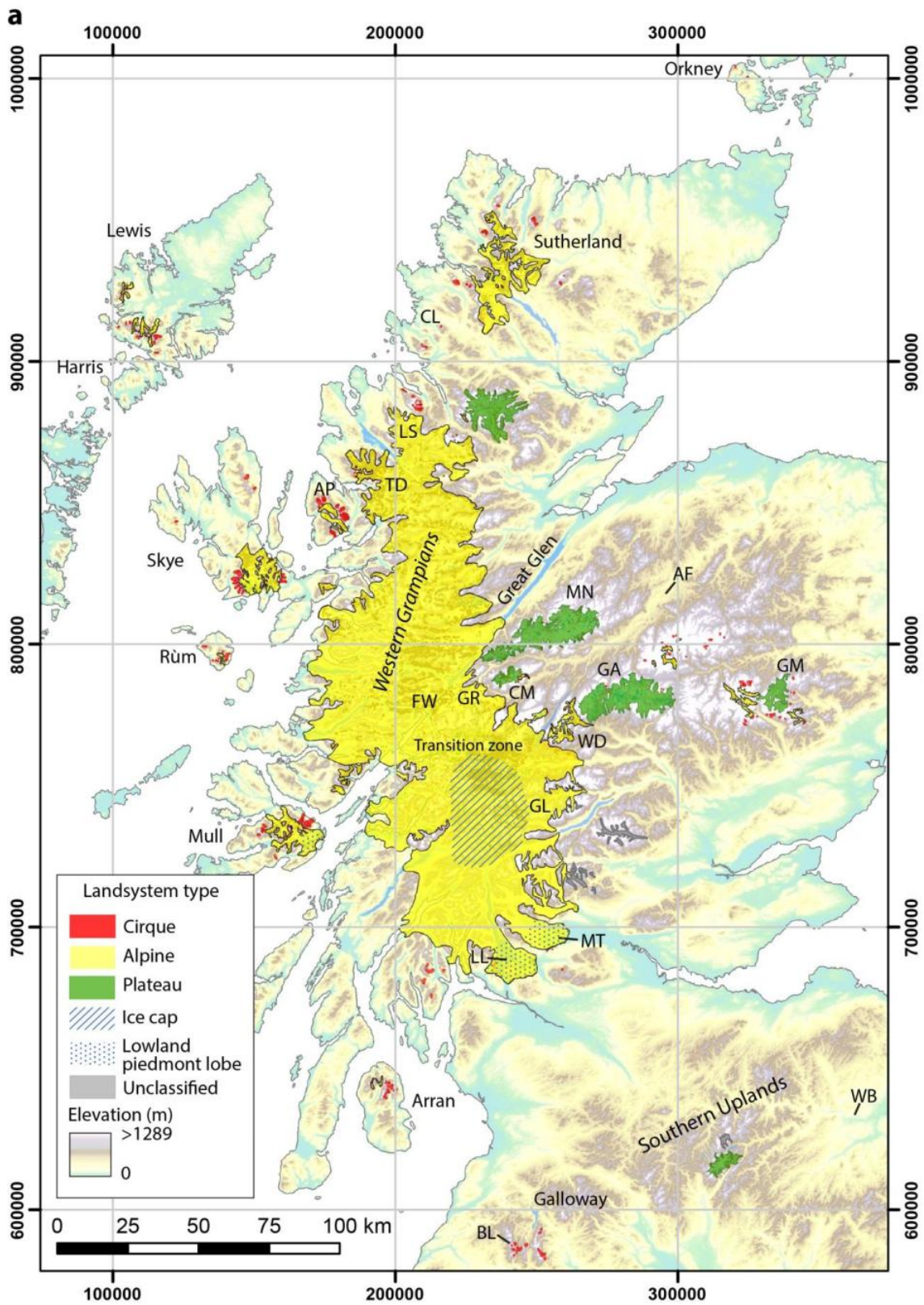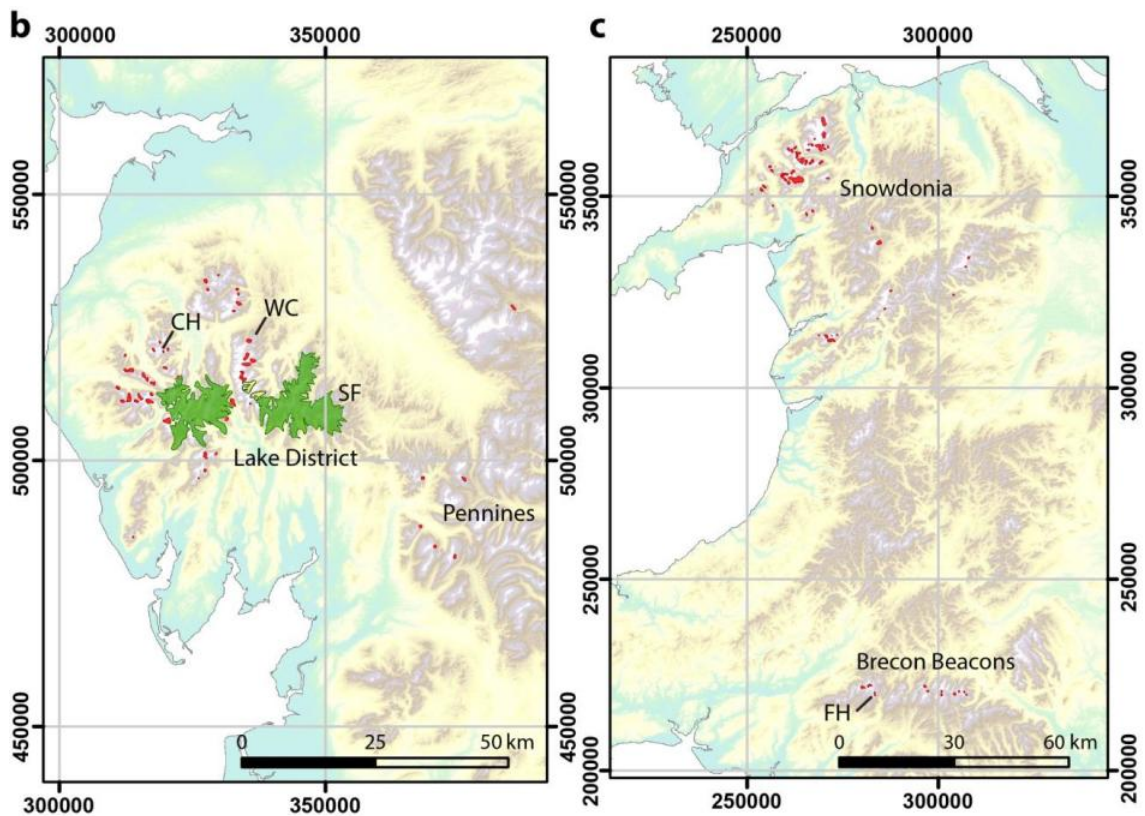
Figure 4.8 Spatial distribution of LLS landsystems in Britain. a) Scotland is dominated by the West Highland Glacier Complex, the majority of which comprised an alpine icefield landsystem. The hatched area represents the approximate extent of the ice cap landsystem. The limits of this landsystem are uncertain, particularly to the south for which geomorphological mapping is poor, although mapping of the termini of the Lomond and Menteith lobes accords with the lowland piedmont lobe landsystem. CL: Cul Mor, LS: Loch na Sealga, AP: Applecross, TD: Torridon, AF: Abernethy Forest, MN: Monadhliath, CM: Creag Meagaidh, GA: Gaick, GM: Glen Mark, GR: Glen Roy, FW: Fort William, WD: West Drumochter, GL: Glen Lyon, LL: Loch Lomond, MT: Menteith, BL: Balminnoch Loch, WB: Whitrig Bog. b) England hosts evidence of the plateau icefield landsystem nourished in the mountains of the Lake District, flanked by cirque glaciers. CH: Crag Hill, WC: Wolf Crags, SF: Shap Fells. c) The cirque glacier landsystem alone is represented by the LLS glacial geomorphology of Wales, where conditions were only just above the threshold required for glaciation. FH: Fan Hir. British coastline and present-day waterbodies reproduced from Ordnance Survey © Crown Copyright and Database Right 2015. Ordnance Survey (Digimap Licence). GB SRTM Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA.

Geomorphological evidence for the lowland piedmont landsystem is uncommon and has been mapped in only three locations, including the termini of the Lomond and Menteith lobes at the south of the West Highland Glacier Complex, and (possibly) the Spelve-Don lobe of the Mull icefield. In each of these cases, Gilbert-type glaciomarine deltas formed during the retreat of the Late Devensian Ice Sheet were subsequently overridden by outlet glaciers from LLS icefields. Consequently, these locations show evidence of glacitectonism, including thrust-block and composite moraine ridges, and features which may indicate the surging of these glaciers during the LLS.

To the east of the West Highland Glacier Complex, the plateau icefield landsystem is represented in the Beinn Dearg, Monadhliath, Creag Meagaidh and Gaick mountains (Sissons, 1974; Finlayson, 2006; Finlayson et al., 2011; Boston, 2012a). In each of these instances, valley outlet glaciers were fed by ice that had accumulated on the plateaux, rather than in cirque or valley head source areas. In the southeastern Grampians, a small plateau icefield was nourished on the plateau area above Glen Mark, although given the lack of landform evidence in this upland area, the presence of ice was largely inferred from the configuration of glaciers in the neighbouring valleys (Sissons and Grant, 1972). Plateau icefields were nourished in the Tweedsmuir Hills and the mountains of the English Lake District, the latter representing the most southerly location in the British Isles to have been extensively glaciated during the LLS. Evidence of a LLS icefield landsystem is present in the Cairngorm mountains, where Bennett and Glasser (1991) and Sissons (1979c) mapped geomorphological evidence indicative of an alpine icefield but which Standell (2014) has subsequently suggested may have hosted plateau ice.

Although it seems very likely that the northern sector of the West Highland Glacier Complex was characterised by the alpine icefield landsystem, south of the Great Glen, a different landsystem has been observed. Mapping around the vicinity of Glen Lyon showed numerous landsystem elements indicative of topographically discordant ice flow, such as recessional moraines aligned obliquely across valleys (Golledge, 2007). These features, in conjunction with streamlined bedrock in the high cols, striations, roches moutonées and erratics, require an ice mass of substantial thickness to have been situated over the area, the radial flow of which could override all but the highest mountain summits (Golledge, 2007). In light of this field evidence and the results of numerical modelling, Golledge and Hubbard (2005) and Golledge (2006) proposed that the upper limit of the LLS ice cap must have occurred at around 900 m.a.s.l. This contrasts with the topographically constrained mountain icefield proposed by Thorp (1981, 1984, 1991) for this area. However, as the ice retreated and downwasted, the importance of topography increased. This perhaps explains why the interpreted ice margin positions in the east are more discordant with the underlying topography than those to the west of Loch Lyon, as by the time the ice margin had reached this area, the underlying topography was exerting more control on the direction of ice flow.

A slightly different landform assemblage has been observed in the area between Glen Lyon and the Great Glen. Turner et al. (2014) observed that within the region of the Great Glen, hummocky moraine, described as irregular mounded terrain ranging between chaotic mounds and nested linear ridges, was found almost exclusively within the Rannoch Moor

basin. Both linear elements in the pattern of hummocky moraine mounds around Loch Ba (approximately 5km north of Figure 4.7b), mapped by Wilson (2005), and transverse moraines mapped just east of this by Turner et al. (2014), indicate ice flow into the Rannoch Basin from the west and southwest. This evidence accords with the reconstruction of the ice cap presented by Golledge (2007), the highest point of which occurs in the vicinity of Stob Ghabhar (approximately 9 km west-northwest of Figure 4.7b). In reality, this intermediate zone probably represented a transition between the ice cap and alpine icefield landsystem styles, becoming increasingly alpine as the ice retreated and downwasted to become more topographically constrained. It is possible that the complex pattern of hummocky moraine on Rannoch Moor results from the partial preservation of pre-existing moraines that were formed during an early readvance phase of the last British Ice Sheet and were preserved under a LLS ice cap, which flowed largely through internal deformation (Golledge, 2007).

It is important to note that classification of some areas within the limits of LLS glaciation was inhibited by a lack of detailed geomorphological mapping. This is the case for the possible icefield in the northern Tweedsmuir Hills. Likewise, although the southeastern sector of the West Highland Glacier Complex is classified as alpine icefield, this reflects the strong evidence of this landsystem signature elsewhere within the limits of this ice mass and may require revising if future geomorphological mapping indicates that a different landsystem is represented.

A rather striking overall asymmetry is apparent from the pattern and style of LLS glaciers across Britain, but particularly across Scotland (Figure 4.8). There is a clear difference between the predominantly alpine icefield style of glaciation in the western Grampians and Scottish islands, and the plateau icefields that are found further east. This contrast partly reflects the present-day landscape and the degree of topographic constraint on the LLS ice masses. There is a strong signature of glacial erosion in the western Grampians, where features typical of alpine glaciation, such as cirques, sharp ridges and deep incised basins, are common (Sutherland and Gordon, 1993). Conversely, the eastern Grampians are characterised by upland plateau surfaces that appear to have experienced only selective glacial erosion. The largest features in these areas comprise glacial breaches and troughs from pre-existing valleys generally aligned parallel to the direction of ice flow (Clayton, 1974). Similarly, further south the topography is both lower in elevation and relief, reflecting a transition from the resistant metamorphic lithologies that comprise the Highlands, to the sedimentary rocks which underlie the Midland Valley (Evans, 2003c). In this area, LLS glaciation was restricted to small plateau icefields and cirque glaciers.

Sutherland and Gordon (1993) argued that Scotland was likely glaciated during the first ice sheet glaciation of the Quaternary, approximately 2.4 Ma, and that the landscape at this time was probably comparable to that of the present day. It was suggested that the largest scale glacial features, such as troughs, must have been formed during successive episodes of ice sheet glaciation, whilst smaller, higher elevation features, such as cirques, could have been formed by more restricted ice cap glaciation (Sutherland, 1984; Sutherland and Gordon, 1993). This suggestion accords well with the distribution and style of LLS ice masses, explaining why in the western Grampians, where the high elevation terrain and snow-bearing westerly winds can initiate glaciation, glacial erosion occurs more frequently and produces more large scale features than in the east. The underlying topography influences the style of subsequent glaciations (as discussed below). In the eastern Grampians, although the upland plateaux provide a suitable source area for glaciers, the ice that covers these surfaces is often thin and cold-based, thus having a very minimal erosive impact and preserving the topography. Consequently, the LLS glaciers probably had a minimal impact on the long-term landscape evolution of Britain, modifying the pre-existing topography, rather than creating it.

### 4.3.7 Retreat styles

From the geomorphological evidence, it is clear that the retreat style of the LLS glaciers was not uniform across Britain and that variation exists within each of the landsystem models. Hummocky moraine is extensive within the limits of LLS glaciation, to the degree that it was previously considered to be diagnostic of glaciation during the period (Sissons, 1974). Sissons (1961) suggested that these features resulted from widespread stagnation of ice during the phase of rapid warming at the end of the LLS. However, more recent investigations of these landforms from geomorphological mapping and sedimentology has argued that many hummocks are actually recessional moraines (Bennett, 1991; Bennett and Boulton, 1993a, b; Lukas and Benn, 2006), which formed at the margins of actively retreating glaciers by a combination of ice push and subaerial deposition at the glacier terminus (Benn and Lukas, 2006). Since each ridge is inferred to have been formed during a minor readvance or stillstand during overall retreat, the position of these ridges can thus be used to reconstruct a series of palaeo ice-front positions, allowing the pattern of retreat to be assessed (Lukas and Benn, 2006). Using this framework, three principal retreat styles have been identified from the LLS geomorphological evidence (Figure 4.9). These are termed 'active retreat', 'two-phase retreat' and 'uninterrupted retreat'.
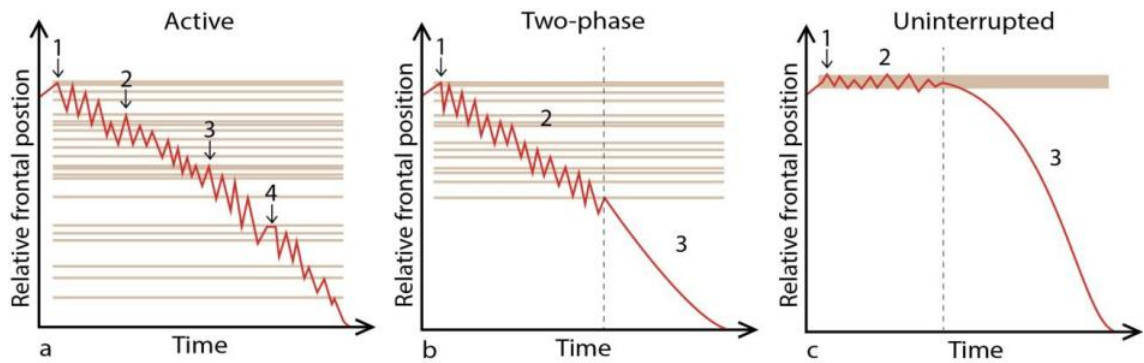
Figure 4.9 Conceptual retreat styles of LLS glaciers. Brown horizontal bars indicate the formation of a recessional moraine during a minor readvance or stillstand. a) Active oscillatory retreat throughout general deglaciation. 1. Maximum extent, 2. Readvance beyond the previous retreat position, forming an overridden moraine, 3. Glacier readvance to the same position during multiple readvance phases, perhaps producing a larger, complex push moraine ridge, 4. Period of stability and thus localised stagnation at retreat position, b) Two-phase retreat. 1. Maximum extent, 2. Phase of active retreat (left of the dashed line), 3. Phase of retreat uninterrupted by stillstands or readvances with a consequent lack of recessional moraines formed. c) Uninterrupted retreat throughout deglaciation. 1. Maximum extent, 2. The glacier oscillates slightly, reaching same margin position during several minor readvances, perhaps facilitating formation of a larger terminal moraine than under scenarios (a) and (b), 3. Phase of uninterrupted retreat (right of the dashed line) from the terminal position and a consequent absence of recessional moraines.

### 4.3.7.1 Active retreat

Active retreat is characterised by valleys in which a continuous (or near-continuous) sequence of recessional moraines stretches from the glacier terminus to the former source area (Figures 4.9a and 4.10). The distribution of these moraines indicates that retreat of the glaciers which formed them was interrupted by stillstands and minor readvances throughout, with the glaciers responding (with minor readvances or retreats) to changes in the climate until very shortly before they disappeared completely. In some cases, such as the icefield in Sutherland, northwestern Scotland, examination of the sediments comprising the recessional moraines indicated that some of these readvances were substantial enough that the glacier readvanced back to, or over, the previous readvance position (Benn and Lukas, 2006) (Figure 4.9a, 2 and 3). Where the glacier stabilised at a particular position, localised stagnation may have occurred (Figure 4.9a, 4). In the case of the Sutherland icefield, glacier size appears to have exerted some control over the frequency of moraine formation, with Lukas and Benn (2006) calculating that the smaller glaciers formed moraines on average every 7-23 years whilst the larger ones formed them every 3-11 years. This trend is perhaps counterintuitive as changes in climate are often muted by larger glaciers with slower response times (MacLeod et al., 2011). This landform signature implies that these glaciers were relatively 'clean' and did not support thick mantles of supraglacial debris, which would inhibit rapid glacier response to climate change (Benn and Lukas, 2006).

Applying the same methodology to the recessional moraines formed by LLS glaciers in other localities in Britain has the potential to aid in quantitatively assessing the dynamics of LLS glacial retreat across Britain. However, this analysis requires high resolution mapping and is not presently possible for all LLS ice masses. Of the three principal retreat styles discussed here, active retreat appears to have been the most widespread in Britain during the LLS, at least in the areas for which detailed geomorphological mapping exists. This includes the northern half of the West Highland Glacier Complex (Bennett, 1991; Bennett and Boulton, 1993a), the Sutherland icefield (Lukas and Benn, 2006), the Outer Hebrides (Ballantyne, 2006; Ballantyne, 2007a), Mull (Ballantyne, 2002a) and many of the valleys in the Lake District, where recessional moraine sequences continue up-valley and onto the plateau areas (McDougall, 1998, 2001).
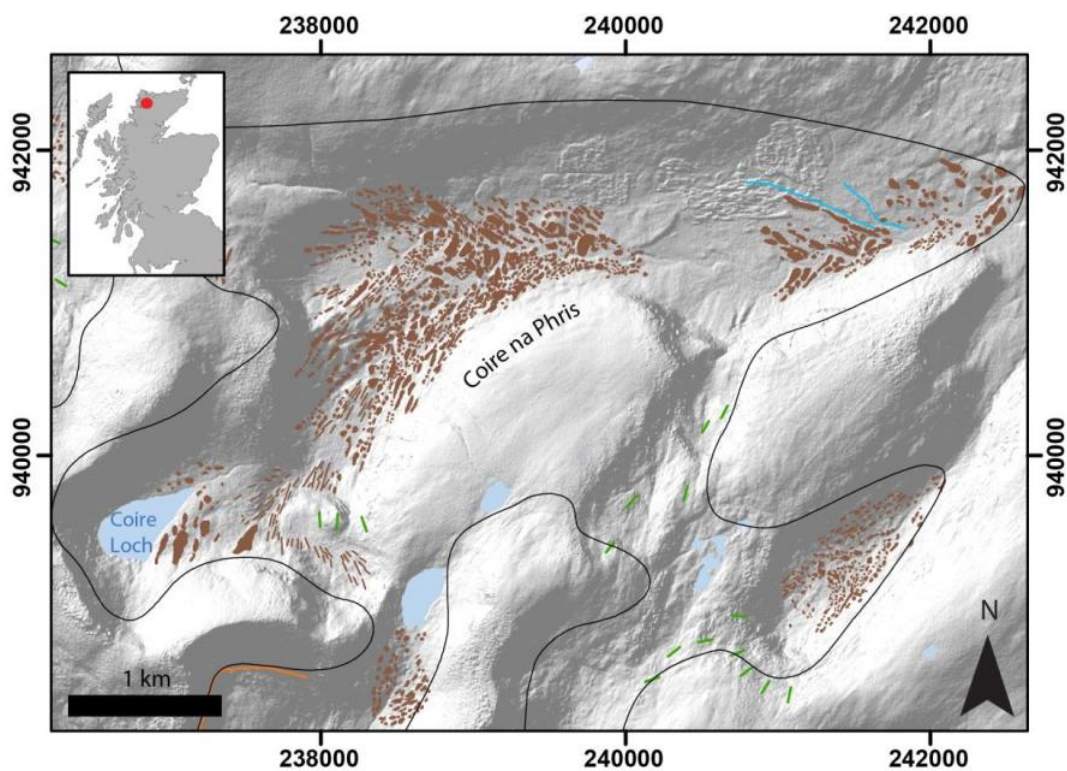


Figure 4.10 An example of the geomorphological signature of active retreat until the late stages of deglaciation (after Lukas and Lukas, 2006a, b). Recessional moraines form a near-continuous sequence from the terminus of the Coire na Phris glacier, Sutherland, Scotland, to its source area at Coire Loch. This glacier was extremely active, readvancing and retreating rapidly in response to fluctuations in climate during the stadial. Gaps within the sequence may indicate brief periods of uninterrupted retreat, a change to the sediment regime that prevented moraine formation, erosion of moraines following deglaciation or gaps in the mapping. Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

*4.3.7.2 Two-phase retreat*

A rather different retreat pattern has been inferred (e.g. Benn, 1990, 1992; Benn et al., 1992), primarily on the Isle of Skye, Scotland, which is termed here 'two-phase retreat' (Figure 4.9b). In this case, recessional moraines are largely restricted to the outermost terminal zone of each glacier, although the number of ice front positions represented varies between catchments (Benn, 1990). A transition then occurs to terrain indicative of local *in situ* ice stagnation, such as chaotically arranged morainic mounds or thick sheets of irregular drift (Benn et al., 1992). Alternatively, the geomorphology up-valley of the recessional moraines may be suggestive of rapid, uninterrupted retreat, for example where the orientation of flutings or erratic dispersal relates to the maximum position of the glacier terminus, indicating that the course of glacier retreat was uninterrupted by subsequent readvances (Benn, 1990; Benn et al., 1992). It seems likely that the onset of uninterrupted retreat occurred when the ELAs increased during deglaciation, forcing large areas of some icefields (e.g. Skye) to become ablation zones. The reoccurrence of recessional moraines after the onset of uninterrupted retreat occurs in only two locations on Skye, at Harta Corrie and Coire Dubh. Both of these are high cirque sites (Benn, 1990; Benn et al., 1992), the elevation of which was probably sufficient that, once the icefield had melted, small glaciers in these cirques re-established equilibrium with climate conditions. This would have produced a second phase of active retreat until the ELAs finally rose above even the highest cirques and deglaciation was complete.

Although this two-phase pattern of deglaciation is widespread across the Isle of Skye, it was long considered to have been absent from mainland Scotland (Bennett and Boulton, 1993a). However, recent work from Torridon has reconstructed a two-phase deglaciation pattern. In this locality, an abundance of chaotic morainic mounds have been mapped in the area which fed much of this sector of the LLS icefield (McCormack, 2011) (Figure 4.11). McCormack (2011) suggested that the transition to stagnation terrain in the form of chaotic debris mounds could possibly be attributed to a high supraglacial debris load which insulated the ice, causing it to become disconnected from the accumulation areas, and thereby inactive. Likewise, although recessional moraines were observed in the upper valleys of some of the glaciers in the West Drumochter Hills, these are largely absent above 550-630 m OD, suggesting that the final stages of retreat were also uninterrupted in this locality.
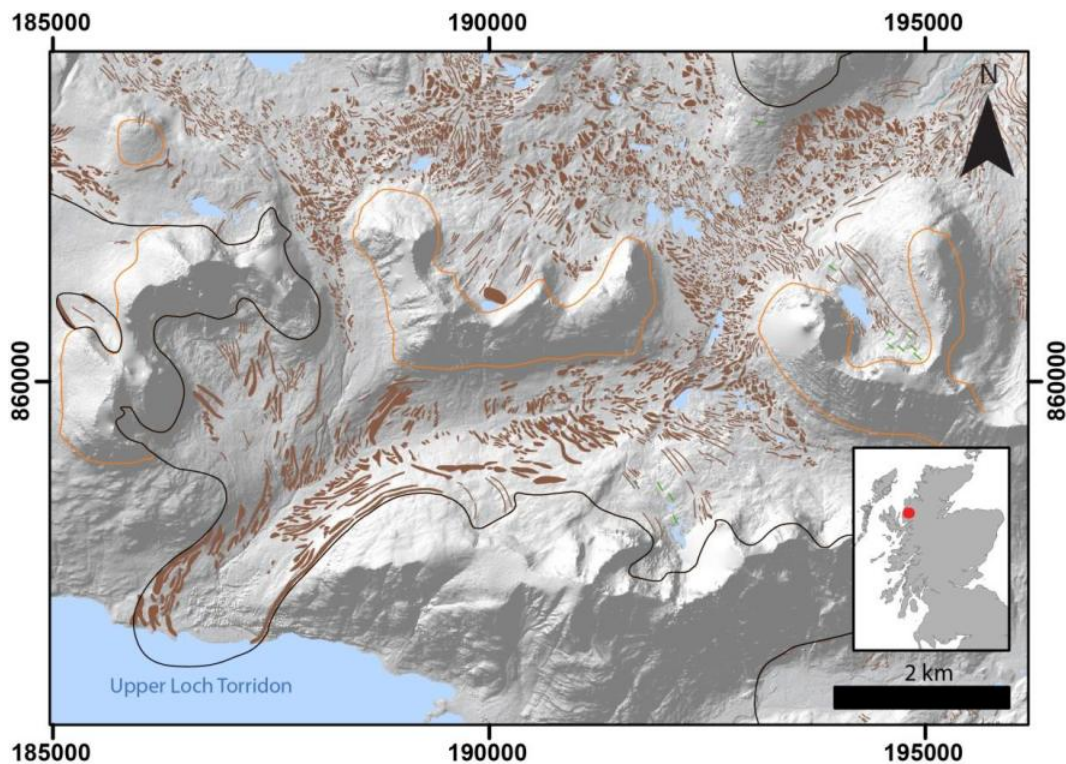
Figure 4.11 An example of the geomorphological signature of two-phase retreat (after McCormack, 2011). The extent of the LLS glacier which descended from the Torridon icefield into Upper Loch Torridon is marked in its lower reaches by a sequence of prominent recessional moraines. This contrasts with the chaotic mounds, indicative of ice stagnation, which are found in its former source area. Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

### 4.3.7.3 Uninterrupted retreat

The third style of retreat, 'uninterrupted retreat', is rather uncommon and occurred in areas where recession from the maximum ice limit appears to have been largely uninterrupted by stillstands or readvances, as indicated by an absence of recessional moraines. In some cases, the associated geomorphological record takes the form of a large terminal moraine ridge (as described by Sissons, 1980a), formed by the glacier oscillating to and from the same margin position (Figure 4.9c, 2), inside of which recessional ridges are absent (Figure 4.12). Examples of such features include the terminal moraines at Bowscale and Scales tarns (Figure 4.12a), and on the northeastern flank of Crag Hill, Lake District, England (Sissons, 1980a). Similar features are found in Wales at Llyn y Fach (Figure 4.12b) in the Brecon Beacons, and at Melynllyn (approximately 2 km north of Figure 4.3b), Snowdonia (Bendle and Glasser, 2012). These features are also present at both Balminnoch Loch (Figure 4.12c) (Cornish, 1981) and Cul Mor Assynt (Sissons, 1977a) in Scotland. Such features indicate that the glacier remained at its maximum extent for a prolonged period, allowing sufficient debris to accumulate and form a large moraine ridge at the terminus. The absence of moraines inside this outer limit indicates that, once deglaciation commenced, it was uninterrupted by stillstands or readvances. Between the

two-phase and uninterrupted retreat styles there are numerous examples of large outer moraines inside of which one or two recessional ridges are preserved (for instance Wolf Crag in the English Lake District (Sissons, 1980a)).
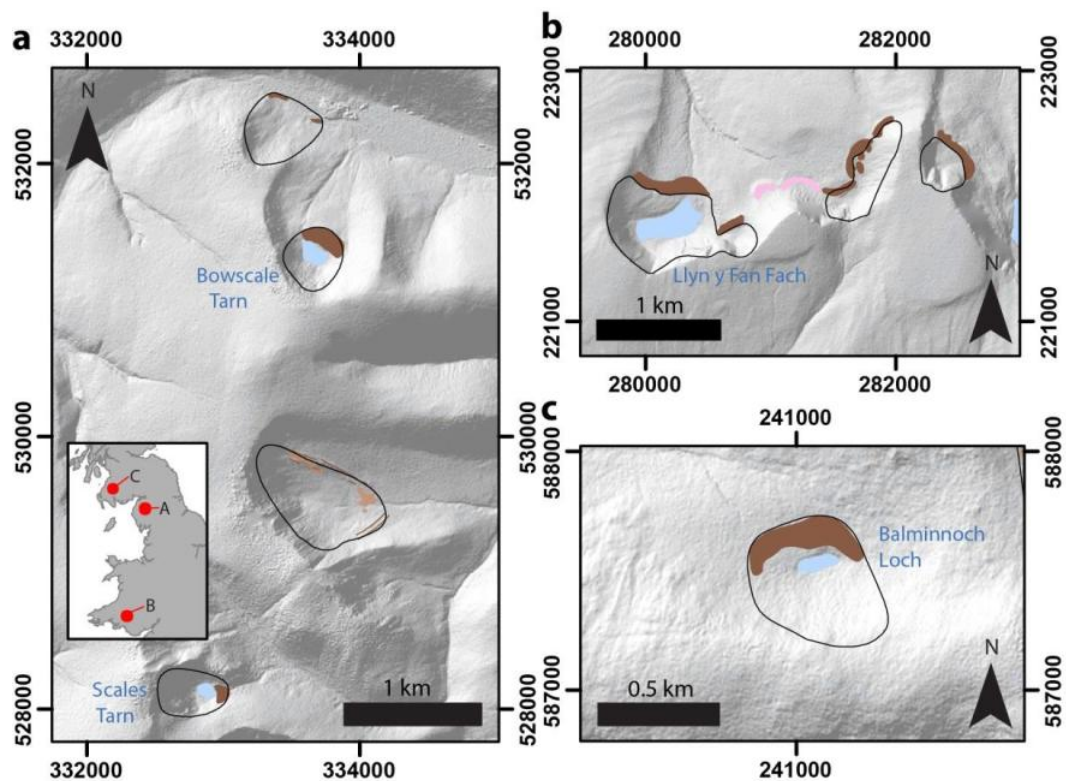


Figure 4.12 Examples of the geomorphological signature of uninterrupted retreat at a) Bowscale and Scales tarns, Lake District, England (after Clark and Wilson, 2001; Sissons, 1980a), b) Llyn y Fan Fach, Brecon Beacons, Wales (after Shakesby, 2007) and c) Balminnoch Loch, Galloway, Scotland (after Cornish, 1981). In each instance, the maximum extent of the LLS glacier is marked by a single, large terminal moraine, inside of which no recessional moraines are present. Hill-shaded images derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre. Key as in Figure 4.1.

An absence of recessional moraines does not necessarily always relate to uninterrupted retreat and alternative explanations for localised gaps in the depositional landform record are discussed in Section 4.4.1. These include: a cessation of moraine deposition, resulting from a change in debris availability; changes to bed topography; and erosion of moraines subsequent to deglaciation. The reasons for the tendency of apparently uninterrupted retreat to have occurred only in the relatively smaller palaeo-glaciers outlined above are unclear. Potentially, the evidence could indicate that these small ice masses were indeed incapable of maintaining quasi-stable fronts after the peak of the LLS maximum (uninterrupted retreat). Alternatively, they may have failed to deliver sufficient debris to their margins to form moraines and hence the lack of recessional moraines is a reflection of restricted basin sizes.

Since the three retreat styles have been inferred from a database comprising previously mapped features, it is possible that an apparent lack of moraines may reflect a gap in the mapping, rather than in the moraine record. This may be the case particularly for the uninterrupted retreat style as, whilst both the active and two-phase retreat styles are represented in high resolution mapping (e.g. Benn, 1990; Lukas and Lukas, 2006a, b), the features which indicate uninterrupted retreat have generally been mapped at a coarser scale where some of the geomorphology may have been simplified. Additionally, the signature of apparently uninterrupted retreat is present at only a small number of sites, for some of which the nature of LLS moraine evidence has been challenged (e.g. Jansson and Glasser, 2008). At many of the sites associated here with uninterrupted retreat, the terminal moraine impounds a tarn or lake (Figure 4.12), which may itself have eroded or obscured inner recessional moraines indicative of active retreat. Given these uncertainties, the palaeoglaciological and palaeoclimatic implications of apparent uninterrupted retreat requires further study.

## 4.4 Discussion: controls on LLS extent and retreat style: the relative importance of topography vs. climate

### 4.4.1 Local controls on moraine formation and preservation

An assumption in much of the mapping that has previously been conducted of LLS glacial geomorphology is that the distribution of features accurately represents the ice-marginal position throughout deglaciation, allowing inferences about the palaeoclimate at the time of formation to be made (e.g. Lukas and Benn, 2006). However, variations occur in the preservation of geomorphological evidence between (even adjacent) valleys of the same landsystem type. For example, Boston et al. (2015) observed that, of the 53 reconstructed outlet glaciers from the Monadhliath icefield, morainic evidence was absent from 23 valleys whilst 17 had clear sequences of recessional moraines. Likewise, moraines are largely absent from the banks of lochs Nevis, Morar, Ailort, Shiel, Sunart and Linnhe which were occupied by the westward draining lobes of the main West Highland Glacier Complex (Bennett and Boulton, 1993a), whereas the eastwards flowing outlet glaciers are well-constrained by evidence. In the English Lake District, numerous valleys that were glaciated during the LLS host sequences of recessional moraines and yet some glaciers, such as the Langdale glacier (McDougall, 2001) or several of those in the Shap Fells (McDougall, 2013), formed little if any morainic evidence. Similarly, although the outlet glaciers of the Mull icefield are marked by sequences of recessional moraines, these are almost entirely absent from the terrain occupied by the Spelve-Don lobe (Ballantyne, 2002a).

There are several possible reasons for these localised gaps in the depositional landform record, relating either to the tendency for glaciers to deposit no moraines, or the poor preservation potential of moraines during, and subsequent to, deglaciation. Boston et al. (2015) proposed a number of reasons to explain this phenomenon in the Monadhliath Mountains. In terms of factors influencing formation, these included: low debris turnover, low debris availability, uninterrupted retreat, and differences in bed topography. A low turnover of debris would certainly inhibit moraine formation and accords well with the notion of cold-based ice forming a component of plateau icefields, such as those reconstructed in the Monadhliath Mountains (Boston et al., 2015) and, potentially, the eastern hills of the Lake District. However, this does not account for the lack of moraines associated with the western outlets of the West Highland Glacier Complex. Furthermore, it is difficult to reconcile the differences between valleys with differential sediment availability. However, this factor offers a potential alternative explanation to the two-phase retreat pattern exhibited in some locations. It seems highly probable that LLS glaciers would have readvanced over large volumes of paraglacial rockslope debris, produced subsequent to the earlier ice sheet retreat, and that this debris contributed to the thick drift and sequences of recessional hummocky moraine commonly found inside the limits of LLS glaciation (Ballantyne, 2002b). It is possible that, in some areas, this supply of debris was exhausted mid-way through deglaciation and that, although active retreat may have continued, there was insufficient debris for moraine formation, giving the apparent geomorphological signature of a second phase of uninterrupted retreat.

It is also possible that some valley glaciers experienced uninterrupted retreat while their neighbours experienced active retreat, probably due to localised variations in glacier catchment topography, elevation, aspect, and glacier hypsometry. Furthermore, bed topography has been shown to play an important role in both the formation, and preservation, of moraines (Barr and Lovell, 2014 and references therein). Indeed, topography probably played a significant role in the lack of moraines along the sides of the western lochs, that were occupied by marine-terminating, calving glaciers, which likely flowed rapidly and stripped deposits from the valley sides (Bennett, 1991). It is probable that the maximum extents of these glaciers were governed by the topography of their basins, with the ice-margins stabilising and forming moraines only at pinning points where the grounding line encountered topographic highs or narrowings of the valley (Greene, 1992). Once the glacier terminus was released from such pinning points, deglaciation may have been uninterrupted until the next pinning point was encountered. It is also possible that accumulation of sufficient debris to form moraines was not possible

on the steep slopes which flank these valleys and that, if moraines had formed, these would rapidly have been reworked by hillslope processes.

An absence of moraines may also reflect efficient removal of glacigenic debris at ice margins where glacial and proglacial fluvial systems are well-coupled. In such instances, the majority of sediment is removed rapidly from the terminus by meltwater streams, the migration of which erodes any moraines which might begin to form, and is usually deposited in extensive outwash plains and sandar (Benn et al. 2003). Such features were not commonly formed by the LLS glaciers in Britain, indicating that outwash from these glaciers was insufficient to remove debris from the glacier terminus, accounting for the extensive moraine sequences found within the LLS glacial limits.

### 4.4.2 Topographic influence on LLS landsystems

Topography appears to have been an important control in the landsystem and retreat styles of the LLS, at both regional and local scales. Due to their relatively small extent, compared to that of the Late Devensian British-Irish Ice Sheet, the LLS glaciers would have had minimal erosive impact on the underlying topography, largely inheriting the recently deglaciated landscape that then became glaciated (Evans, 2015a). This underlying topography thus exerted a strong control on the type of LLS landsystem represented by the geomorphology. For instance, where the topography is dominated by steep slopes and narrow mountain ridges, the gradient of high elevation surfaces was too great to allow sufficient snow accumulation to nourish glaciers. Ice was consequently restricted to the lower gradient slopes that form the backwalls of cirques and the valley heads, producing glaciers that were confined to this lower elevation topography. These glaciers were nourished by snowfall delivered directly onto the glacier surface, or blown onto the glacier from neighbouring ridges, or via avalanches. Conversely, gently undulating plateau areas provided surfaces that could act as accumulation areas to feed valley glaciers in the adjacent valleys (Rea et al., 1998; Rea and Evans, 2003). Indeed, Manley (1959) identified that the smaller the area of the plateau, the higher its summit must rise above the regional firn line in order for glaciers to form on the surface.

Topography is also likely to have played a significant role in the earliest, and latest, stages of LLS glaciation. Time-slices from numerical modelling of the LLS glaciers have shown that ice began to accumulate in the Rannoch Moor area from the very beginning of the stadial (Hubbard, 1999; Golledge et al., 2008) (Figure 4.13). It has been proposed by Golledge (2007, 2008) that the high summits of the mountains to the west of the moor would have acted like a snow-fence (Andrews et al., 1970; Hulton, 1997), their high relief facilitating accumulation of large volumes of snow and enabling glacier formation. During

the early stages of the LLS, the underlying relief of this area would have been important in determining the declining influence of topographic control as the icefield thickened from a more alpine icefield style of glaciation to an ice cap that overran all but the highest mountain summits.

Topographic influence was particularly important to the cirque glacier landsystem, which dominated in locations peripheral to the major LLS ice masses, where conditions were only just above the threshold required for glaciation. The volume of debris that forms LLS moraines within each cirque is generally small, indicating that the amount of headwall retreat during the stadial was low (Evans, 2015a). Thus, it seems highly probable that the LLS glaciers modified, rather than created, the cirques and that LLS glaciation of these sites was dependant on the pre-existence of cirques prior to glaciation. North and northeasterly facing cirques were particularly favourable for LLS glacier development, offering the maximum protection from insolation, especially where their backwalls cut into upland plateau areas or escarpments from which accumulating snow could be blown onto the glacier surface (Mitchell, 1996). For example, at their maximum extent, the majority of LLS cirque glaciers faced between northwest and east. In the English Lake District, of a total 158 cirques, at least 82% were occupied by ice during the Lateglacial, most probably during the LLS (Evans, 2015a). Those which were not occupied by ice during this period, generally were at low elevations or had southerly aspects (Evans, 2015a). Likewise, in Snowdonia, all but 4 of the 38 reconstructed LLS glaciers had an aspect of northwest to east (Bendle and Glasser, 2012), in the Brecon Beacons this applied to 12 of 14 glaciers, and in the Galloway Hills, 9 of 11 glaciers fell into this category (Cornish, 1981).

Figure 4.13 Time-slices from the optimum run of numerical modelling (Golledge et al., 2008), where the modelled extent of ice most closely matched with the empirical limits. Ice readvances early in the stadial and, in many locations, reaches its maximum extents between 12.6 and 12.4 ka. Increased aridity from the middle of the stadial causes net recession, with the majority of ice having retreated before the onset of rapid warming at the transition into the Holocene at approximately 11.7 ka. Reprinted from Quaternary Science Reviews, Vol. 27, Golledge, N.R., Hubbard, A. and Sugden, D.E., High-resolution numerical simulation of Younger Dryas glaciation in Scotland, 888-904. Copyright (2008) with permission from Elsevier.

### 4.4.3 Palaeoclimatic influence on LLS landsystems

Notwithstanding the discussion above, the pattern of glacier extent and, therefore, landsystem distribution, was not solely determined by topography. If this had been the case, it might be expected that the same landsystem would be found in locations where the underlying landscape was at a similar elevation and had similar topography (i.e. alpine mountain ridges or undulating upland plateaux). Figure 4.8 shows the distribution of LLS glaciers against the underlying topography, and it is clear that although all the LLS glaciers were sourced in upland areas, numerous locations exist where terrain was above the ELA of the nearest LLS glaciers, yet remained unglaciated. This is particularly clear in the central Scottish Highlands where a substantial area of land lies above 1000 m OD (Standell, 2014), but is also the case in the English Pennines and much of central Wales.

Ballantyne (2007a) observed a strong linear northward decline in the ELAs of reconstructed LLS glaciers and attributed this to a northwards decline in ablation season temperatures, although he noted that some of the estimated ELAs had been calculated from glacial reconstructions that have since been revised. The warmer ablation season temperatures in southern Britain partly explain why LLS glaciation was limited to cirque and niche glaciers, despite the presence of what were favourable topographic conditions for glaciation, particularly in Snowdonia. However, a decrease in temperatures driven by latitudinal differences does not account for the presence of various different types of landsystem at the same latitude. From Figure 4.8, it is apparent that a west to east transect across Scotland, running through Fort William, intersects the following landsystems: the probable zone of transition between the ice cap centred on Rannoch Moor and the more alpine style icefield north of the Great Glen; the alpine icefield in the West Drumochter Hills; the plateau icefield on the Gaick; a high elevation, but ice-free area; and the collection of cirque glaciers and small alpine and plateau icefields in the southeastern Grampians.

The existence of a precipitation gradient across Scotland during the LLS has long been recognised (Sissons, 1979a, 1980b). Golledge et al. (2008) found that numerical modelling could best replicate the empirical limits of the LLS glaciers at their maximum positions by imposing a 60% south-north and an 80% west to east precipitation reduction to the north and east of Rannoch Moor. A further reduction in precipitation from 12.5-12.0 ka BP was imposed, representing a more arid climate during this part of the stadial. Furthermore, as the LLS glaciers expanded to their maximum positions, it is likely that the West Highland Glacier Complex itself exerted increasing influence on the LLS climate, cooling air masses as they passed over the icefield and prompting precipitation (Benn and Ballantyne, 2005). Consequently, the climate of the central Highlands was more arid, effectively starving the glaciers in this area of precipitation and limiting their size, perhaps explaining why ice masses on the Monadhliath, Gaick and Glen Mark plateau areas did not evolve into topographically unconfined ice caps. The northwards reduction in precipitation probably contributed to the substantial ice thickness, and therefore ice cap landsystem, of the West Highland Glacier Complex in the vicinity of Rannoch Moor, and its decrease in thickness north of the Great Glen, despite cooler temperatures and similar underlying topography.

Previous studies (e.g. Ballantyne, 2002a; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010) have estimated palaeo-precipitation volumes using the method proposed by Ohmura et al. (1992), which calculates precipitation from the 3-month mean summer temperature at the ELA. Using this method, LLS mean annual precipitation on the Isle of

Mull was estimated at 23% higher than at present (Ballantyne, 2002a), and at between the present-day value and 25% higher on the Isle of Harris (Ballantyne, 2007a). In contrast, the precipitation volumes calculated for inland sites, such as the West Drumochter Hills (Benn and Ballantyne, 2005) and Creag Meagaidh (Finlayson, 2006), were statistically indistinguishable from present-day values, whilst those using the Ohmura et al. (1992) approach for the Monadhliath were slightly less than modern day totals. Calculations using this method therefore support the notion of a steep precipitation gradient from west to east across Scotland.

Golledge et al. (2010) have also argued that the LLS climate was characterised by increased seasonality resulting from the formation of more extensive sea ice. LLS summers were approximately 10°C cooler than present-day conditions, but winters may have been up to 30°C colder than present (Golledge, 2008). Under such conditions, the ablation season each year would have been relatively short in duration and the glaciers would have required a smaller volume of precipitation to remain in climatic equilibrium at their maximum extent. Numerical modelling that incorporates these much greater temperature ranges has indicated that both temperature and precipitation were much lower during the LLS in Scotland than was previously calculated from empirical based glacier reconstructions (Golledge et al., 2010). These findings have been used to derive a new temperature-precipitation equation with an added seasonality constant (Golledge et al., 2010). Application of this new equation to the Skye icefield, gave values which range from 60-80% of present-day values if summer rainfall dominated precipitation, to 45-60% for a neutral precipitation regime, to as little as 35-45% of present-day mean annual precipitation if winter snowfall dominated (Ballantyne et al., 2016). Likewise, Boston et al. (2015) found that the Golledge et al. (2010) equation gave mean annual precipitation values for the Monadhliath that were lower than present-day values, even when a summer precipitation bias was applied. These results concur with regional proxy records indicative of cold and dry winters during the LLS (as detailed in Golledge et al., 2010). In light of these results, recalculation of precipitation and temperature values from other LLS glacier reconstructions will be necessary before regional assessment of the patterns in LLS palaeoclimate across Britain can be made.

The nature of the LLS palaeoclimate has important implications for the dynamics and retreat patterns of these glaciers. The widespread distribution of recessional hummocky moraine inside the limits of the LLS glaciers throughout much of the northwest Scottish Highlands, indicates that these glaciers retreated in an oscillatory fashion, responding to fluctuations in climate, until the later stages of deglaciation. This behaviour was assumed

to have been driven by heavy snowfall, with mean annual precipitation totals up to 26% higher than modern day levels, leading to high mass turnover and the ability to respond rapidly to climate variations and hence the production of dense sequences of recessional moraines (Benn and Lukas, 2006).

The LLS palaeoclimate was considered to be a principal control on the two-phase deglaciation pattern observed on the Isle of Skye (Benn et al., 1992). The presence of complete sequences of early Flandrian pollen sequences in sediments from inside the LLS limits in Glen Sligachan and Glen Arroch, was inferred to show that a significant area of the LLS icefield had deglaciated before the rapid warming at the end of the stadial (Benn et al., 1992). It was suggested that initial deglaciation was active, producing the extensive sequences of recessional moraines, and was driven by reduced accumulation as the climate became more arid, before the onset of rapid warming at the end of the LLS caused uninterrupted retreat (Benn et al., 1992). This is in accord with a chironomid-inferred record of mean summer temperatures from Whitrig Bog, in southeast Scotland (Figure 4.14), which showed that the coolest temperatures (approximately 7.5°C) occurred early in the stadial (Brooks and Birks, 2000). Temperatures then increased gradually to about 9°C before increasingly rapidly at the transition into the Holocene (Brooks and Birks, 2000). This warming trend during the latter part of the stadial was also inferred from a similar chironomid record at Abernethy Forest in the central Highlands (Brooks et al., 2012). The notion that LLS glacier retreat commenced during the mid-stadial, under gradually warming temperatures, is supported by recalibrated cosmogenic exposure dates in numerous locations throughout Scotland (Ballantyne, 2012). However, these ages conflict with radiocarbon dates and varve sequences from the Lomond glacier and Glen Roy which indicated that ice advance continued until late in the stadial (Palmer et al., 2010; MacLeod et al., 2011).
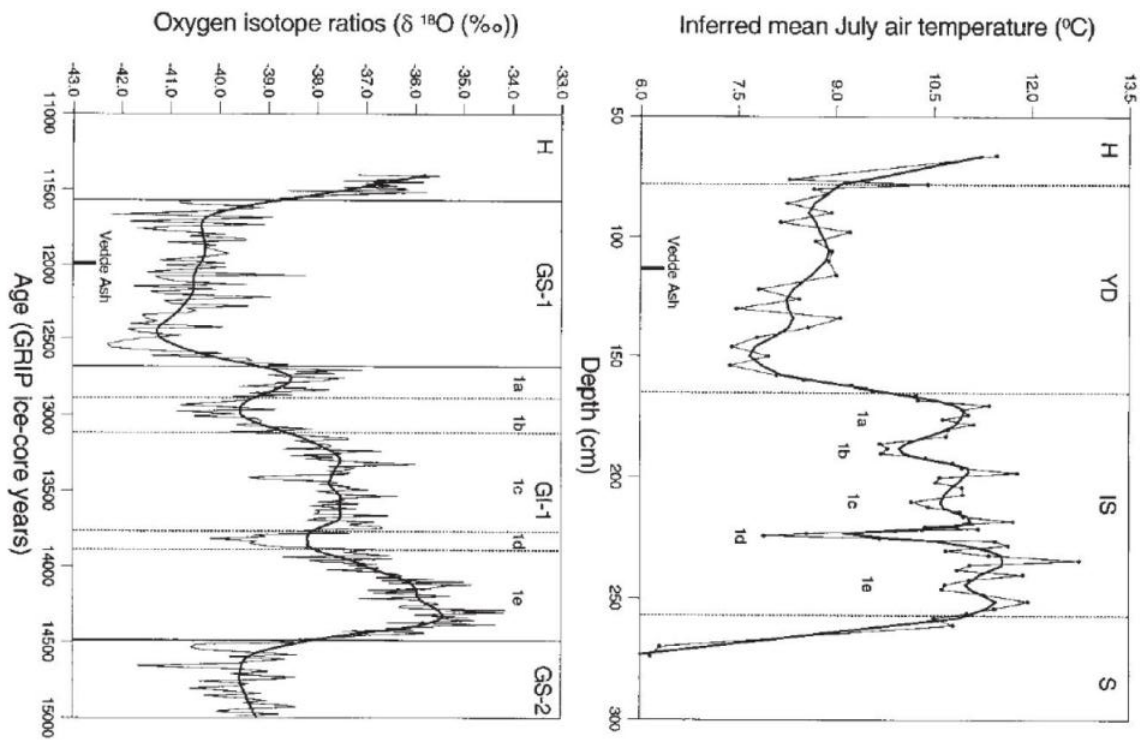
Figure 4.14 Lateglacial and early Holocene chironomid-inferred mean July air temperatures compared with GRIP oxygen isotope data. Both records indicate that the coolest temperatures were experienced early in the stadial, rising gradually initially before rapid warming occurred at the onset of the Holocene. Reprinted from Journal of Quaternary Science, Vol. 15(8), Brooks, S.J. and Birks, H.J.B., Chironomid-inferred Lateglacial air temperatures at Whitrig Bog, southeast Scotland, 759-764. Copyright (2000) with permission from John Wiley and Sons.

This apparent variability in the timing of the response of glaciers to the changing climate is perhaps to be expected. Glaciers with high elevation source areas may have continued to be nourished whilst those in lowland areas may have become isolated from their accumulation areas and began to downwaste more rapidly (Benn et al., 1992). Likewise, glaciers with large catchments may have responded more slowly, perhaps accounting for the continued advance of the Roy and Lomond glaciers under a warming climate. Numerical modelling (Golledge et al., 2008) driven by a scaled GRIP temperature pattern, and under parameters that were able to closely replicate the extent of LLS glaciation as inferred from the empirical evidence, suggested that the maximum extent of LLS glaciation occurred between 12.6 and 12.4 ka BP (Figure 4.13). This again supports the notion that, for many of the LLS glaciers, retreat commenced during the mid-stadial under gradually warming temperatures, rather than being triggered by rapid warming at the close of the stadial. This perhaps explains the dominance of the active retreat deglaciation pattern throughout much of Scotland, with localised variations in retreat dynamics reflecting the effect of local conditions of individual glaciers, such as catchment topography and elevation (including the presence of nearby snow-contributing areas and avalanche prone

slopes), glacier size and hypsometry and microclimate variations, superimposed onto this general trend.

## 4.5 Conclusions

- Compilation of glacial geomorphology associated with the LLS in Britain into a GIS database by Bickerdike et al. (2016) has facilitated the identification of five landsystem models which encompass the majority of the terrain glaciated during the LLS.

- The five landsystems are the cirque/niche glacier landsystem; the alpine icefield landsystem; the lowland piedmont lobe landsystem; the plateau icefield landsystem and the ice cap landsystem.

- These landsystems reflect increasing ice thickness and decreasing topographic control on the flow of ice. The cirque glacier landsystem is typical of localities where conditions were only just above the threshold required for glaciation and is characterised by the restriction of ice to topographically favourable sites.

- Alpine icefields formed in areas where the underlying topography comprised steep slopes and sharp mountain crests. These glaciers were warm-based, producing sequences of recessional moraines, and were confined to interconnected valleys. In a limited number of locations, outlet glaciers from these icefields formed lowland piedmont lobes when they extended out of the valleys onto more open terrain.

- Plateau icefields developed on gently undulating upland surfaces which nourished glacier ice and were drained by outlet glaciers. They had complex thermal regimes of both warm and cold-based ice. The ice cap landsystem is present only at the centre of the West Highland Glacier Complex where the geomorphology is strongly indicative of ice flow discordant to the underlying relief, indicating overriding of the topography by a thick ice cap.

- Three retreat styles are represented by the glacial geomorphology. These include active retreat throughout deglaciation, two-phase retreat and uninterrupted retreat. The active retreat style appears to be the most widespread and accords well with retreat commencing during the mid-stadial, under gradually warming conditions, rather than rapid warming at the stadial's end.

- Retreat style was influenced by a combination of climatic and topographic conditions but the preservation of landforms from which retreat dynamics can be derived also reflects localised variation in debris turnover and availability or the removal of glaciogenic material by proglacial fluvial systems.

- Similarly, the extent and landsystem type of the LLS glaciers was strongly influenced by the interplay of pre-existing topography and climate. The snow-fence effect of the mountains west of Rannoch Moor combined with the prevailing south-westerly winds allowed accumulation of large volumes of snow which became the centre of the West Highland Glacier Complex. Likewise, cirque glaciers in peripheral locations were able to develop at sites downwind of plateau surfaces which offered protection from insolation and a source of windblown snow.

- Regional variations in climate also influenced glacier extent and landsystem type. Northward and eastward reductions in precipitation limited the extent of glaciation at more arid sites, but more recent modelling studies (Golledge, 2008; Golledge et al., 2010) suggest that the LLS climate was generally colder, drier and had greater seasonality than previously believed.

- In identifying the type, distribution and possible controls on landsystems associated with the LLS in Britain, this paper provides a regional framework against which modern analogues can be tested to shed greater light on the processes which drove LLS glacial advance and retreat. These landsystem templates also serve as a useful constraint or test of numerical modelling experiments of LLS glaciation and retreat (e.g. Golledge et al., 2008).

# Chapter 5

# Relative age dating of moraines in the English Lake District using soil chronosequences: a pilot study

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## Abstract

This paper presents a soil chronosequence approach to relative age dating of moraines by utilizing age-related characteristics of soils developed on sharp-crested moraine ridges in the English Lake District. The potential for this technique to differentiate between moraines formed during the Loch Lomond (Younger Dryas) Stadial and those dating to earlier phases of moraine construction is explored. Pits were excavated on inset moraine sequences in nine valleys within the study area where field measurements were taken of soil depth and soil horizon thickness, and samples were collected for laboratory analysis of organic content and clay/silt composition with depth. A predictable pattern of a decrease in organic content and a decrease in silt and clay sized particles with increased depth is observed, but these soil properties do not appear to be related to landform age. However, total soil depth and combined E and B horizon thickness generally increase with age and have been used to identify two probable age populations of moraines within the sample. A cluster of features with combined E and B horizons between 9 and 20 cm thick is thought to represent moraines of LLS age. A second population of moraines with combined E and B horizons of 28 cm thick or more is assigned a pre-LLS age, as calibrated by a Lateglacial site in Mosedale. However, between these two populations is a range of intermediate values, probably reflecting the slowing of soil development rates on moraines greater than 10 ka in age and the short duration of time between deglaciation of the last British Ice Sheet and the onset of the Loch Lomond Stadial. This range of intermediate values and the lack of reliable chronological control for these sites has impeded identification of a threshold value between LLS and pre-LLS soils and consequently, this study has been unable to conclusively distinguish between LLS and pre-LLS moraines across the Lake District using this technique.

## 5.1 Introduction

The Younger Dryas Stadial (YD) refers to the abrupt return to severe cold conditions experienced in the Northern Hemisphere following the Last Glacial Maximum (26.5-19 ka) (Clark et al., 2009). Despite the short duration of this event, which occurred between 12.9 and 11.7 ka (Golledge, 2010), the decline in temperatures was sufficient to drive glacier readvance or regrowth in many regions around the margins of the North Atlantic Ocean, including in North America, Canada, Iceland and Scandinavia (Mott and Stea, 1993; Andersen et al., 1995; Lowell et al., 1999; Ingólfsson et al., 2010). In Britain, the period, which is known locally as the Loch Lomond Stadial (LLS), was marked by mild summers and much colder winters (Golledge, 2008), that resulted in the regrowth of a substantial icefield over much of the western Highlands, which was flanked by a number of satellite icefields, ice caps, valley and cirque glaciers around its periphery (Golledge, 2010).

Dating of landforms produced by these glaciers is important for several reasons. For example, absolute dating of terminal moraines that mark the maximum extents of LLS glaciation, allows an assessment of the timing and synchroneity of LLS glacier advance and retreat, both in the UK (e.g. Ballantyne, 2012) and further field (e.g. Ivy-Ochs et al., 1999). This provides an insight into the spatial pattern of deglaciation and the potential factors which drove ice retreat through time (for example climate change). Furthermore, the landform signature of LLS glaciation is superimposed onto features formed during the previous phase of glaciation, when the entirety of Scotland and Ireland, the majority of Wales and much of England were covered by the last (Late Devensian) British-Irish Ice Sheet (Clark et al., 2012). It is not always possible to distinguish between LLS and older landforms based on their morphology; many early studies (e.g. Gray and Brooks, 1972; Sissons, 1974, 1979c, 1980a) commonly used the concept of moraine freshness to differentiate between younger sharp-crested moraines, thought to date from the LLS, and more subdued older features. More recently, however, this approach has been questioned (Wilson, 2002). Dating of landforms is therefore crucial in ensuring that features are not erroneously attributed to a particular glacial period, as this can lead to incorrect palaeoglaciological reconstructions and palaeoclimatic inferences being made. However, despite both the importance of chronological constraints and advancements in dating techniques, the number of dates on LLS landforms is both relatively low and spatially uneven.

One region where differentiating between LLS and older features has proved particularly contentious is the English Lake District. The first detailed geomorphological mapping of LLS glacial landforms in this area was conducted by Sissons (1980a), who reconstructed a

series of 64 alpine style valley and cirque glaciers (Figure 5.1). The maximum extents of 40 of these were marked with clear terminal moraines whilst the limits of the others were inferred from the distribution of hummocky moraine and drift limits. Subsequent remapping of the geomorphology (McDougall, 1998, 2001, 2013), combined with an increased understanding of the glaciology of modern-day plateau icefields in Norway and Iceland (Rea et al., 1998; Rea and Evans, 2003; Evans et al., 2006), informed by Manley's (1959) theoretical concept of critical summit breadths above equilibrium line altitudes (ELAs), has prompted a radical reinterpretation of the geomorphological evidence in the Lake District. It has been argued that, during the LLS, substantial plateau icefields as well as, and in some cases, instead of, cirque glaciers developed in the upland areas of the central and eastern Lake District (Rea et al., 1998; McDougall, 1998, 2001, 2013; Brown et al., 2011, 2013; Evans, 2013). Ice on the plateaux was drained by a series of outlet glaciers in the valleys, which in some cases had similar extents to those reconstructed by Sissons (1980a) but which in many cases were much more extensive (Figure 5.1). There is strong evidence in favour of a plateau icefield configuration, such as the presence of moraines at the heads of particular valleys, including Greenup Gill, Langstrath and Hayeswater, which lead up onto the plateau surfaces themselves (McDougall, 2001, 2013). Furthermore, a plateau icefield configuration accounts for the variations in extent and ELA of the glaciers reconstructed by Sissons (1980a). Numerical modelling, driven by an ELA record scaled from the GRIP record, retrodicted a much greater extent of ice than proposed by Sissons (1980a), both on the plateau surfaces and in the valleys (Brown et al., 2013).

Whilst it seems highly probable that the LLS glaciation in the Lake District was more extensive than suggested by Sissons (1980a), chronological constraint is poor and the age of landforms, and consequently the extent of LLS glaciation, remains uncertain (Wilson et al., 2013). Early dating work generally involved contrasting the lithostratigraphic and palynological characteristics of cores taken from sites inside and outside of the proposed glacier limits (Walker, 1965, 1966; Pennington, 1978) (Figure 5.1). If a site escaped glaciation during the LLS, a Lateglacial tripartite sequence of sediments should be preserved whereas at sites that were glaciated during the LLS, the oldest sediments will have a Holocene age. However, the number of sites in the Lake District suitable for this "inside"/"outside" differentiation approach is relatively few, particularly within the LLS limits, and the results of this approach have been challenged in some cases (e.g. McDougall, 2001). In some locations, cosmogenic isotope dating has been used to provide absolute dates for deglaciation (Ballantyne et al., 2009; Hughes et al., 2012; Wilson et al., 2013). This technique has been used to assign LLS ages to moraines, for example in Keskdale (Hughes et al., 2012) and Lingmell Gill (Ballantyne et al., 2009). However, at

other sites, including Rosthwaite, Watendlath and Wythburn (Figure 5.1), the majority of dates collected were anomalously old (several indicating an implausible pre-LGM age), suggesting that most samples are affected by nuclide inheritance (Wilson et al., 2013). Given both the very resistant nature of the volcanic rocks in the central Lake District and the short glacier transport distances involved, it is very possible that glacial erosion did not remove the 2-3 m of rock required to reset the 'cosmogenic isotope clock' (Ballantyne et al., 2009, Wilson et al., 2013). The difficulties associated with previous attempts to constrain the age of moraines in the Lake District make this a critical location to trial new approaches to dating these landforms.

This paper details the results of an experimental pilot study to explore whether soil chronosequences can be used to differentiate between glacial landforms of different ages, specifically between LLS and pre-LLS aged moraines in the Lake District. This method has been applied to multiple moraine assemblages in various locations, including Labrador, Canada (Evans and Rogerson, 1986), the Sierra Nevada and Wyoming, USA (Birkeland and Burke, 1988; Hall and Shroba, 1993; Berry, 1994), Norway (Mellor, 1987; Messer, 1988; McCarroll and Ware, 1989; Evans, 1999), and South Georgia (Bentley et al., 2007), but it has not previously been applied to moraines in Britain. The technique involves measurement of the characteristics indicative of soil development (such as total soil depth, B horizon thickness and colour) to infer a relative age for the soil and, therefore, for the landform on which it has developed. Usage of this technique elsewhere has proved a simple and inexpensive way to differentiate between assemblages of moraines of different age groups (Evans and Rogerson, 1986; Evans, 1999). As such, this study aims to test whether soil chronosequences can be used to identify whether there are different populations of soil development (and thus age) within moraine sequences commonly associated with LLS glaciation and, if so, whether these can identify the limits of LLS glaciation in relation to those proposed by Sissons (1980a) and McDougall (1998, 2001, 2013). Application of this technique to moraines in the Lake District is particularly attractive, given the problems associated with other dating methods in this region.
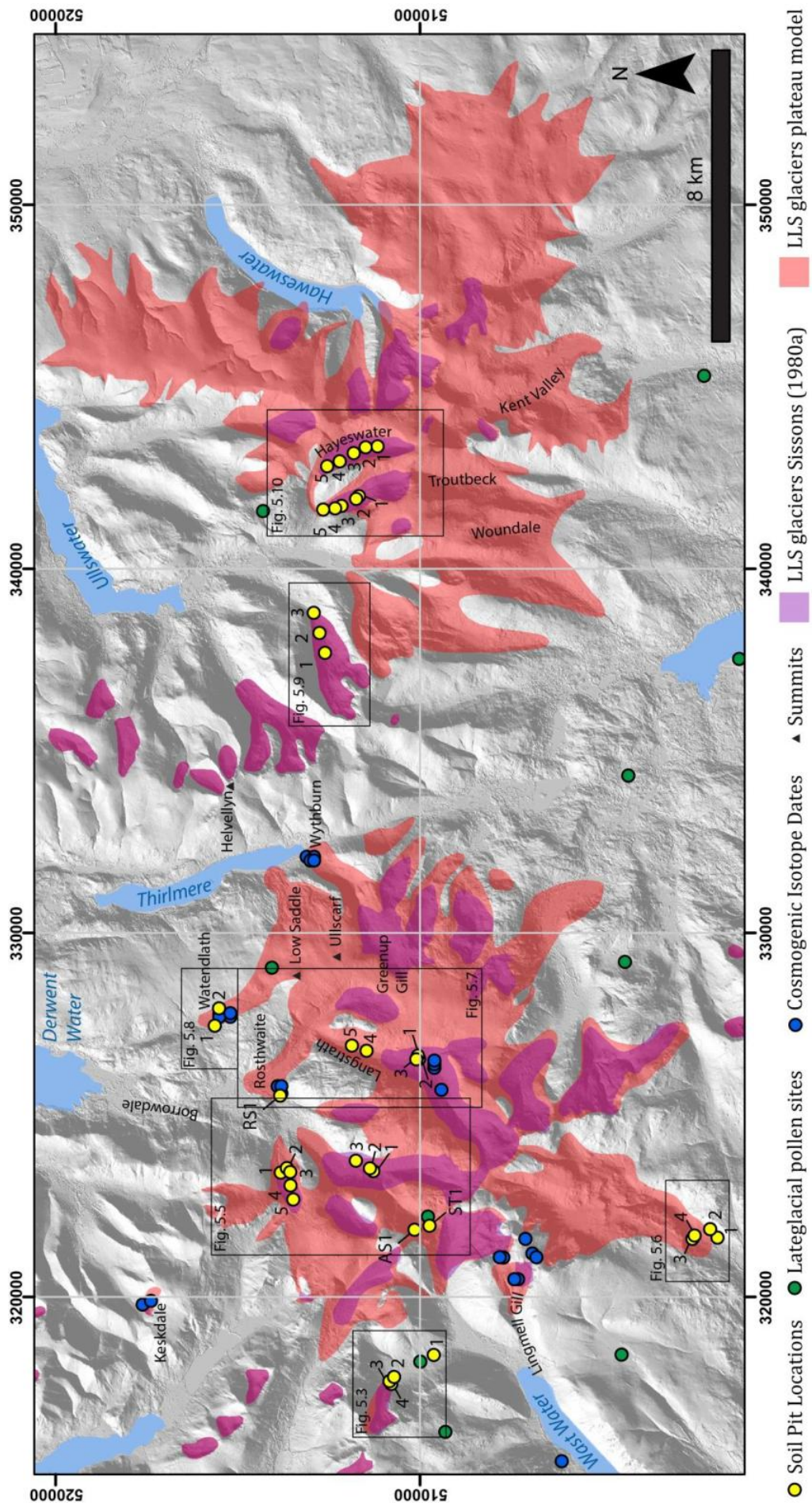
Figure 5.1 Lake District study site. Locations of soil pits sampled for this study shown in yellow alongside locations of existing dates. The extents of Sissons' (1980a) alpine style glaciers are shown in purple, overlaying the more extensive plateau icefield configuration proposed by Brown et al. (2013) and McDougall (2001, 2013).

## 5.2 Methods and site selection

Soil chronosequences are suites of soils formed under similar climatic, topographic and vegetation conditions, the spatial position of which is assumed to relate to time since formation (McCarroll, 1991; Huggett, 1998). The principle of the technique is that in a developing/maturing soil there are morphological, chemical and mineralogical characteristics that evolve with time and that these changes can be measured to give an indication of the relative age of the soil itself and of the underlying landform (Birkeland, 1978). For example, a series of recessional moraines, where landform age increases with distance from the glacier source, should show a parallel increase in soil age, as indicated by these developmental characteristics.

For this study, soil pits, approximately 40x40 cm in area, were excavated on the crests of moraines in each of the sampled valleys (Figure 5.1). For each pit, the turf layer was removed intact and the excavated soil was deposited onto a tarpaulin, enabling the pit to be refilled easily and ensuring that after a short period, evidence of the pit is untraceable. The majority of excavated soil profiles (e.g. Figure 5.2) comprise a thin layer of organic matter and vegetation on the surface, underlain by a dark, often black, layer of top soil (the A horizon) which is rich in organic material. In podzols, the most common soil type in the Lake District valleys, the A horizon is underlain by a lighter grey layer from which clay and other minerals have been eluviated or leached (E horizon). Below this subsoil (the B horizon) is present, comprising parent material which has been altered by pedogenic processes and in which minerals accumulate, usually having been leached from the horizons above. This layer is firmer in texture and contains less organic material than the layers above. This B horizon is usually brighter in colour than the horizons above and below it and may take the form of different layers over which a transition in soil characteristics may occur. At the base of each pit, the C horizon represents unchanged parent material and is recognisable by its very low organic content, lighter colour and increase in gravel sized particles. In some instances, a layer of oxidised parent material (C ox; Birkeland, 1978) is found below the B and C horizons.
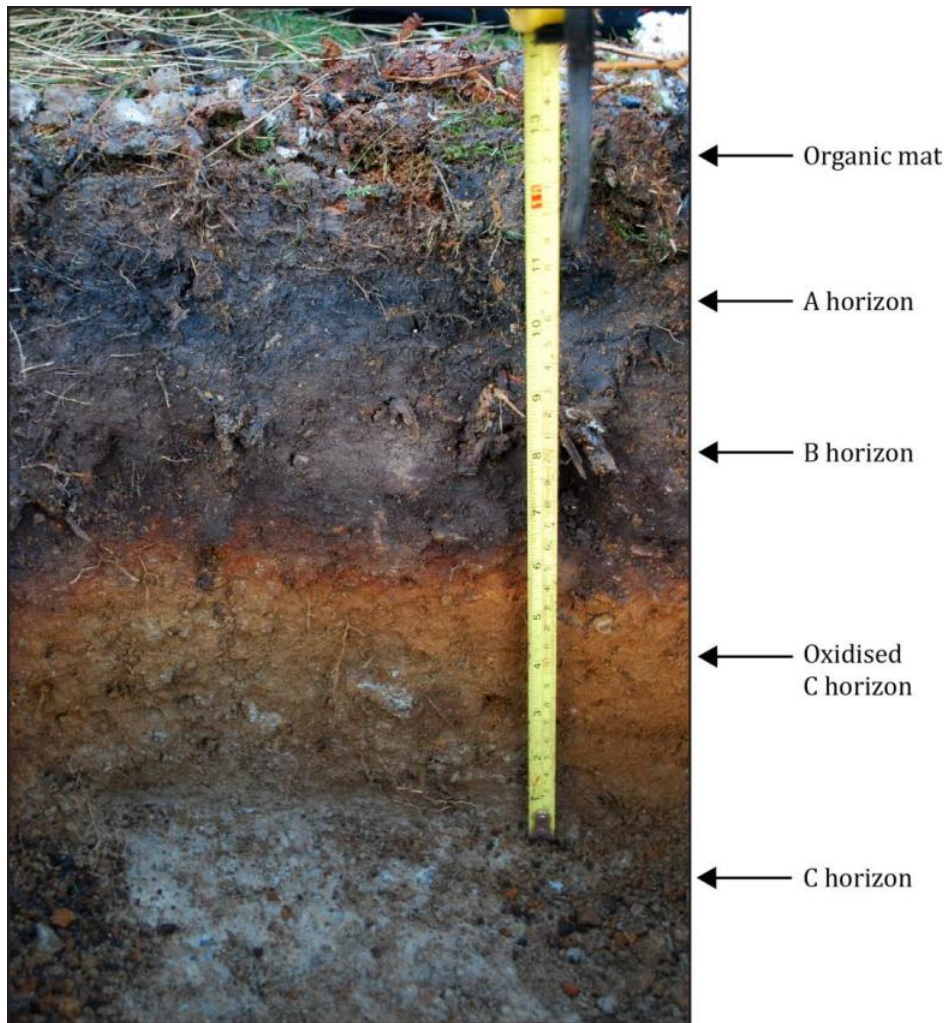
Figure 5.2 Example of a soil profile from Deepdale, Lake District. The top of the profile displays an organic mat, below which the black layer represents the organic rich A horizon. Below this is a B horizon which transitions from a more grey-brown colour to darker shades at its base. At the bottom of the profile there is an orange oxidised C horizon and the more olive-grey unchanged parent material (C horizon).

Soil horizons, and visible layers within each horizon, were identified based on field observations, were photographed and their depths recorded. A sample was taken from each for laboratory analysis. The colour of the wet soil samples was determined using a Munsell colour chart. The colour enrichment (reddening) of each layer was calculated using the colour development equivalent (CDE) approach of Buntley and Westin (1965), for which the chroma is multiplied by a soil colour value where 10R = 7, 2.5YR = 6, 5YR = 5, 7.5YR = 4, 10YR = 3, 2.5Y = 2 and 5Y = 1. Between 3 and 5 g of each sample was dried at 105°C overnight to remove moisture and was then ignited at 550°C for 4 hours in a furnace. This allowed loss on ignition (LOI) to be calculated, giving an indicator of organic matter content. Particle size analysis was conducted using a laser granulometer to provide a measure of silt/clay translocation. Peroxide digestion was performed on each sample to remove organic material, leaving approximately 0.5 g of sediment for analysis. The LOI

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and cumulative percentages of silt and clay sized particles for A, E and B horizons, were plotted against the mid-point depths of each layer from which the samples were collected. Where samples could be collected for the C ox and C horizons, these were processed but, because these layers were not present in all pits, they are not depicted graphically, ensuring that the plots could be compared directly. Soil horizon thicknesses were used to construct soil profiles for each pit, with 1 mm on the soil profile representing 1 cm of soil depth in the field, and were overlain onto maps of the sampling sites. The approximate distance that each pit lies along the former glacier flowline from either the cirque backwall, or former ice-divide position in the instance of plateau icefields, was measured to allow the down-valley (increasing age) progression of characteristics to be plotted. The chronosequence of each valley is detailed within the Results (Section 5.3), whilst the cumulative results and the regional chronosequence are considered in the Discussion (Section 5.4).

Jenny (1941) identified five factors which drive soil formation: parent material, topography, climate, organisms and time. In order to assume that the soil characteristics measured can be considered to be a surrogate for time, all those factors other than time need to be kept consistent. For example, all sampling locations have a similar bedrock geology of Eycott and Borrowdale Volcanics (Evans and McDougall, 2015), from which a local till, the Blengdale Glacigenic Formation, has been derived (Evans, 2015b). This provides a consistent parent material from which the soil has formed. Likewise, topography is known to influence soil formation, including at the individual landform scale (McCarroll and Ware, 1989). For example, on a single moraine, microtopography can influence the volume of precipitation received, the duration of snow cover and the distribution of vegetation (Parkinson and Gellatly, 1991). Exposure to wind on ridge crests may lead to finer particles being winnowed away from the soil, giving the false impression of coarser grain sizes. McCarroll and Ware (1989) found that individual moraine morphology also impacted the process of soil development; the shallowest brown soils formed on steep, exposed ridges which had minimal translocation potential, whilst deeper podzol soils formed on broad, subdued moraine crests.

To overcome the potential influence of topography on soil development, all soil pits within this study were sited on crests of relatively sharp moraine ridges, ensuring that all the soils sampled were likely to be well-drained. The necessity to sample only well-drained soils meant that valleys where moraines are present but are less well-drained, could not be sampled. The close proximity of the sites, which all occur within 26 km of each other, minimises spatial variability in climatic conditions between sites. Likewise, the proximity

of sites to each other ensures that all valleys should have experienced a relatively similar vegetation history and all samples were taken from areas with the same vegetation pattern of grasses and occasional bracken, with no pits being excavated in areas that had clearly been afforested and/or improved pasture. A particular consideration has been to minimise the likelihood of previous disruption of the soil during formation, such as by ploughing or settlement, which would inhibit clear horizon formation. Consequently, there was a greater availability of younger moraines in upland valley sites than older features, because the latter are more likely to be located in lower elevation areas and are consequently more likely to have been anthropogenically modified.

In addition to fulfilling the requirements above, sampling sites were selected so that moraines of a variety of proposed ages would be included, allowing potential identification of different populations of features. These included i) moraines proposed by Sissons (1980a) to be of LLS age; ii) moraines proposed by McDougall (1998, 2001, 2013) or Brown et al. (2013) to be of LLS age, but which fall outside Sissons' (1980a) limits; and iii) moraines which fall beyond the limits of LLS glaciation in all published reconstructions. However, although it would have been desirable to sample moraines of each of these three groups in every valley, this was not possible. For example, in Deepdale, Sissons' (1980a) limit incorporates the outermost moraine in the valley and no older moraines were present down-valley of this limit for sampling. Likewise, no moraines were present down-valley of McDougall's (1998, 2001) limit for the Rosthwaite and Watendlath glaciers, although cosmogenic isotope dating suggests that these features do predate the LLS (Wilson et al., 2013). Mosedale was selected for sampling as, uniquely, this site hosts moraines assigned by Sissons (1980a) and Brown et al. (2013) to the LLS and chronological control which supports moraines in the lower valley predating the LLS (Evans et al., 2015).

A clear difference in soil development between those moraines assigned to the LLS by Sissons (1980a) and those outside these limits (either inside the more extensive plateau icefield limits or outside of them), could indicate that only these moraines date to a later readvance (i.e. the LLS) and that LLS glaciation was restricted in its extent. If a clear difference in soil development is apparent between moraines within the limits of the more extensive LLS icefield and those outside of the limits, this would imply that the proposed limits of the plateau icefield represent the limits of LLS glaciation. If no different populations are apparent throughout the total sample of moraines, then this would suggest one of three scenarios: i) that all features within the sample predate the LLS (which conflicts with the distribution and morphology of the landform evidence, absolute

dating which supports the existence of LLS glaciers in the Lake District and numerical modelling); ii) that all sampled moraines are LLS in age (requiring a very extensive LLS icefield to form the outermost features), or iii) that the resolution of the soil chronosequence technique is insufficient to reflect differences in landform age within this area.

## 5.3 Results

### *5.3.1 Mosedale*

Mosedale, the westernmost sampling location, is a crucial site in this study because independent chronological constraint already exists for the moraines in this valley (Ballantyne et al., 2009; Evans et al., 2015). Morainic mounds are present on the valley floors of both Mosedale and Wasdale, into which Mosedale feeds (Figures 5.3 and 5.4). The majority of these moraines have been attributed to the later stages of deglaciation of the Late Devensian Ice Sheet, with the limit of LLS glaciation having long been associated with a pair of sharp-crested latero-frontal moraines, at Black Comb (around 200 m OD) in the upper valley (Sissons, 1980a; Brown et al., 2013) (Figure 5.4). Pits MD4 and MD3 were sampled on these ridges. In the lower valley, at approximately 140 m OD, a circular, sharp-crested ridge, infilled by peat, has been identified as a snow avalanche impact pit (Brown et al., 2011, 2013). Lithostratigraphic and pollen analysis of a core taken from within this pit revealed a Lateglacial tripartite sequence of sediments (Evans et al., 2015). This indicates that at least the moraines down-valley of this site, including MD1, predate the LLS. Both the total soil depths down to the base of the B horizon and the combined thickness of E and B horizons increase sequentially along this valley (Figure 5.3). Combined E and B horizon thickness and total soil depth do not appear to indicate that MD3 and MD2 belong to two different age populations, as published reconstructions of the extent of LLS glaciation (Sissons, 1980; Brown et al., 2013) suggest they do; indeed, the difference between them is comparable to the difference between proposed LLS samples, MD3 and MD4. A smaller difference is observed between the two 'older' pits (MD1 and 2), particularly with regard to total soil depth to the B horizon base. Conversely, no trend in B horizon reddening was apparent, MD1, 2 and 4 having a CDE of 9, whilst MD3, the only podzol at this site, displays a CDE of 16. In each pit, the percentage LOI decreased with depth (see Appendix 2). Both the oldest and youngest pits (MD1 and MD4 respectively) have comparable LOI percentages (A horizons of 32% and 29%, upper B horizons of 22% and 21%, lower B horizons of 11% and 14%), and there is no sequential pattern of these values in the intermediate pits. Likewise, the percentage of soil that comprises silt and clay sized particles generally decreases down the soil profile but there is no sequential change

in either the maximum or minimum B horizon silt/clay percentages with distance down-valley.
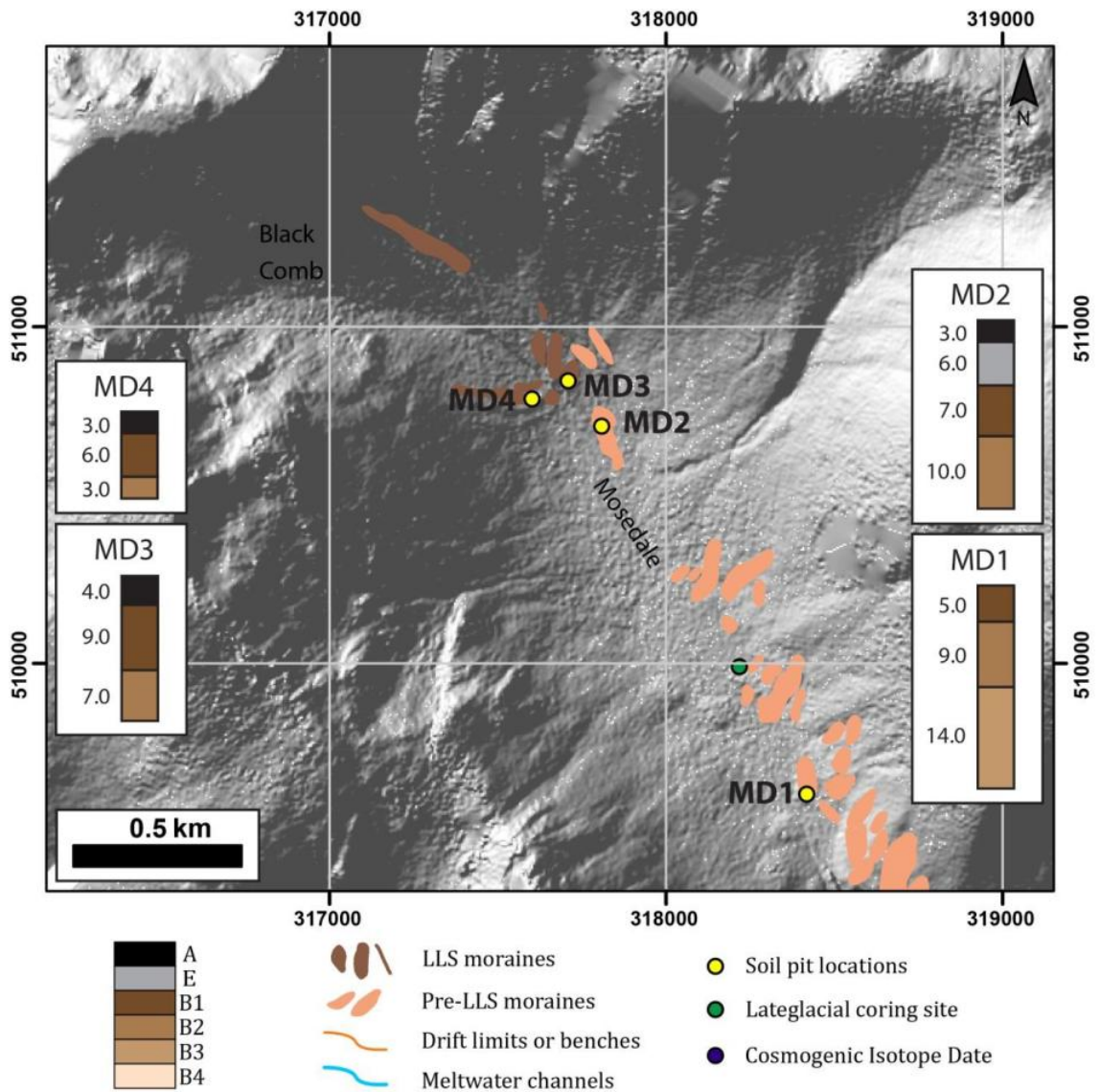


Figure 5.3 The soil chronosequence for Mosedale. For each of the four soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. MD3 and 4 were sampled on LLS moraines proposed by Brown et al. (2013) and Sissons (1980a) whereas MD1 and 2 were sampled on 'pre-LLS' moraines. A pre-LLS age for MD1 is confirmed by the presence of Lateglacial sediments in the snow avalanche impact pit (as indicated). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Figure 5.4 Moraines in Mosedale. The older, larger moraines, including the site of MD1, are visible in the foreground whilst the moraines marking the LLS limit can be seen beside the stream, just below the snowline at the rear of the image. Photograph from H.L. Bickerdike.

Table 5.1 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence in Mosedale. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Mosedale | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| MD4 | 9.0 | 9.0 | 12.0 | 9 | 1.52 |
| MD3 | 16.0 | 16.0 | 20.0 | 9 | 1.58 |
| MD2 | 23.0 | 17.0 | 26.0 | 16 | 1.74 |
| MD1 | 28.0 | 28.0 | 28.0 | 9 | 2.97 |

## 5.3.2 Honister

A well-developed and continuous sequence of recessional moraines, which are orientated obliquely down-valley, is found at around 300 m OD in Little Gatesgarthdale, the valley to the east of the Honister Pass (Figure 5.5). Moraines in this valley all fall within the LLS limits proposed by both Sissons (1980a) and McDougall (1998, 2001) and were sampled to test whether they appeared to represent a single age population, as would be expected given their location. In the upper and middle valley, these features are particularly clear and can rise to between 4 and 5 m in height. Many of the areas between the moraines are filled with peat, subduing the relief. There is a gradual transition between these features and more subdued ridges in the lower valley (McDougall and Pearce, 2015) and thus sampling further down-valley was not attempted as the subdued relief of moraines would suggest that these features are less well-drained. Five pits were sampled at this site. Total soil depths increase from 21 cm at the youngest site (HON 5), to the greatest depth at HON3 (32 cm). However, the sequence is interrupted by an anomalously shallow, clast-rich soil at HON 2 (13 cm deep). Similarly, although the three most up-valley pits at Honister show an increase in combined E and B horizon thickness, the pattern is interrupted by the anomalously shallow soil at HON2. The second oldest site at HON1 has a total soil depth (21 cm) that is comparable to that of HON5, and a combined E and B horizon thickness of 18 cm, which is comparable to that of HON 4. All of the moraines sampled appear to belong to a single age population. There is no sequential trend in B horizon CDEs between these pits, this value fluctuating between 6 and 18. Whilst there is a decrease in both the percentage of mass LOI and silt/clay sized particles down the soil profiles (Appendix 2), there appears to be no coherent pattern in these characteristics between pits.

Table 5.2 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence at Honister Pass. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Honister | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| HON5 | 14.0 | 9.0 | 21.0 | 18 | 1.37 |
| HON4 | 19.0 | 19.0 | 23.0 | 9 | 1.76 |
| HON3 | 23.0 | 19.0 | 32.0 | 16 | 2.13 |
| HON1 | 18.0 | 14.0 | 21.0 | 16 | 2.20 |
| HON2 | 9.0 | 9.0 | 13.0 | 6 | 2.25 |

Figure 5.5 The soil chronosequence for Honister Pass and Seathwaite. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. A possible deepening of soil and thickening of B horizon can be observed from the younger to older moraines at Honister Pass but this pattern is disrupted by a particularly deep pit at HON3 and a much shallower soil at HON4. McDougall (2001) proposed that the LLS glacier extended down-valley as far as Seatoller, whereas Sissons (1980a) argued that it lay in the vicinity of HON1. The chronosequence between Styhead Tarn and Seathwaite is also shown. An increase in both soil depth and B horizon thickness with age can be observed. Legend on Figure 5.3. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

### 5.3.3 Seathwaite

According to McDougall's (2001) reconstruction, the Seathwaite glacier was nourished by ice on the plateau and in the valleys to the east of Great Gable. Five pits were sampled in this valley, two in the vicinity of Styhead Tarn and Aaron Slack in the upper valley (STY1 and AS1) at approximately 500 m OD, and a further three in the lower valley (at approximately 180-200 m OD) below the confluence with Grains Gill (Figure 5.5). All of the sample sites fall inside the limits of LLS glaciation proposed by McDougall (1998, 2001) but Sissons' (1980a) reconstruction suggested that the tributary valley leading to Styhead Tarn was ice-free. These sites were sampled to ascertain whether the moraines in the lower valley (GG1-3) represented a different age population to those at AS1 and STY1. With the exception of the second most up-valley pit (AS1), the soils show a progressive increase in depth, from 16.5 cm soil depth at the uppermost pit in the valley (STY1), to 28 cm for the oldest pit (GG3). Likewise, the combined thickness of the E and B horizons increases from 12.5 cm and 13 cm in the upper valley (STY1 and AS1), to 20.5 and 23cm (GG2 and GG3 respectively). There is no apparent relationship between soil reddening and relative age, because CDEs calculated for the B horizons do not change sequentially down-valley. Although the percentage of mass LOI generally decreases down each soil profile, there is no relationship with relative soil age; B horizons at the youngest site have low organic content (a maximum LOI of 7%), whereas the B horizons in the subsequent pit (AS1) have extremely high organic content, with an LOI of 83% (Appendix 2). Soils in the lower valley had upper B horizons where LOI percentage ranged from 28-30%. It is possible that there is a general increase in the percentage of silt and clay particles in the B horizons with distance down-valley, but the (often large) variation between the upper and lower B horizons makes it difficult to confidently assess the presence, and strength of this relationship.

Table 5.3 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence between Styhead Tarn and Seathwaite. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Seathwaite | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| STY1 | 12.5 | 10.0 | 16.5 | 12 | 2.16 |
| AS1 | 13.0 | 13.0 | 15.0 | 3 | 2.55 |
| GG1 | 16.5 | 12.5 | 20.5 | 12 | 3.20 |
| GG2 | 20.5 | 13.5 | 25.5 | 16 | 3.29 |
| GG3 | 23.0 | 15.0 | 28.0 | 9 | 3.72 |

### *5.3.4 Eskdale*

The upper reaches of Eskdale are thought to have been occupied by the southwestern lobe of the LLS plateau icefield, as reconstructed by Brown et al. (2013), whereas Sissons (1980a) argued that this area had remained ice-free. This site was chosen for sampling as, if Brown et al.'s (2013) age assignments are correct, this valley should host both LLS and pre-LLS moraines ridges. Four pits were excavated on moraines, at approximately 100 m OD, near the proposed limit of the LLS glacier (Figure 5.6). Two pits, ESK 1 and 2, were on moraines just outside or on the limit, and two, ESK 3 and 4, just inside it, as inferred from the reconstruction proposed by Brown et al. (2013). The soil depths in these pits were significantly deeper than most observed in the Lake District study area; total soil depths to the base of the B horizon varied from 26 to 39 cm and the combined E and B horizons measured 28 cm (ESK4), 37 cm (ESK3), 25 cm (ESK2) and 26 cm (ESK1) from youngest to oldest. This does not appear to indicate that these moraines belong to two different populations within the valley, as would be expected if Brown et al.'s (2013) age assignments are correct, though this could reflect the small number of moraines sampled. Furthermore, given that these soils are much deeper than those sampled in other locations, it seems reasonable that they may belong to an older age population. The unfortunate lack of sharp-crested moraines further up-valley of these sites impeded sampling of additional soil pits to determine whether a transition to apparently younger moraines occurs within this valley. With so few samples in the valley, it is difficult to assess whether B horizon reddening increases with age (CDEs vary from 6 to 12 to 9 to 8 from the youngest moraine to the oldest). Likewise, although there is a decrease in both LOI and silt/clay particle size percentages with increased depth, no clear trend is apparent in relation to relative age. Lower concentrations of organic material were present in the B horizons of the apparently younger moraines (ESK3 and 4). The range in silt/clay percentages between the upper and lower B horizon layers suggests that variation between layers of the B horizon in the same pit are greater than any relationship with age.
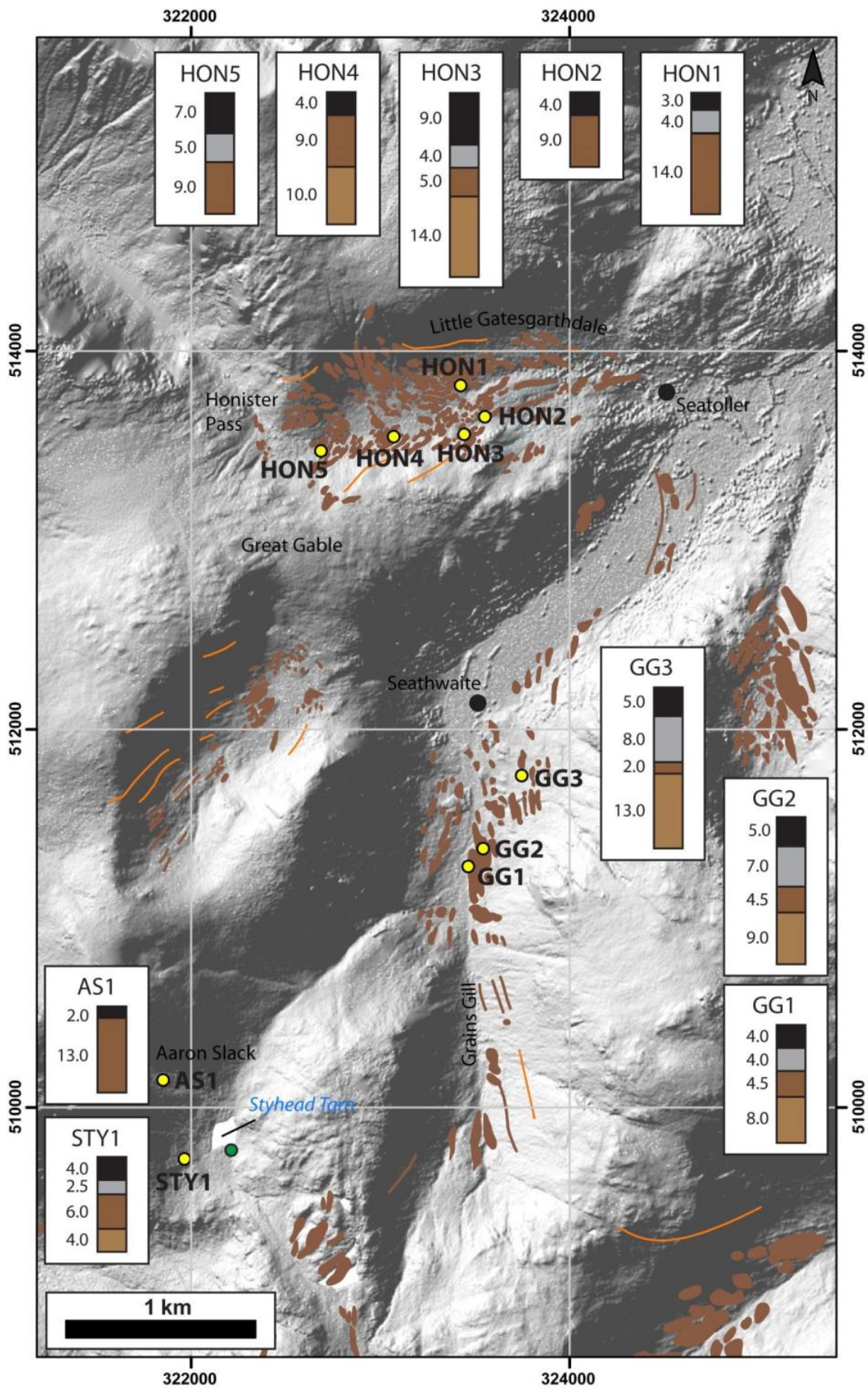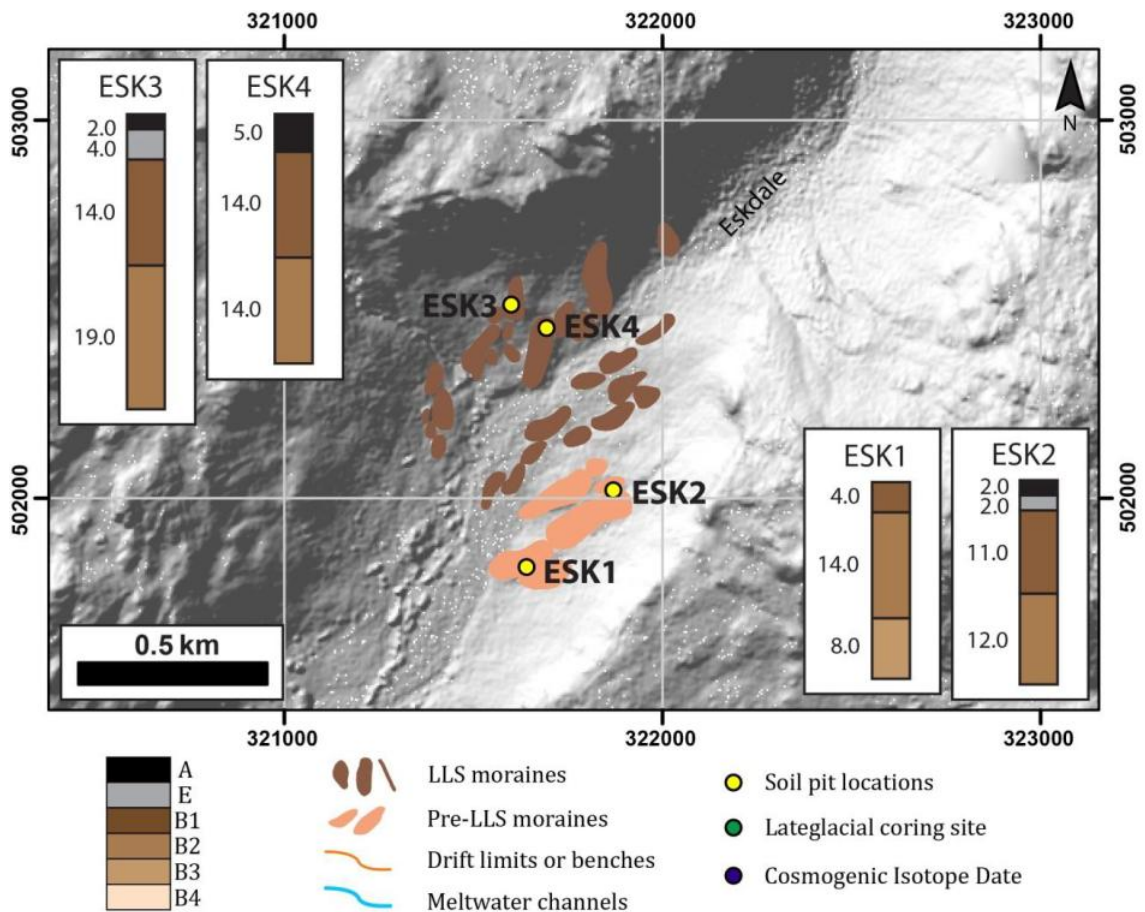
Figure 5.6 The soil chronosequence for Eskdale. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. The inner moraines (ESK3 and 4) show particularly deep soils and thick combined E and B horizons compared to those found in the majority of sampled valleys. The measurements are comparable with the Lateglacial moraine in Mosedale, suggesting that the features in Eskdale may predate the LLS. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Table 5.4 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence in Eskdale. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Eskdale | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---------|-------------------------------|--------------------------|----------------------|-----|-------------------------------------|
| ESK4 | 28.0 | 28.0 | 33.0 | 6 | 5.98 |
| ESK3 | 37.0 | 33.0 | 39.0 | 12 | 5.98 |
| ESK2 | 25.0 | 23.0 | 27.0 | 9 | 6.33 |
| ESK1 | 26.0 | 26.0 | 26.0 | 8 | 6.60 |

### 5.3.5 Rosthwaite and Langstrath

Langstrath valley descends northwards from the central upland area of the Lake District, descending into Borrowdale at the village of Rosthwaite (Figure 5.7). Prominent recessional moraines are present in the upper valley, particularly clear examples of which are found below Langdale Combe and at Stake Pass. Three inset moraines, at approximately 230 m OD, were sampled in this area of closely-spaced moraine ridges (LN1, 2 and 3). These sites are approximately 500 m down-valley of a series of surface exposure dates (indicated on Figure 5.7), which range from a maximum age of between 18.1±1.3 and 17.4±1.2 ka and a minimum of 13.1±0.9 and 13.6±0.9 ka, depending on the isotope production rate applied (Wilson et al., 2013). Two previously unmapped moraines were sampled in the lower Langstrath Valley, one at Blea Rock and the other about 400 m further down-valley (170 and 160 m OD). The final moraine sampled, at the foot of the valley, is a large terminal moraine between the villages of Rosthwaite and Stonethwaite. Although further sampling in the vicinity of this terminal moraine was attempted, excavated pits showed signs of anthropogenic activity and thus were unsuitable for inclusion in this study. The Rosthwaite site is attractive for soil sampling because of the presence of four surface exposure ages on boulders on and within the moraine. However, since three of these ages predate LGM glaciation, it is highly probable that they have been affected by nuclide inheritance, and are therefore unreliable. The fourth age (between 19.0±1.1 and 16.6±1.0 ka) suggests that these moraines may have formed during LGM deglaciation but, given the nuclide inheritance of the other samples, inferences made from this single date should be made with caution. This site was selected as it includes i) moraines within Sissons' (1980a) limit, ii) moraines outside Sissons' limit but inside McDougall's (1998, 2001) limit and iii) the Rosthwaite moraine, which cosmogenic isotope dating suggests predates the LLS. Thus, this site provided an opportunity to investigate whether any age breaks, as inferred from sudden changes in soil development, could be identified in this moraine sequence.

The total depth of soil varies between 23 and 18 cm for the pits in the upper and middle reaches of the valley, aside for the anomalously shallow LN2, before increasing markedly on the Rosthwaite moraine (RS1, 30 cm) (Figure 5.7). This trend is mirrored by the combined E and B horizon thicknesses; shallow soils at LN2 and LN4 interrupt what may be a general trend of increased thickness of soil horizons with distance down-valley but the small number of pits reduces confidence in this pattern. However, the Rosthwaite moraine again represents a substantially more developed soil, and thus may belong to an older population of moraines within the Lake District. It is unfortunate that moraines are largely absence from between pits LN5 and RS1, making it impossible to investigate this

pattern further. B horizon reddening does not change sequentially along the valley, with the high values having been detected in samples from both the upper and middle reaches of Langstrath, and therefore is probably not reflective of the relative age of the landforms. The percentages of mass LOI and silt/clay sized particles both decrease with depth in the soil profile (Appendix 2). However, the LOI percentage in the B horizons seems relatively consistent between pits; with the exception of LN5, the percentage of LOI of the upper B horizon of each pit ranges from 15 to 19%, but there is no sequential trend between pits. Both the minimum and mean B horizon LOIs also show no pattern with age. The same is true of the percentage of silt and clay sized particles which fluctuate between pits but do not change consecutively with distance down-valley.

Table 5.5 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence at Rosthwaite and Langstrath. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Langstrath | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| LN2 | 9.5 | 8.0 | 13.5 | 8 | 3.75 |
| LN3 | 16.0 | 10.0 | 21.0 | 9 | 3.78 |
| LN1 | 19.0 | 15.0 | 23.0 | 24 | 3.84 |
| LN4 | 13.0 | 10.0 | 18.0 | 16 | 5.14 |
| LN5 | 19.0 | 15.0 | 23.0 | 9 | 5.58 |
| RS1 | 30.0 | 30.0 | 30.0 | 16 | 8.82 |
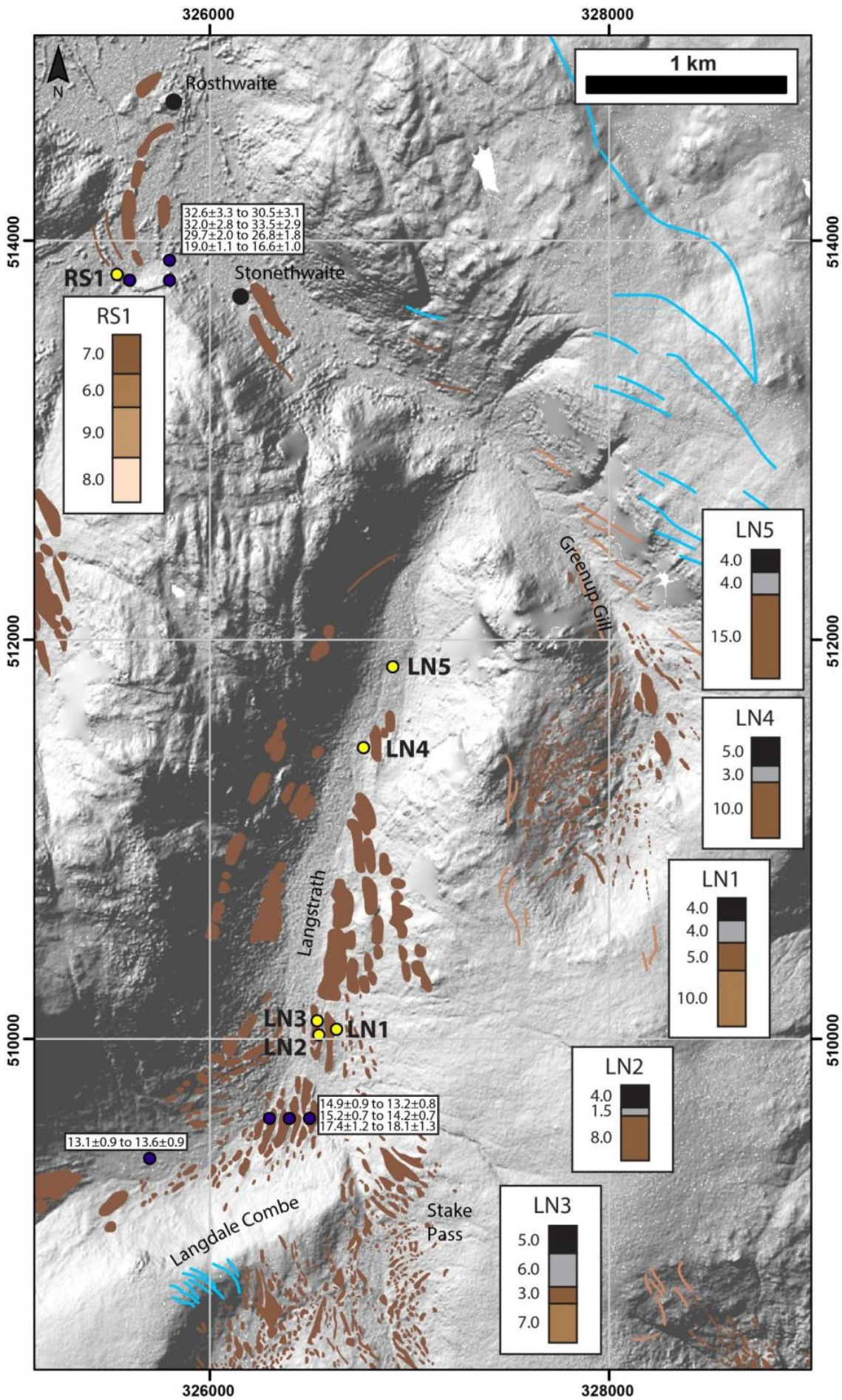
Figure 5.7 The soil chronosequence for Rosthwaite and Langstrath. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. Both combined E and B horizon thickness and total soil depth show an increase in thickness between LN5 in the middle reaches of Langstrath and the Rosthwaite moraine, suggesting that this feature may be significantly older than those sampled up-valley. Legend on Figure 5.6. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

### *5.3.6 Watendlath*

Two pits were excavated on moraines in the vicinity of Watendlath Tarn, (260 m OD) which McDougall (2001) proposed to have been overridden by a glacier draining the area of Low Saddle during the LLS. Of four surface exposure dates (Figure 5.8) collected on boulders inside the proposed limit, three predated the LGM (Wilson et al., 2013), and therefore probably reflect nuclide inheritance. The final date fell between 15.0±0.8 and 15.6±1.0, suggesting a pre-LLS age for these features (Wilson et al., 2013). This site was sampled in order to determine whether the development of soil on the Watendlath moraines is comparable to that on the pre-LLS, chronologically constrained outer moraine sampled at Mosedale. If so, this would indicate that the Watendlath moraines predate the LLS and that the fourth cosmogenic isotope date is correct, whereas a less developed soil would support McDougall's (1998, 2001) interpretation of these features as LLS in age. Pits excavated on two moraines within this valley gave total soil depths of 15.5 and 19 cm and combined E and B horizon thicknesses of 12.5 and 19 cm. B horizon reddening was consistent between the two pits (CDE = 9) as were B horizon silt/clay percentages (60, 64 and 66%) and mass LOI (8-10%) (Appendix 2).

Table 5.6 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence at Watendlath. The table is organised so that the youngest soil appears at the top and the oldest at the bottom.

| Watendlath | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| WT2 | 12.5 | 10.0 | 15.5 | 9 | 3.77 |
| WT1 | 19.0 | 19.0 | 19.0 | 9 | 4.13 |
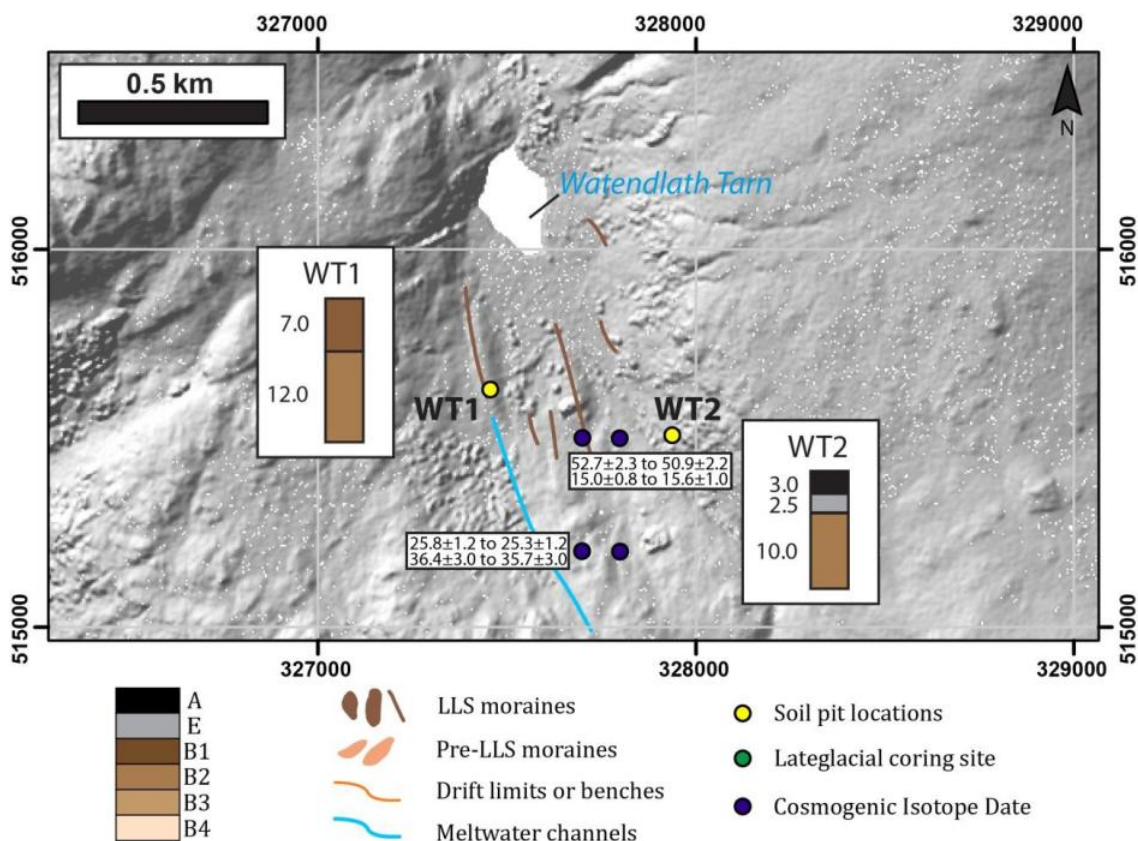
Figure 5.8 The soil chronosequence for Watendlath. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. Three of the four cosmogenic dates predate the LGM and the fourth dates to approximately 15-15.6 ka. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

### 5.3.7 Deepdale

Deepdale is an east-northeast facing valley in the Helvellyn range of the Lake District, the floor of which, at approximately 250 and 300 m OD, is occupied by a sequence of recessional moraines. Three moraines were sampled within Deepdale, all of which fall inside Sissons' (1980) reconstructed LLS limit. Although sampling on a moraine beyond this limit would have allowed comparison between the "inside" and "outside" pits, beyond DP3 the valley sides are characterised by gullied drift, rather than moraines. Thus, the aim of sampling this valley was to determine whether all of the moraines appear to belong to the same age population. The three soil profiles have consistent thicknesses of B horizons (between 10.5 and 9 cm) and total soil depths (13.5, 14.5 and 13 cm) (Figure 5.9). The consistency of these values does suggest that the three moraines belong to a single age population. Substantial thicknesses of oxidised parent material were observed at the base of the youngest and oldest pits but were absent from DP2. B horizon reddening decreased from a CDE of 10 at the youngest site, falling to 6 for the older two sites, whilst the maximum LOI percentage in the B horizons increased from 18% at DP1 to approximately 25% for DP2 and 3. However, these trends may simply reflect variations within the small

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sample set. Cumulative silt and clay sized particle percentages for the upper B horizons are broadly consistent, ranging from 77 to 66%, but the lower B horizon which comprises the majority of the B horizon in DP2 was as low as 35%. There is a range of between 13% and 26% for the LOI percentages of the B horizons throughout the pits. The consistency of soil depths and B horizon thicknesses suggests that these soils are all of similar ages.
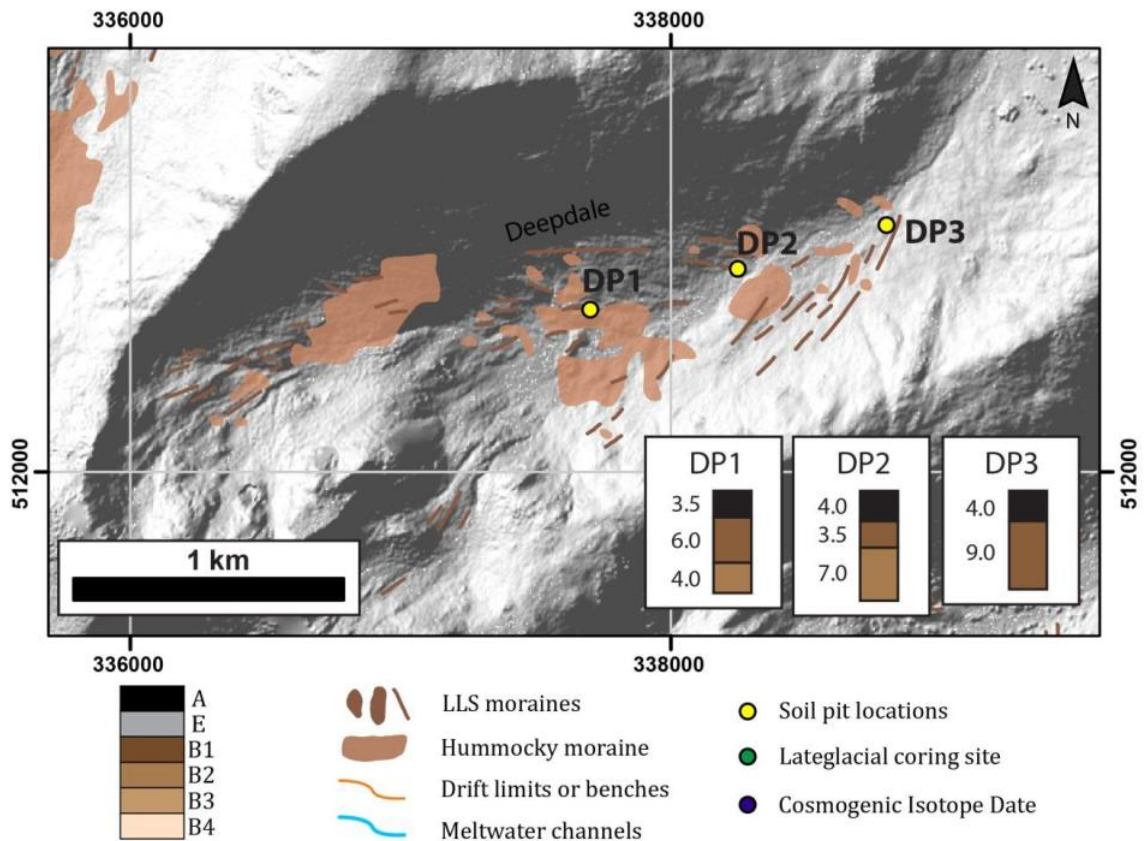


Figure 5.9 The soil chronosequence for Deepdale. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. This valley shows the most consistent soil depths in the study area with relatively thin horizons and total soil depth. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Table 5.7 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence in Deepdale. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Deepdale | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| DP1 | 10.0 | 10.0 | 13.5 | 10 | 1.99 |
| DP2 | 10.5 | 10.5 | 14.5 | 6 | 2.56 |
| DP3 | 9.0 | 9.0 | 13.0 | 6 | 3.13 |

## 5.3.8 Pasture Bottom

Pasture Bottom is a north-northwest facing valley in the eastern Lake District, the valley floor of which lies at approximately 250 m OD. The floor and lower valley sides are occupied by a sequence of well-developed, closely-spaced recessional moraine ridges, some of which bifurcate, and which appear to climb valley sides at an almost consistent angle (McDougall et al., 2015b) (Figures 5.10 and 5.11). Sissons (1980a) proposed that an alpine style glacier had occupied this valley, whereas McDougall (2013) argued that the valley glacier had been fed by ice from the plateau to the south. With respect to the terminal positions for the glacier, Sissons (1980a) deemed that the last large moraine ridge marked the maximum extent, whereas McDougall (2013) used a drift limit on the northeastern valley side, inferring that the glacier had been approximately 350 m longer at its maximum extent. Five pits were excavated on moraines within the valley (Figure 5.10), four within Sissons' (1980) limit and a fifth that was outside the limit proposed by Sissons (1980a) but inside that proposed by McDougall (2013). These sites were chosen in order to investigate whether one or two age populations appeared to be present within this valley. Total soil depths generally increase with distance down-valley, except for the fourth pit (PB4) which is slightly shallower than would be expected, owing to the absence of an E horizon. Likewise, with the exception of PB4, the thickness of the E and B horizons combined increases with depth. An increase in soil thickness is observed for the final (oldest) pit in the sequence (PB5), potentially indicating that soil development commenced on this moraine much earlier than it did on the other moraines sampled within the sequence and that this feature belongs to an older age population than those sampled up-valley. The CDE for B horizon reddening was at its maximum at PB3 (CDE = 12) and decreased for both PB2 and PB4 (CDE = 6). The percentages of silt and clay sized particles and of mass LOI increase from PB1 to PB4 but then decreased markedly for the much deeper pit of PB5.

Table 5.8 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence in Pasture Bottom. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Pasture Bottom | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| PB1 | 9.0 | 7.5 | 14.0 | 6 | 1.44 |
| PB2 | 11.0 | 8.0 | 16.0 | 6 | 1.55 |
| PB3 | 17.5 | 14.0 | 21.5 | 12 | 1.99 |
| PB4 | 14.0 | 14.0 | 19.0 | 8 | 2.18 |
| PB5 | 33.0 | 28.0 | 39.0 | 6 | 2.49 |
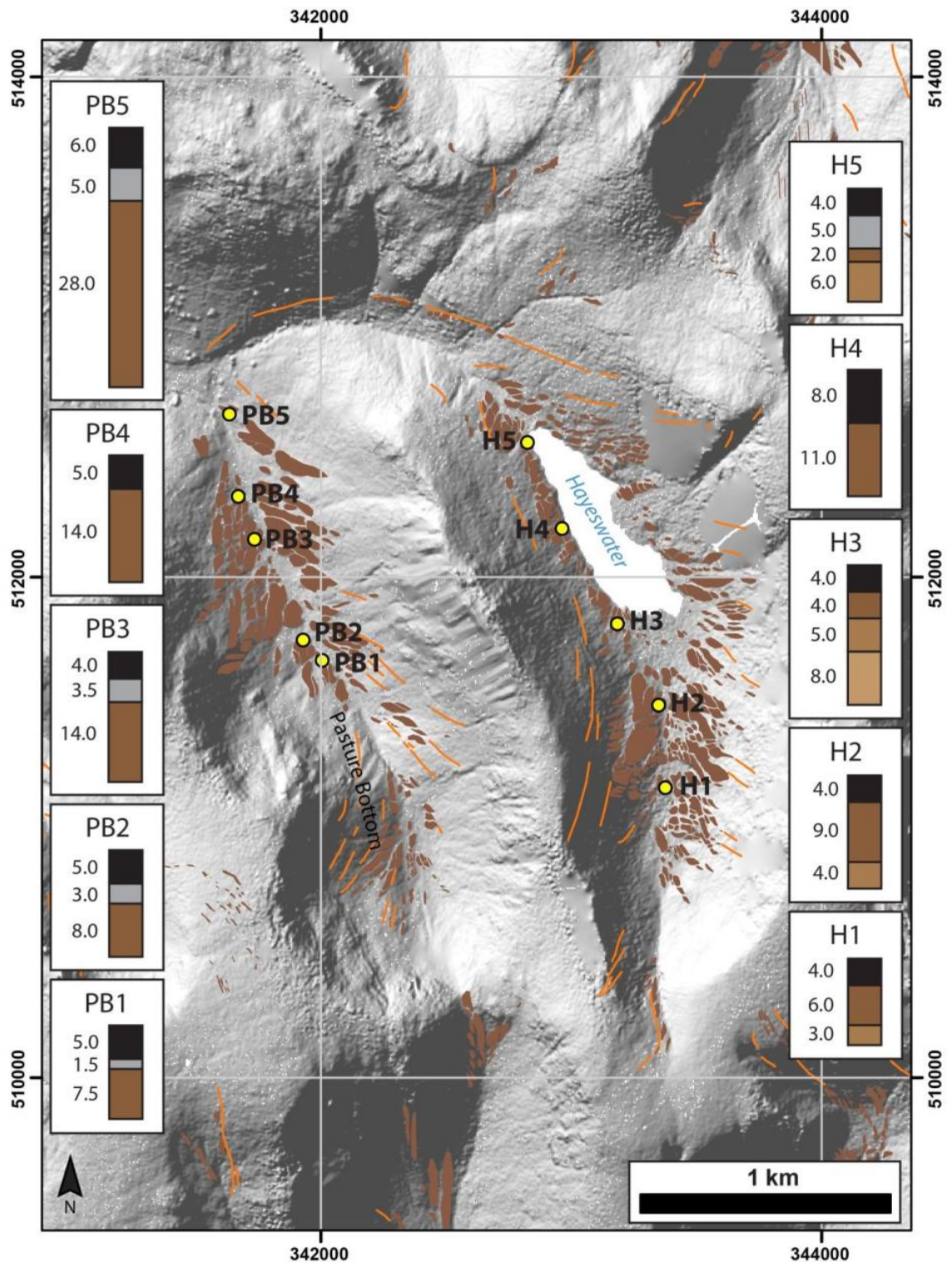
Figure 5.10 The soil chronosequence for Pasture Bottom and Hayeswater. For each of the soil profiles, 1 mm on the figure represents 1 cm of soil depth in the field and the thickness of each horizon (in cm) is noted beside the relevant layer. Soils are relatively thin in the upper valleys but show a dramatic increase on the outermost moraine in Pasture Bottom, suggesting that this moraine was formed during an earlier phase of glaciation. Legend on Figure 5.9. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies Inc. provided by the NERC Earth Observation Data Centre.

Figure 5.11 Moraines in Pasture Bottom, looking up-valley from the vicinity of PB3. Photograph from H.L. Bickerdike.

### *5.3.9 Hayeswater*

Hayeswater is adjacent to the valley of Pasture Bottom and both valleys coalesce in their lower reaches into the trunk valley that drains northwards into Ullswater. However, the lower reaches of the valley rise steeply from their confluence with Pasture Bottom, from about 200 m OD to 430 m OD, where the majority of the valley floor is occupied by Hayeswater Reservoir. Like Pasture Bottom, Hayeswater valley is occupied by a well-developed sequence of recessional moraines, from which it can be inferred that deglaciation was characterised by active, oscillatory retreat. The asymmetry of moraines and their continuation from the upper valley floor onto the plateau behind is compelling evidence that the glacier was fed by a plateau icefield (McDougall, 2013; McDougall et al., 2015b). Five soil pits were excavated in this valley, the oldest of which is situated at the down-valley end of the reservoir (Figure 5.10). Sampling down-valley of this point was restricted by the absence of sharp-crested moraine ridges. Steep valley sides in the lower valley may have prevented moraine deposition or slope processes may have removed evidence of any morainic features since deglaciation (McDougall et al., 2015b). Although moraine ridges are present for 300 m down-valley of the reservoir, these were more subdued in their morphology, and thus their soils would have been incompatible with the other results in the study. Sampling from one moraine was attempted but excavation was inhibited by numerous cobbles which had impeded the formation of clear soil horizons. Consequently, all five pits were located inside the limits proposed by Sissons (1980a) and McDougall (2013) and thus were sampled to determine whether these features belong to a

single age population. Total soil depths increase initially, from a minimum of 13 cm, to reach their maximum at H3 (21 cm) and then decrease slightly to 17 and 19 cm. This pattern is mirrored by the trend in combined E and B horizon thicknesses, although an E horizon was only present in H5. The consistent nature of the relatively shallow soils in this valley suggests that all features may belong to the same age group. B horizon reddening is consistent along the valley, with H4 having a CDE of 6 whilst the other pits display a CDE of 8. The LOI percentage decreases with depth in each pit but there is no trend between this and relative age. The relationship between percentage of silt and clay particles and depth within the soil pit is less clear than observed at other sites and no relationship with relative age is apparent.

Table 5.9 Combined E and B horizon thickness, total soil depths, maximum B horizon reddening and distance from the cirque backwall/ice-shed for the soil chronosequence in Hayeswater valley. The table is organised so that the youngest soils appear at the top and the oldest at the bottom.

| Hayeswater | E and B horizon thickness (cm) | B horizon thickness (cm) | Total soil depth (cm) | CDE | Distance from backwall/ice-shed (km) |
|---|---|---|---|---|---|
| H1 | 9.0 | 9.0 | 13.0 | 8 | 1.24 |
| H2 | 13.0 | 13.0 | 17.0 | 8 | 1.57 |
| H3 | 17.0 | 17.0 | 21.0 | 8 | 1.92 |
| H4 | 11.0 | 11.0 | 19.0 | 6 | 2.35 |
| H5 | 13.0 | 8.0 | 17.0 | 8 | 2.73 |

## 5.4 Discussion

### 5.4.1 Loss on ignition, particle size and B horizon reddening

The results appear to show that the use of soil chronosequences may have some potential to differentiate between LLS and pre-LLS moraines in the Lake District, but that distinctions between the two age groups are not as clear as have been observed in other studies (e.g. Birkeland, 1978; Evans, 1999). Both the percentage of mass LOI and the cumulative percentage of silt and clay sized particles decrease with increased depth in individual pits (see Appendix 2). However, the values of these properties for the B horizon fluctuate between pits, both within individual valleys and between valleys in the study area, rather than changing sequentially, and no distinct change between moraines of proposed LLS age and older features could be detected. This is perhaps unsurprising, as Birkeland (1978) also found that organic content and particle size were not useful in differentiating between Quaternary deposits on Baffin Island, Canada. Vanden Bygaart and Protz (1995) observed that although organic matter in soil profiles increased with age on sand dunes in Ontario, Canada, the most rapid increase was within the first 2 ka. Since the moraines sampled in the Lake District are, at their youngest, around 10 ka older than this,

it is likely that the rate of increase in organic matter is now much slower and that trends in this characteristic are obscured by local variations. Likewise, the maximum reddening of the B horizon, as measured by the CDE value, shows no down-valley trend (Figure 5.12) and, with the exceptions of Seathwaite and Deepdale, the $R^2$ values are very low (Table 5.10). In several valleys, such as Mosedale and Hayeswater, the CDE value remains constant throughout the valley, aside for a single anomaly in the middle of the sequence of moraines. Indeed, the CDE for MD1, the pit assigned a pre-LLS age based on its location down-valley of the Lateglacial coring site in Mosedale, is lower than for many of the features assigned to the LLS by Sissons (1980a) and McDougall (1998, 2001, 2013). It may be that the difference in moraine ages is insufficient to have allowed significant development of this characteristic or may be the case that the magnitude of local scale variability obscures any possible trends throughout valleys.
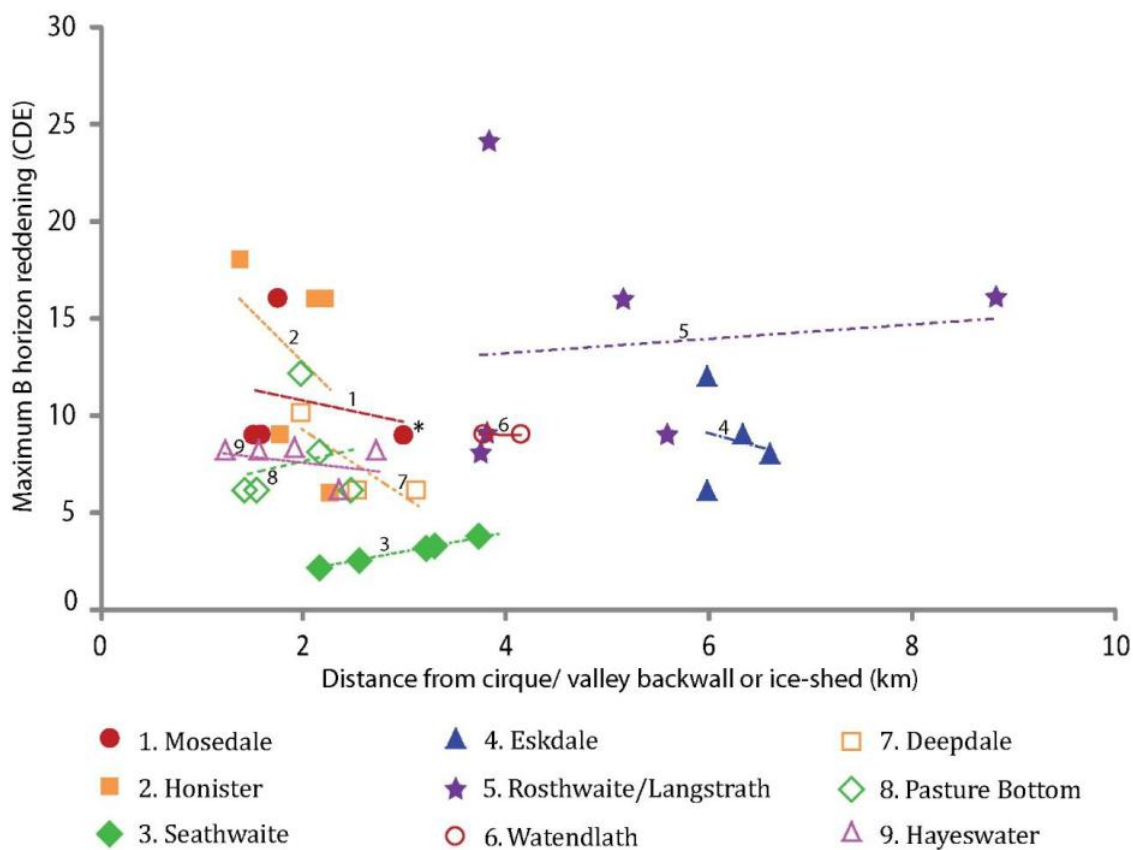


Figure 5.12 Maximum reddening (CDE) of the B horizon against distance from the cirque backwall or ice-shed (an assumed proxy for relative age). Linear trendlines show that CDEs appear unrelated to age and are consistent between pits in many of the valleys sampled. MD1, the pit assigned a pre-LLS age based on its location down-valley of a Lateglacial coring site, is marked with an asterisk.

Table 5.10 R² values for the relationship between maximum B horizon reddening (CDE) and distance from cirque/valley backwall or ice-shed.

| Site | Sample Size | R² |
|---|---|---|
| Mosedale | 4 | 0.044 |
| Honister | 5 | 0.144 |
| Seathwaite | 5 | 1.000 |
| Eskdale | 4 | 0.029 |
| Rosthwaite/ Langstrath | 6 | 0.014 |
| Watendlath | 2 | n/a |
| Deepdale | 3 | 0.749 |
| Pasture Bottom | 5 | 0.036 |
| Hayeswater | 5 | 0.136 |

### 5.4.2 Total soil depth and combined E and B horizon thickness

Total soil depth (Figure 5.13) and the combined thickness of the E and B horizons (Figure 5.14) do appear to relate to landform age, with this relationship being shown in seven of the nine valleys sampled. This is unsurprising, given the successful use of soil depth and B horizon thickness when differentiating between different age groupings of moraines (and other landform types) in various locations (e.g. Birkeland, 1978; Evans and Rogerson, 1986; Hall and Shroba, 1993; Evans, 1999). In several valleys (Mosedale, Pasture Bottom and Langstrath) there is a jump from relatively shallow soils in the upper valleys to much deeper soils on the outermost moraines. Soil depth and combined E and B horizon thickness appear closely related to distance from the cirque or valley backwall, which acts as a proxy for the age. This is reflected by relatively high R² values for these sites (Tables 5.11 and 5.12), although it should be noted that the strength of this relationship is perhaps surprising if it is assumed that the jump in soil thickness reflects the presence of two age populations of moraines within these valleys.

Within the total sample set there is a high proportion of sites with comparatively thin soils and horizons, clustered between 9 and 20 cm based upon E and B horizons combined, (72% of the total sample set), perhaps reflecting the predominance of young moraines in the study area (Figure 5.15). However, there is not a distinct break in combined horizon thickness between the cluster of apparently younger landforms and the outermost features in Mosedale, Pasture Bottom and Langstrath. Intermediate thicknesses of soil are present in Mosedale, Honister, Seathwaite and Eskdale. Even if the deepest soils are assumed to predate the LLS, in the absence of chronological control it is not possible to determine where in the range of intermediate thicknesses the threshold value between LLS and older landforms would lie. Likewise, if the samples could be split into two age

groups it would be possible to test whether there is a statistical difference between the thicker and thinner soils but, given the intermediate values of soil thickness and lack of robust chronological control, this is not possible.
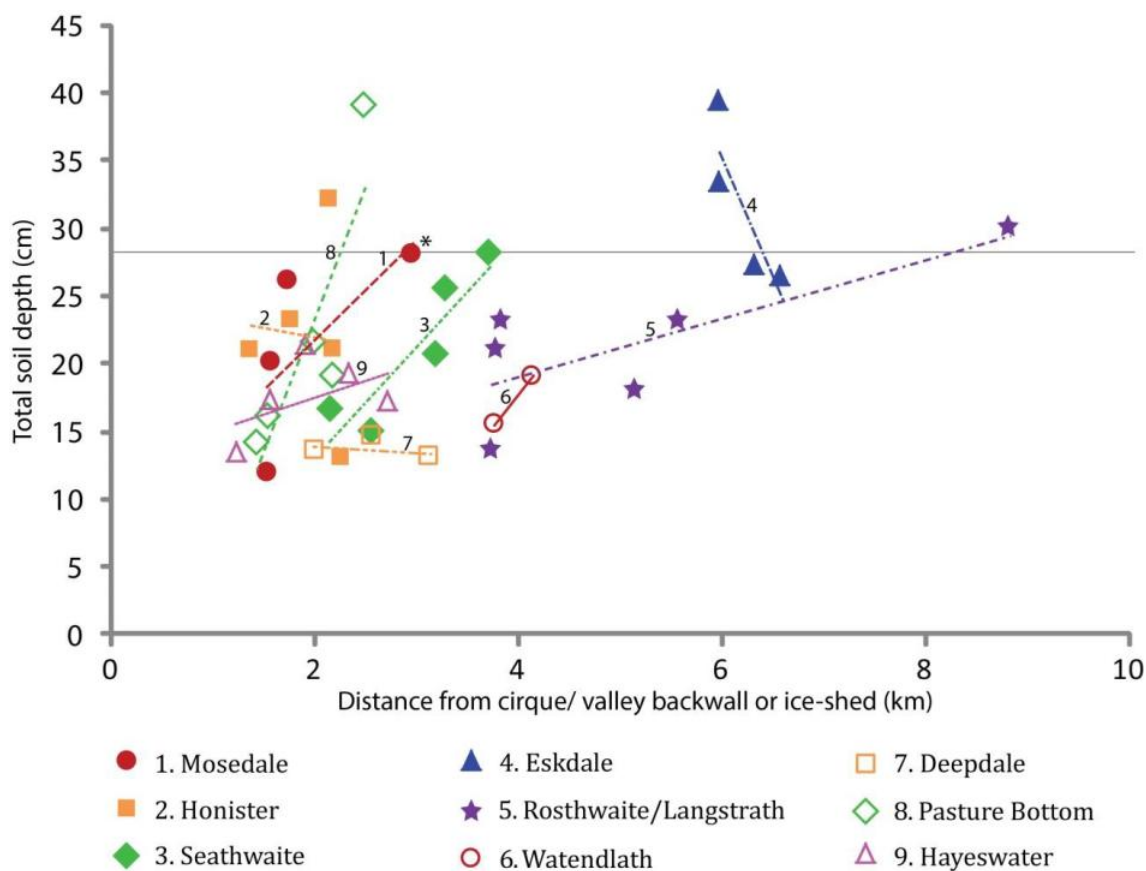


Figure 5.13 Total soil depth against distance from the cirque backwall or ice-shed. Linear trendlines show that total soil depth generally increases with distance from the backwall but there are some exceptions where the pits sampled are closely spaced along the former glacier flowline. The horizontal grey line represents the total soil thickness at site MD1, which has been assigned a pre-LLS age based on its location down-valley of a Lateglacial coring site.

Table 5.11 $R^2$ values for the relationship between total soil depth and distance from cirque/valley backwall or ice-shed.

| Site | Sample Size | $R^2$ |
|---|---|---|
| Mosedale | 4 | 0.487 |
| Honister | 5 | 0.006 |
| Seathwaite | 5 | 0.832 |
| Eskdale | 4 | 0.764 |
| Rosthwaite/ Langstrath | 6 | 0.593 |
| Watendlath | 2 | 1.000 |
| Deepdale | 3 | 0.108 |
| Pasture Bottom | 5 | 0.737 |
| Hayeswater | 5 | 0.245 |

Within the sample set there are some anomalies. Eskdale is notable on Figure 5.14, as the combined E and B thicknesses decrease with distance down-valley. This can largely be attributed to the thick horizons which have developed on the second youngest (second in the sequence down-valley) moraine and the fact that the pits are closely spaced along only a short part of the flow path of the former glacier. In Deepdale, E and B horizon thickness decreases very slightly on the final moraine in the sequence but this value is only 1.5 cm lower than the maximum thickness on the other moraines. This may again reflect the relatively close spacing of the moraines or may indicate that deglaciation was quite rapid between the sites of the pits, allowing little time for soil formation on each moraine before the next was exposed by the retreating ice.
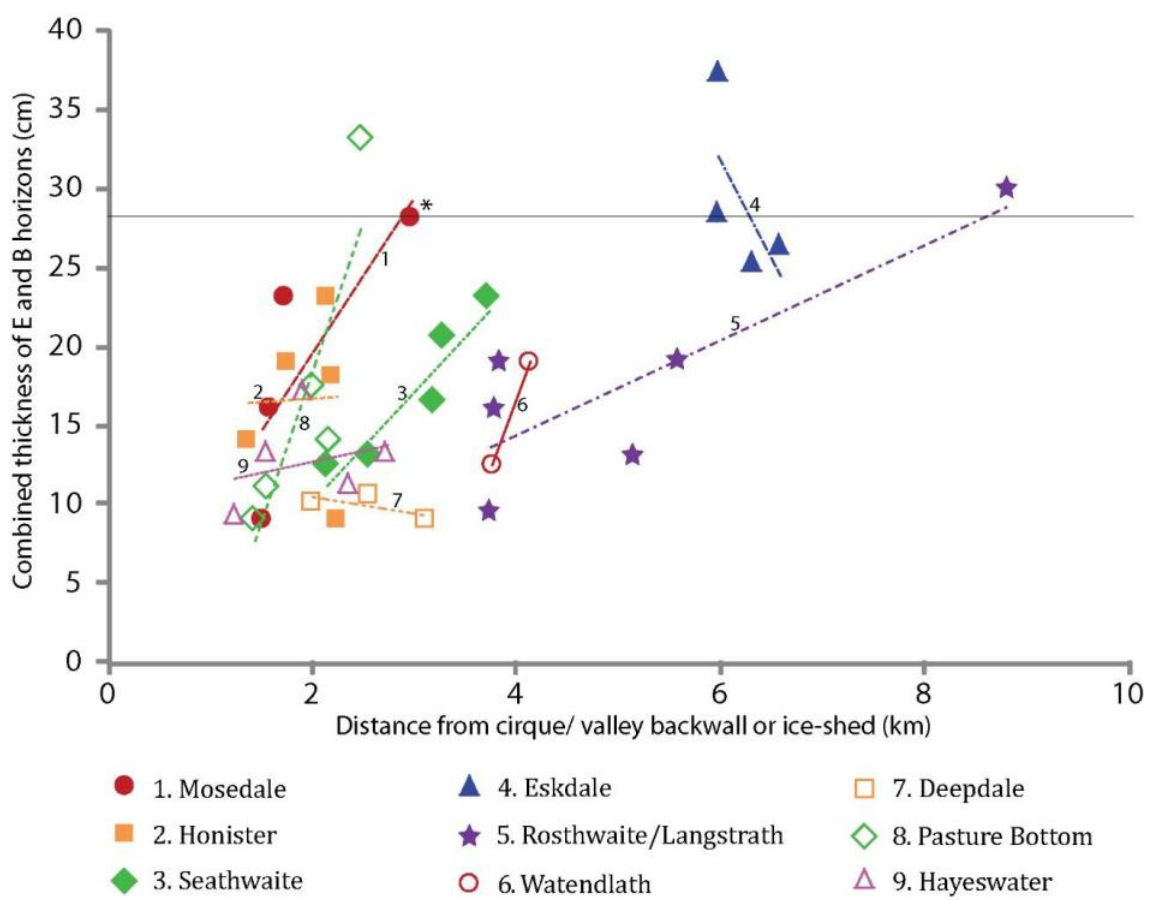


Figure 5.14 The combined E and B horizon thickness against distance from the cirque wall or ice-shed. Linear trendlines show that, for the majority of sites, there is a general increase in horizon thickness with increased distance down-valley (a proxy for relative age). The horizontal grey line represents the combined E and B horizon thickness at site MD1, which has been assigned a pre-LLS age based on its location down-valley of a Lateglacial coring site.

Table 5.12 R² values for the relationship between combined E and B horizon thickness and distance from cirque/valley backwall or ice-shed.

| Site | Sample Size | $R^2$ |
|---|---|---|
| Mosedale | 4 | 0.650 |
| Honister | 5 | 0.001 |
| Seathwaite | 5 | 0.893 |
| Eskdale | 4 | 0.438 |
| Rosthwaite/ Langstrath | 6 | 0.705 |
| Watendlath | 2 | 1.000 |
| Deepdale | 3 | 0.430 |
| Pasture Bottom | 5 | 0.751 |
| Hayeswater | 5 | 0.075 |
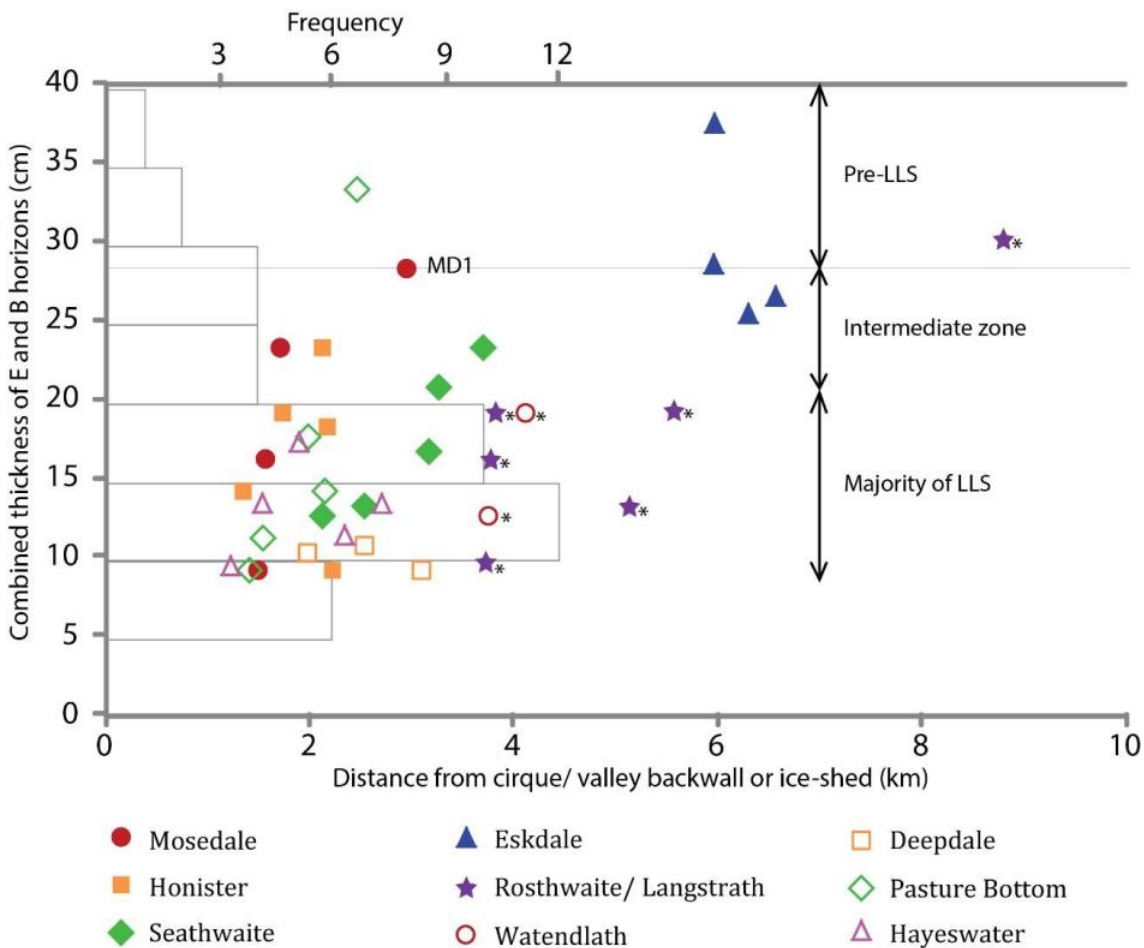


Figure 5.15 Summary figure of the chronosequence data for combined E and B horizon thickness. The frequency of combined E and B horizon thicknesses shows a clustering of relatively thin soils between 9 and 20 cm, thought to represent LLS aged landforms. There is then a zone of intermediate soil thicknesses. Finally, there is a group of the thickest soils, including the chronologically constrained pre-LLS moraine MD1, the inner moraines at Eskdale, the outermost moraine at Pasture Bottom and the Rosthwaite moraine. Asterisks indicate where soil pits fall in the vicinity of pre-LLS cosmogenic isotope dates, although it is probable that at least some of these dates have been affected by nuclide inheritance and, thus, may not be reliable.

### 5.4.3 Chronosequence derived ages for the Lake District moraines

The clear lithostratigraphic and palynological evidence for Lateglacial sediments in the snow avalanche impact pit in Mosedale (Evans et al., 2015), up-valley of site MD1 (Figure 5.1), indicates that this moraine predates the LLS. Both the total depth of soil (from the top of the A horizon to the base of the B horizon) and the B horizon thickness of this pit measured 28 cm (as indicated on Figures 5.13, 5.14 and 5.15). Given the strong stratigraphic evidence that this moraine predates the LLS, this site can be used to calibrate the chronosequence and it is reasonable to infer that other moraines within the study area exhibiting a similar total depth of soil, must be of a similar age (i.e. predate the LLS). When total soil depth is considered, this would include several moraines the locations of which are difficult to reconcile with a pre-LLS age, including the oldest moraine in Seathwaite, which lies within the proposed LLS limits of both Sissons (1980a) and McDougall (2001) (Figure 5.1), and a moraine mid-way along the sequence at Honister Pass (HON3).

Rather than examining total soil depth, it may be more robust to deduct the thickness of the A horizon to leave a value for combined E and B horizon thickness only to get an indicator of relative age (Figures 5.14 and 5.15). Usage of this property indicates that only the outer moraines at Rosthwaite, Pasture Bottom and the inner moraines at Eskdale are of comparable age to the Mosedale outer moraine (Figures 5.14 and 5.15). For example, for the outermost moraine sampled in Pasture Bottom (PB5) both E and B horizon thickness (33 cm) and total soil depth (39 cm) exceed the values for the Lateglacial moraine in Mosedale (28 cm). Similarly, in Eskdale, the younger inner moraines have combined E and B horizons of 28 and 37 cm, indicating that they are at least as old as the Mosedale outer moraine. The depths of soil on these inner moraines are thus probably more reflective of the true age of all of the moraines sampled in Eskdale as it is impossible to reconcile their position with having been exposed before the moraines sampled a short distance further down-valley. A pre-LLS age for the Rosthwaite moraine (as indicated on Figure 5.15), at the foot of Langstrath, can be reconciled with the youngest exposure date collected at this site by Wilson et al. (2013), although, as discussed above, the reliability of this single age is questionable. If the Rosthwaite moraine belongs to a different age population to those in upper Langstrath, this might account for the absence of recessional moraines between the outer limit and the onset of the closely-spaced recessional moraines further up-valley, which were also found to have shallower soils developed on their crests. However, dates from within the neighbouring lobe at Watendlath, as reconstructed by McDougall (2001), indicate that glaciation here occurred prior to the LLS (Wilson et al., 2013), yet both pits excavated at this site were much shallower than that on the Rosthwaite moraine. It is possible that the area of the Rosthwaite moraine (approximately

100 m OD) became ice-free earlier than at Watendlath, which is at a much higher elevation (270 m OD), which would explain the difference in soil thicknesses between the two sites. Although cosmogenic isotope dates indicate that several other moraines, in addition to those at Eskdale, Pasture Bottom and Mosedale, predate the LLS (as indicated by asterisks on Figure 5.15), the reliability of these dates is questionable (as discussed above in Section 5.3.5 and 5.3.6) and therefore these dates cannot be used to calibrate the chronosequence.

With the exception of the moraines displaying substantial soil development, the majority (72%) of combined E and B horizon thicknesses are clustered between 9 and 20 cm (Figures 5.14 and 5.15). This cluster of values appears to represent one age population of moraines. For some valleys, such as Deepdale and Hayeswater, all moraines belong to this apparent age population and fall inside the LLS limits of both Sissons (1980a) and McDougall (2013), although no LLS chronological control is available for these sites. In other valleys, such as Seathwaite and Honister, only those moraines further up-valley fall into this category. However, the break between these values and those for the deeper soils (greater than or equal to 28 cm) is not clear cut, making it impossible to determine where in the intermediate values the threshold between LLS and pre-LLS lies.

Moraines which gave intermediate values of soil development include MD2 (Mosedale), which has a total depth of 26 cm and a combined E and B horizon thickness of 23 cm, which lies just beyond the LLS glacier limits proposed by Brown et al. (2013) and Sissons (1980a). A similar depth and thickness of soil was observed on the outermost moraine at Seathwaite, and at HON3 at Honister Pass. Whilst it is possible that these moraines predate the LLS, their positions inside proposed LLS limits is hard to reconcile with an older age and it is perhaps more plausible that the soil profile from MD2 is underdeveloped for its true age. The presence of recessional moraines beyond the limits of the LLS glaciers reconstructed in several valleys, including Eskdale and Wasdale (Brown et al., 2011, 2013), suggests that Late Devensian Ice Sheet deglaciation was interrupted by numerous readvances and stillstands. Cosmogenic isotope dating from Lingmell Col suggests that this area was exposed by downwasting of the Late Devensian ice by approximately 17.8 ka, but a date of 14.2±1.7 ka to 13.5±1.4 ka midway along the northwestern shore of Wastwater (Figure 5.1), indicates that deglaciation of the valleys may have been much later (Ballantyne et al., 2009). Consequently, the difference in time that elapsed between the exposure of pre-LLS and LLS moraines is relatively short and consequently the transition from one age population to the other, as reflected by the intermediate values of soil development (Figure 5.15), is more gradual than abrupt. The moraines sampled for this study do appear to broadly fall into two age populations.

However, the lack of a distinct break between the two groups and the reliance on the single chronological control site to calibrate the chronosequence, mean that this study has been unable to conclusively distinguish LLS from pre-LLS landforms in the Lake District.

This limitation of the technique is perhaps to be expected, given the potentially narrow window of time for soil formation between pre-LLS deglaciation and the onset of LLS glaciation. It also explains fluctuations in soil thickness between moraines of the same age population. Soil development rates calculated for other regions demonstrate that, after an initial period of rapid soil development, the rate of development slows. For instance, Evans (1999) calculated that at Jardalen, Norway, initially the soil depth increased at a rate of 5 cm per 100 years, but it then slowed dramatically after 700 years to a maximum of 0.06 cm per 100 years. A similar pattern in soil development, albeit over a longer timescale, was identified by Bentley et al. (2007), using cosmogenic isotope dates on moraines in South Georgia. A decrease in the rate of soil formation on landforms of older ages, such as those sampled in this study, would lead to comparably small differences in the depth of soil that would be expected to form on moraine crests, meaning that local variations between pits might obscure the pattern of depth change with age. A lack of fine age resolution when using soil chronosequences was recognised by Berry (1994) in the Sierra Nevada, California, and was attributed to a combination of slow soil formation rates, moraine crest erosion, and (although less relevant to this study) burial of soil at the foot of moraines by colluvium and low volumes of atmospheric dust necessary for soil development in semi-arid climates.

### 5.4.4 Future work

The limitations in the resolution of the soil chronosequence technique highlight the necessity for establishing reliable chronological constraint on moraines within the Lake District, which would then allow more confident threshold values of soil development between different age populations to be identified. Establishment of independent chronologies on moraines at other locations, for instance by using lichenometric data, has allowed soil development rates to be calculated (Messer, 1988; Evans, 1999), which, if available for the Lake District, could be instrumental in confidently identifying the ages of moraines with intermediate soil depths and combined E and B horizon thicknesses. Choice of sites in this study was largely dictated by the necessity to sample only on sharp-crested moraines with well-drained soils. There are a number of other sites within the Lake District where the age of moraines of more subdued relief remains uncertain, such as Woundale, Troutbeck Valley and Kent Valley in the eastern Lake District (McDougall, 2013). Given the low relief of these features, comparison of soil development on these moraines would not be reliably comparable to the chronosequences constructed on the

sharp-crested features and thus would need to be part of a much larger dataset of more pits, both within and between valleys. Further work to extend the distance over which sampling occurred within each valley would contribute to this dataset but might also remove uncertainty in sites such as Eskdale where the trend in soil thickness with age appears unreliable. Future application of this method within the study area must therefore be conducted in tandem with implementation of rigorous and systematic dating of landforms to establish reliable and independent chronological control.

## 5.5 Conclusions

- Soils within nine valleys in the English Lake District have been sampled in a pilot study that tests whether the soil chronosequence method can differentiate between LLS and pre-LLS aged moraines.

- Of the soil properties measured, percentage LOI and silt/clay particle percentages were found to decrease with increased depth in individual pits. However, in none of the valleys sampled was there a relationship between these characteristics and relative age. The maximum soil reddening of the B horizon either fluctuated unpredictably or remained consistent within valleys and no sequential relationship with age was identified.

- Total soil depth and, particularly, combined E and B horizon thickness generally increases with distance from the cirque or valley backwall, which is a proxy for landform age. However, the small sample size means that this relationship is relatively weak. Exceptions to this rule occur only where sampling took place over a short section of the flow path of the former glacier.

- The majority of moraines sampled (72%) have combined E and B horizons clustered between 9 and 20 cm in thickness, representing a younger age population of moraines, inferred to be of LLS age. A second population of older moraines with thicker combined E and B horizons (28 cm or more) also features within the sample. The latter features are thought to predate the LLS age because their combined E and B horizons are thicker than those on the outer moraine at Mosedale, which has been assigned a pre-LLS age from a nearby Lateglacial coring site.

- Intermediate values of horizon thickness are present between these two populations and may reflect the short duration of time for soil formation between exposure by retreating LGM ice and exposure during early LLS retreat.

- The very limited chronological constraint makes it impossible to determine a threshold value between the two apparent age populations within this range of

intermediate values. Thus, this pilot study has been unable to conclusively distinguish between LLS and pre-LLS landforms across the English Lake District but might have the potential to do so if improved chronological constraint is developed.

# Chapter 6
# Conclusions and Implications

## 6.1 Key conclusions

This thesis has drawn together the published evidence for the geomorphology of the LLS in Britain in order to build a coherent picture of the extent, style and dynamics of glaciation during the stadial. Secondly, a pilot study to test whether soil chronosequences could be used to provide chronological control on landforms formed by LLS glaciers versus those formed by deglaciation of the Late Devensian Ice Sheet has been conducted. This conclusion addresses the research questions posed in Section 1.2 and briefly highlights future work that might fruitfully build on the findings of this study.

### 6.1.1 What is the range of evidence that has been used to map the LLS glaciation in Britain?

A key contribution of this thesis is a map and GIS database compiled of LLS glacial geomorphology from the published literature (Chapter 2). Synthesis of this evidence into the GIS database has allowed identification of the range of landforms used to map the LLS across Britain. This evidence includes moraines, drift and boulder limits, drift benches, periglacial trimlines, meltwater channels, eskers, striations and roches moutonneés, protalus ramparts and ice-dammed lakes. Of the 95,000 individual features, the majority are moraine mounds or ridges, which are found at the termini and within the limits of the majority of LLS glaciers. However, the nature of the evidence mapped in the published literature has varied between regions. In many of the upland areas previously occupied by mountain icefields, particularly in the western Scottish Highlands and islands, the geomorphology is dominated by extensive sequences of recessional hummocky moraines. Geomorphological mapping of the northern sector of the West Highland Glacier Complex (Bennett and Boulton, 1993a) focused predominantly on recessional moraines and included no evidence (such as trimlines) from which the vertical extent of these glaciers could be inferred. This contrasts with the landform assemblage observed in some areas of the central Highlands, where ice-marginal meltwater channels are well-developed on upland plateau areas. Glaciofluvial features are much rarer and, although there are examples of eskers and outwash plains found inside and at the margins of the LLS glaciers, these features are comparatively rare and indicate that glacial and proglacial fluvial systems were generally poorly-coupled. Compilation of this map highlights the previously fragmented and spatially inconsistent nature of LLS glacial geomorphological mapping in Britain and highlights gaps within the current mapping, most conspicuously, the lack of mapping for the southwest Scottish Highlands.

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### 6.1.2 How robust are the previously mapped limits of LLS glaciation, both spatially and temporally?

Compilation of the LLS glacial map and database, as detailed in Chapter 2, required a comprehensive and systematic review of the existing published literature (Chapter 3). Critical assessment of the evidence from sub-regions in Scotland, England and Wales was important in deciding which material was suitable for inclusion in the GIS database, particularly at localities where conflicting interpretations exist. As a result of the review it is clear that, although there is a wealth of published literature concerning the LLS in Britain, there is still a large degree of uncertainty regarding the limits and timing of the maximum extents of these glaciers. The relative paucity of absolute dating on LLS features is particularly problematic and has inhibited both differentiation between LLS and older landforms, and identification of the timing and synchroneity with which the LLS glaciers reached their maximum extent. In some cases, even where dating has been undertaken, problems associated with techniques have led to the reliability of chronological evidence being questioned (Wilson et al., 2013; Boston et al., 2015; Small and Fabel, 2016). The ambiguous genesis of some specific landforms has also led to uncertainty when identifying the limits of LLS glaciation, such as in the Brecon Beacons. However, this problem is associated with cirque glaciers only and, thus, relates to a relatively small area of the total extent of LLS glaciation.

In some areas, the limits of LLS glaciation are well-constrained both spatially, by clear landforms which have been the subject of high resolution geomorphological mapping, and temporally, by robust absolute dating at the former glacier termini. The termini of the Lomond, Menteith and Callander glaciers are examples of sites where there is a high degree of confidence in both the timing and extent of LLS glaciation. In many locations, the extent of glaciation has been reconstructed from high resolution mapping but chronological evidence indicating a LLS age is only present within specific valleys and has been interpolated to a whole icefield. Examples of this include the icefields which occupied the Ben Hee and Beinn Dearg upland areas and the isles of Skye and Mull. More speculative limits exist along the western margins of the West Highland Glacier Complex, particularly where these glaciers terminated beyond the current coastline, and in the English Lake District. The southern sector of the West Highland Glacier Complex is, in many places, unmapped and undated and therefore, is very poorly constrained. Possible future research to directly address some of these areas of uncertainty has been proposed within Section 3.5.

### 6.1.3 What was the style of LLS glaciation and what factors influenced this?

Examination of the GIS database has aided identification of five LLS glacial landsystems: the cirque/niche glacier landsystem; the alpine icefield landsystem; the lowland piedmont lobe landsystem; the plateau icefield landsystem and the ice cap landsystem (Chapter 4). The cirque glacier landsystem was typical of localities where conditions were only just above the threshold required for glaciation. Alpine icefields were found in upland areas of sharp mountain crests and steep slopes and, in a limited number of cases, fed lowland piedmont lobe glaciers. Conversely, plateau icefields developed on upland areas of gently undulating summits, with thin, cold-based ice on the plateau surfaces feeding warm-based glaciers in the surrounding valleys. The ice cap landsystem was present in the vicinity of Rannoch Moor, at the centre of the West Highland Glacier Complex. Three styles of retreat have been inferred from the moraine record: active retreat throughout deglaciation, which seems to have been most widespread during the LLS; two-phase retreat; and uninterrupted retreat, which was restricted to a limited number of small cirque glaciers. Mapping of the spatial distribution of each landsystem suggests that the extent, style and retreat dynamics of the LLS glaciers were largely influenced by a combination of pre-existing topography and palaeoclimate. Glaciation was facilitated where topographic conditions aided the accumulation of snow, such as the mountains which surround Rannoch Moor or topographic hollows downwind of upland plateau areas. Superimposed onto this is the regional climatic effect of precipitation gradients across Britain, as discussed in Chapter 3, which contributed to an arid climate in the central Scottish Highlands and restricted glaciation.

### 6.1.4 Can soil chronosequences be used to resolve some of the current chronological uncertainty associated with LLS ice limits?

As highlighted throughout this thesis, chronological control on LLS landforms is relatively scarce. This has led to uncertainty in both assessing the timing and synchroneity with which LLS glaciers reached their maximum extents and differentiating between LLS and older landforms. The soil chronosequence approach was piloted in the English Lake District (Chapter 5) to establish whether it could prove an alternative to other dating techniques, which have been problematic in this region. Although the soil characteristics of B horizon reddening, loss on ignition and particle size did not show any relationship with relative age, total soil depth and, in particular, combined E and B horizon thickness do appear to increase with age. Combined E and B horizon thicknesses appear to show a cluster in values between 9 and 20 cm, thought to indicate moraines of LLS age, whilst calibration using existing chronological control in Mosedale suggests that pre-LLS moraines are present at four sites within the study area. However, there are also

intermediate values of E and B horizon thickness. The lack of a distinct break between the two populations, coupled with the reliance on the single chronological control site to calibrate the chronosequence, inhibits identification of the threshold between the two populations and means that currently the technique cannot conclusively distinguish between LLS and pre-LLS landforms in the English Lake District.

## 6.2 Implications and further work

A key aim of this thesis was to bring together the published geomorphological evidence of LLS glaciation in Britain to develop a more holistic understanding of the extent, style and dynamics of glaciation during the stadial. This marks the first time that this has been undertaken for these glaciers at a regional scale, incorporating evidence from Scotland, England and Wales. This project was partly inspired by compilation of evidence associated with the last British-Irish Ice Sheet (Clark et al., 2004; Evans et al., 2005), which stimulated research to remap the ice sheet bed and to identify the pattern and timing of ice sheet retreat. It is hoped that, in the same way, this project will prompt renewed and systematic research on the LLS glaciers in Britain. The geomorphological and ice extent shapefiles were disseminated alongside the papers which comprise Chapters 2 and 3, enabling researchers to use this data directly in future projects, e.g. to remap areas of uncertain or poorly recorded geomorphological evidence or in numerical modelling experiments. The landsystem models are a key outcome of this thesis and provide a regional framework for the style of LLS glaciation in Britain that could be used to test whether the style of LLS glaciation in other areas (e.g. North America (Lowell et al., 1999), the European Alps (Ivy-Ochs et al, 2009), and Scandinavia (Andersen et al., 1995)) was of a similar or different style to that in Britain. As such the following recommendations for future research are made.

### 6.2.1 Remapping of speculative areas

As highlighted in both Chapters 2 and 3, it is clear from the GIS database and glacial map that there is a large area in the southwestern Scottish Highlands (indicated by the red dashed line in Figure 3.30) that was glaciated during the LLS but for which high resolution geomorphological mapping is almost entirely absent. Likewise, glacial geomorphology of the eastern outlet glaciers of the West Highland Glacier Complex and of a possible icefield in the northern Tweedsmuir Hills is poorly or only partially recorded. Consequently, the limit of LLS glaciation is speculative in these areas and will remain so until the geomorphology of these areas is more completely mapped.

### 6.2.2 Bathymetric surveying and dating

LLS glaciers terminated beyond the present-day coastline in several of the lochs along Scotland's west coast. Bathymetric surveying in Loch Ainort, on the Isle of Skye, and Loch Hourn, on the Scottish mainland, has indicated that submarine moraines are present in these lochs beyond the LLS limit inferred from the onshore evidence (Dix and Duck, 2000; McIntyre et al., 2011). This raises the possibility that LLS ice may have extended further than previously reconstructed in the sea lochs and, therefore, bathymetric surveying is needed to identify and map possible LLS moraines at such sites. If such features are present, absolute dates on these landforms would both confirm whether they represent LLS glaciation and, if so, could provide valuable chronological control on the timing of the maximum extent and retreat of LLS glaciation along the west coast.

### 6.2.3 Constraint of the vertical ice limits and refinement of the LLS ice extent

The LLS glacier limit presented in Chapter 3 has been derived from published reconstructions (e.g. Benn et al., 1992; Ballantyne, 2002a; Finlayson et al., 2011; McDougall, 2013) which were adapted to accord with the glacial geomorphology compiled in the GIS database (Chapter 2). For the West Highland Glacier Complex, no attempt has been made to incorporate the vertical limits of this ice mass, as these are poorly constrained. Particularly in the northern sector, it would be helpful if future mapping included geomorphological evidence, such as periglacial trimlines, from which the maximum elevation of the icefield can be inferred. Establishing the vertical ice limits would enable the distribution of landsystems along the West Highland Glacier Complex to be identified with more precision than was possible in this thesis from the presently available data (Chapter 4).

### 6.2.4 Updating of palaeoclimatic inferences

Although the results of numerical modelling suggest that the LLS glaciers formed under steep northwards and eastwards precipitation gradients (Golledge et al., 2008), the notion that the LLS climate was more seasonal than previously recognised, suggests that precipitation may have been overestimated in calculations based on glacial reconstructions. In light of this, such palaeoclimatic estimates should be re-examined using the temperature-precipitation equation proposed by Golledge et al. (2010). These results could then be used to scale LLS climate records which drive future numerical modelling.

### 6.2.5 Comparison of retreat dynamics and rates

As identified in Chapter 4, the most widespread style of retreat for the LLS glaciers appears to have involved multiple minor readvances and stillstands as the glaciers

responded to even modest fluctuations in climate (Chapter 4). In the northwest Scottish Highlands, sequences of recessional moraines formed by this style of glaciation have been used to estimate the frequency of moraine formation using the barcode method (c.f. Lukas and Benn, 2006). Application of this methodology to other localities (e.g. the Outer Hebrides, Skye, Mull, the West Drumochter Hills and Monadhliath Mountains) where such sequences of moraines have been identified could provide a quantitative estimate of the retreat of LLS glaciers under a range of topographic and climatic conditions.

### 6.2.6 Compilation of existing LLS dates and implementation of a systematic dating regime

Compilation of dates relating to the build-up and retreat of the last British-Irish Ice Sheet into a GIS database (Hughes et al., 2011) has previously been used to create time-slice diagrams of deglaciation stages of that ice sheet (Clark et al., 2012). Similar analysis has been undertaken for the last Eurasian ice sheets (Hughes et al., 2016), the Laurentide and Innuitian ice sheets (Dyke et al., 2002) and in Antarctica, albeit at a much coarser temporal resolution (Bentley et al., 2014). Analysis of this kind for the LLS glaciers is currently inhibited by the general paucity of dates on LLS landforms and the large errors associated with absolute dating methods in relation to the short duration of the LLS. Therefore, there is a need for a systematic regime of dating on LLS landforms, which would not only allow confident identification of the limits of LLS glaciation in uncertain areas, but would also provide dates to calibrate relative dating methods, such as soil chronosequences. Dating would also shed light on the synchroneity with which the LLS glaciers reached their maximum extents. Furthermore, as absolute dating techniques improve and the errors on these dates decrease, it may be possible to date the retreat of LLS glaciers along former major retreat corridors, providing an important constraint for numerical models of ice retreat.

### 6.2.7 Comparison of the timing and style of British LLS glaciation with other regions

Developing a more coherent picture of the extent, style and timing of LLS glaciation will allow a better understanding of how British glaciation fits within the global context of the Younger Dryas (Alley, 2000; Carlson, 2013). For example, by constraining the timing of the maximum extent of LLS glaciation, it will be possible to determine whether the British LLS glaciers responded rapidly to changes in global climate and whether glaciers in different regions reacted synchronously to these changes. The landsystem models proposed in Chapter 4 can be used to test whether the signature of LLS glaciation is the same in different regions glaciated during the stadial around the North Atlantic. This could potentially provide information on the degree to which the same factors, such as

topography and palaeoclimate, influenced glaciation elsewhere. Ultimately, the LLS (Younger Dryas) provides an important analogue for rapid climate change from which we can learn much about the complex relationships between the cryosphere and the global ocean-climate system.

# Appendix 1

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# Appendix 2

**Soil chronosequence data: loss on ignition and silt and clay sized particles with depth.**

Each column shows data from one site and is organised so that the youngest pits are at the top and the oldest at the bottom. Shown for A, E and B horizons, including sub-layers.



- ● Loss on ignition
- ● Silt and clay sized particles

Loss on ignition

Silt and clay sized particles

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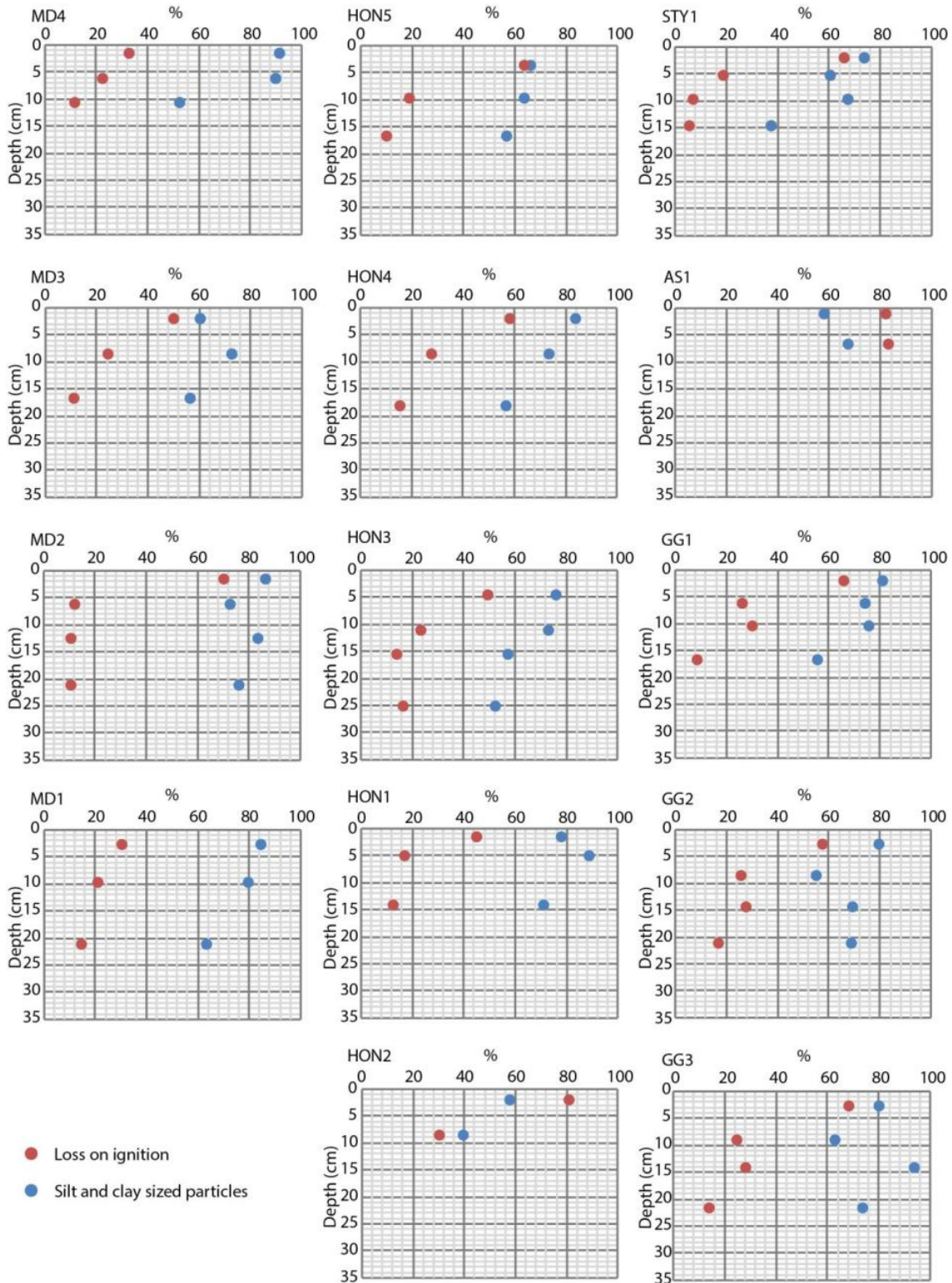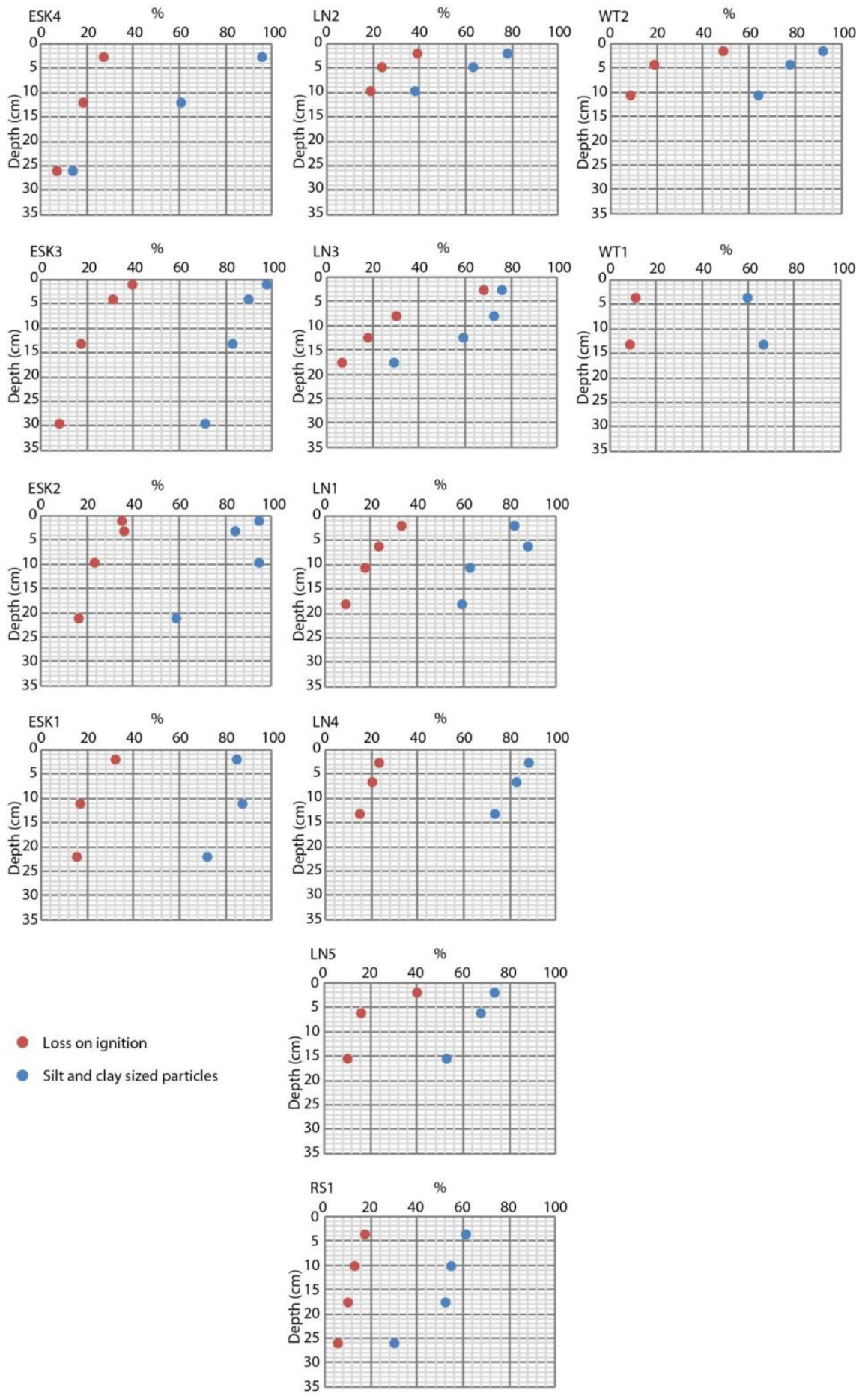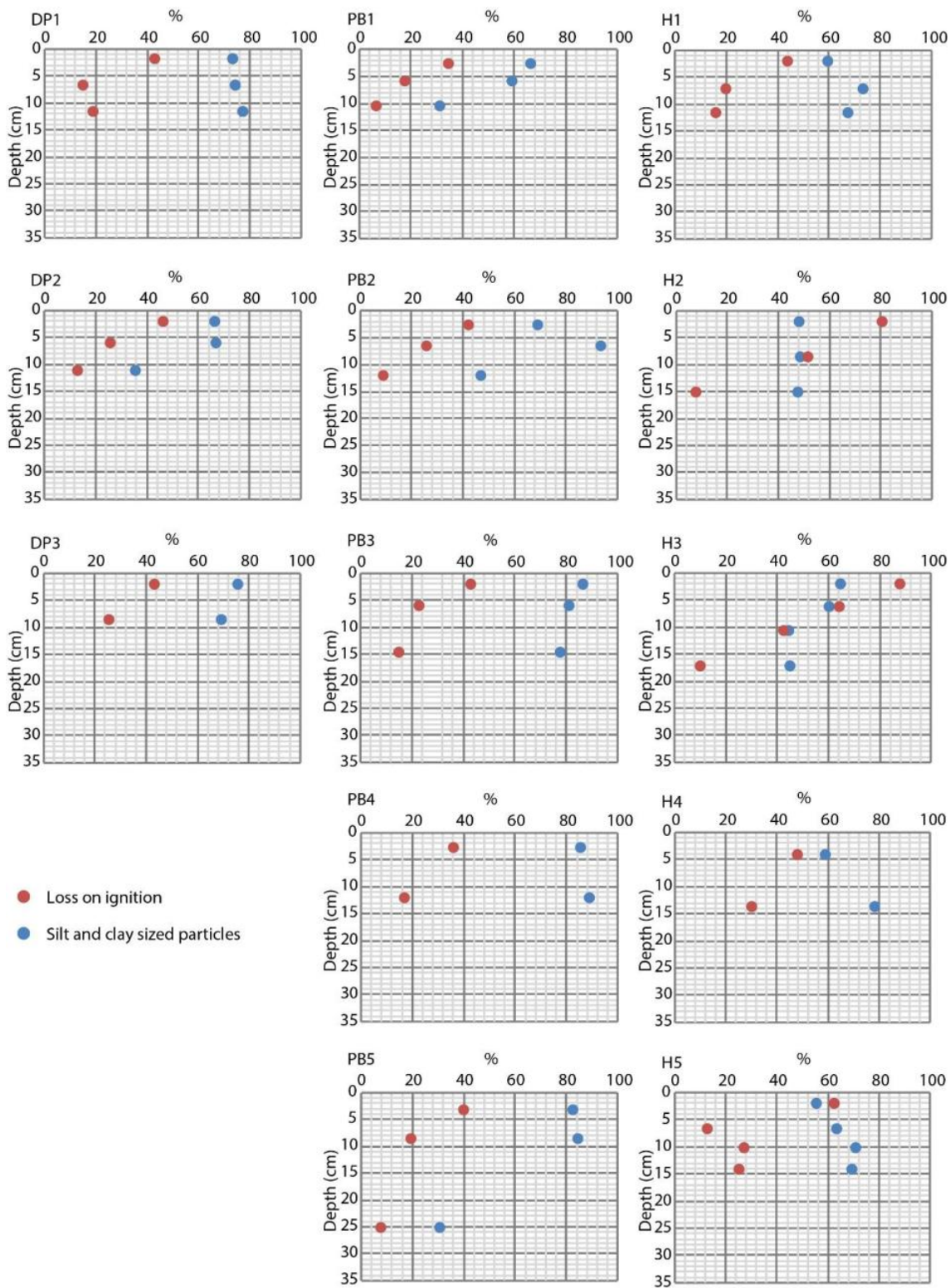
# Soil pit characteristics data

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt+clay (%) |
|---|---|---|---|---|---|---|
| MD4 | A | 3.0 | 5YR 2.5/1 | 5 | 32.4 | 91.1 |
| MD4 | B1 | 6.0 | 10YR 3/3 | 9 | 22.1 | 89.7 |
| MD4 | B2 | 3.0 | 10YR 3/3 | 9 | 11.5 | 52.4 |
| MD4 | C | | 2.5Y 4/3 | 6 | 3.8 | 34.4 |
| | | | | | | |
| MD3 | A | 4.0 | 7.5YR 2.5/1 | 4 | 50.1 | 60.1 |
| MD3 | B1 | 9.0 | 10YR 3/1 | 3 | 24.1 | 72.4 |
| MD3 | B2 | 7.0 | 10YR 3/3 | 9 | 11.2 | 56.1 |
| MD3 | C-ox | 9.0 | 7.5YR 3/2 | 8 | 9.5 | 27.7 |
| MD3 | C | | 2.5Y 4/2 | 4 | 3.3 | 21.6 |
| | | | | | | |
| MD2 | A | 3.0 | 7.5YR 2.5/1 | 4 | 70.0 | 86.4 |
| MD2 | E | 6.0 | 10YR 4/2 | 6 | 12.1 | 72.7 |
| MD2 | B1 | 7.0 | 7.5YR 4/3 | 12 | 10.7 | 83.2 |
| MD2 | B2 | 10.0 | 7.5YR 4/4 | 16 | 10.5 | 75.9 |
| MD2 | C | | 10YR 4/4 | 12 | 7.8 | 73.8 |
| | | | | | | |
| MD1 | B1 | 5.0 | 7.5YR 2.5/1 | 4 | 29.9 | 84.4 |
| MD1 | B2 | 9.0 | 7.5YR 3/2 | 8 | 20.8 | 79.6 |
| MD1 | B3 | 14.0 | 10YR 3/3 | 9 | 14.4 | 63.0 |
| MD1 | C-ox | 9.0 + | 10YR 3/4 | 12 | 11.0 | 60.4 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt+clay (%) |
|---|---|---|---|---|---|---|
| HON5 | OM | 3 | No sample | | | |
| HON5 | A | 7 | 10YR 2/1 | 3 | 63.3 | 65.9 |
| HON5 | E | 5 | 10YR 2/2 | 6 | 18.8 | 63.4 |
| HON5 | B | 9 | 10YR 3/6 | 18 | 9.9 | 56.8 |
| HON5 | C | | 10YR 4/3 | 9 | 9.2 | 43.9 |
| | | | | | | |
| HON4 | OM | 5 | No sample | n/a | n/a | n/a |
| HON4 | A | 4 | 7.5YR 2.5/1 | 4 | 57.8 | 83.4 |
| HON4 | B1 | 9 | 10YR 2/2 | 6 | 27.4 | 73.5 |
| HON4 | B2 | 10 | 10YR 3/3 | 9 | 15.6 | 56.6 |
| HON4 | C | | 10YR 4/3 | 9 | 2.5 | 20.0 |
| | | | | | | |
| HON3 | A | 9 | 7.5YR 2.5/1 | 4 | 49.1 | 75.9 |
| HON3 | E | 4 | 10YR 2/2 | 6 | 23.1 | 72.7 |
| HON3 | B1 | 5 | 7.5YR 4/4 | 16 | 13.9 | 57.1 |
| HON3 | B2 | 14 | 7.5YR 3/2 | 8 | 16.1 | 52.3 |
| HON3 | C | | 10YR 3/3 | 9 | 6.0 | 30.6 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| HON1 | OM | 8.0 | No sample | n/a | n/a | n/a |
| HON1 | A | 3.0 | 10YR 2/1 | 3 | 44.7 | 77.7 |
| HON1 | E | 4.0 | 7.5YR 3/1 | 4 | 17.0 | 88.5 |
| HON1 | B | 14.0 | 7.5YR 4/4 | 16 | 12.6 | 70.6 |
| HON1 | C | | 10YR 4/3 | 9 | 6.4 | 32.0 |
| | | | | | | |
| HON2 | OM | 2 | No sample | n/a | n/a | n/a |
| HON2 | A | 4 | 10YR 2/1 | 3 | 80.8 | 57.4 |
| HON2 | B | 9 | 10YR 2/2 | 6 | 29.9 | 39.2 |
| HON2 | C | | 10YR 4/3 | 9 | 3.6 | 28.6 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| STY1 | A | 4.0 | 10YR 2/1 | 3 | 65.5 | 73.4 |
| STY1 | E | 2.5 | 10YR 3/2 | 6 | 18.6 | 60.4 |
| STY1 | B1 | 6.0 | 10YR 3/4 | 12 | 7.0 | 67.1 |
| STY1 | B2 | 4.0 | 10YR 3/3 | 9 | 5.3 | 37.4 |
| STY1 | Cox | 4 | 10YR 3/4 | 12 | 4.7 | 56.1 |
| STY1 | Iron pan | | | | | |
| | | | | | | |
| AS1 | A | 2 | 10YR 2/1 | 3 | 81.8 | 57.6 |
| AS1 | B | 13 | 10YR 2/1 | 3 | 82.7 | 67.2 |
| AS1 | Cox | 5 | 10YR 2/2 | 6 | 10.7 | 26.6 |
| AS1 | Iron pan | | | | | |
| | | | | | | |
| GG1 | A | 4 | 7.5YR 2.5/1 | 4 | 65.8 | 81.0 |
| GG1 | E | 4 | 7.5YR 3/2 | 8 | 26.3 | 74.1 |
| GG1 | B1 | 4.5 | 10YR 2/2 | 6 | 29.9 | 75.3 |
| GG1 | B2 | 8 | 10YR 3/4 | 12 | 8.2 | 55.5 |
| GG1 | Cox | 8+ | 2.5Y 3/3 | 6 | 6.3 | 38.4 |
| | | | | | | |
| GG2 | A | 5 | 10YR 2/1 | 3 | 57.3 | 79.3 |
| GG2 | E | 7 | 10YR 3/2 | 6 | 25.5 | 54.9 |
| GG2 | B1 | 4.5 | 7.5YR 3/2 | 8 | 27.7 | 69.0 |
| GG2 | B2 | 9 | 7.5YR 3/4 | 16 | 16.8 | 68.8 |
| GG2 | Cox | 5+ | 10YR 3/3 | 9 | 8.1 | 40.8 |
| | | | | | | |
| GG3 | A | 5 | 10YR 2/1 | 3 | 68.0 | 79.9 |
| GG3 | E | 8 | 10YR 3/2 | 6 | 24.5 | 62.7 |
| GG3 | B1 | 2 | 10YR 3/2 | 6 | 28.1 | 93.4 |
| GG3 | B2 | 13 | 10YR 3/3 | 9 | 13.7 | 73.5 |
| GG3 | Cox | 10+ | 10 YR 3/4 | 12 | 7.7 | 47.6 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| ESK4 | OM | 2 | No sample | n/a | n/a | n/a |
| ESK4 | A | 5 | 7.5YR 2.5/1 | 4 | 27.0 | 95.3 |
| ESK4 | B1 | 14 | 10YR 2/2 | 6 | 18.1 | 60.8 |
| ESK4 | B2 | 14 | 10YR 3/2 | 6 | 6.8 | 14.0 |
| ESK4 | C | | 2.5Y 3/3 | 6 | 3.8 | 7.4 |
| | | | | | | |
| ESK3 | OM | 2 | No sample | n/a | n/a | n/a |
| ESK3 | A | 2 | 7.5YR 2.5/1 | 4 | 39.3 | 97.5 |
| ESK3 | E | 4 | 10YR 2/2 | 6 | 31.1 | 89.6 |
| ESK3 | B1 | 14 | 7.5YR 3/2 | 8 | 17.3 | 82.6 |
| ESK3 | B2 | 19 | 10YR 3/4 | 12 | 8.2 | 70.7 |
| ESK3 | C | | 10YR 3/3 | 9 | 6.6 | 41.2 |
| | | | | | | |
| ESK2 | OM | 1 | No sample | n/a | n/a | n/a |
| ESK2 | A | 2 | 10YR 2/1 | 3 | 35.1 | 94.2 |
| ESK2 | E | 2 | 7.5YR 2.5/1 | 4 | 35.9 | 84.0 |
| ESK2 | B1 | 11 | 10YR 3/3 | 9 | 23.4 | 94.4 |
| ESK2 | B2 | 12 | 7.5YR 2.5/2 | 8 | 16.2 | 58.5 |
| ESK2 | C | | No sample | n/a | n/a | n/a |
| | | | | | | |
| ESK1 | B | 4.0 | 10YR 3/2 | 6 | 31.9 | 84.5 |
| ESK1 | B1 | 14.0 | 7.5YR 3/1 | 4 | 17.0 | 87.1 |
| ESK1 | B2 | 8.0 | 7.5YR 3/2 | 8 | 15.6 | 71.7 |
| ESK1 | Cox/C | 7.0 | 10YR 3/3 | 9 | 8.4 | 48.1 |
| ESK1 | Cox | 14+ | 10YR 3/3 | 9 | 6.9 | 44.4 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| LN2 | A | 4.0 | 7.5YR 2.5/1 | 4 | 39.0 | 77.8 |
| LN2 | E | 1.5 | 7.5YR 3/1 | 4 | 23.8 | 63.1 |
| LN2 | B | 8.0 | 7.5YR 3/2 | 8 | 18.9 | 37.9 |
| LN2 | C-ox | 7.5 | 7.5YR 3/4 | 16 | 4.4 | 20.8 |
| LN2 | C | | 10YR 4/3 | 9 | 2.6 | 15.5 |
| | | | | | | |
| LN3 | A | 5.0 | 10YR 2/1 | 3 | 67.9 | 75.7 |
| LN3 | E | 6.0 | 7.5YR 3/1 | 4 | 30.1 | 72.6 |
| LN3 | B1 | 3.0 | 7.5YR 2.5/2 | 8 | 17.8 | 59.2 |
| LN3 | B2 | 7.0 | 10YR 3/3 | 9 | 6.4 | 29.3 |
| LN3 | C | | 10YR 4/3 | 9 | 2.3 | 20.2 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| LN1 | A | 4.0 | 7.5YR 2.5/1 | 4 | 33.1 | 81.8 |
| LN1 | E | 4.0 | 7.5YR 3/2 | 8 | 23.4 | 88.0 |
| LN1 | B1 | 5.0 | 7.5YR 3/3 | 12 | 17.4 | 62.6 |
| LN1 | B2 | 10.0 | 7.5YR 4/6 | 24 | 8.8 | 59.2 |
| LN1 | C | | 7.5YR 4/3 | 12 | 2.6 | 27.0 |
| | | | | | | |
| LN4 | A | 5.0 | 7.5YR 2.5/1 | 4 | 23.0 | 87.8 |
| LN4 | E | 3.0 | 7.5YR 3/3 | 12 | 20.1 | 82.7 |
| LN4 | B1 | 10.0 | 7.5YR 3/4 | 16 | 15.1 | 73.5 |
| LN4 | B2/Cox | 17.0 | 7.5YR 3/4 | 16 | 7.9 | 49.6 |
| LN4 | C | | 7.5YR 3/3 | 12 | 4.3 | 28.9 |
| | | | | | | |
| LN5 | A | 4.0 | 7.5YR 2.5/1 | 4 | 40.0 | 73.2 |
| LN5 | E | 4.0 | 10YR 3/2 | 6 | 15.7 | 67.4 |
| LN5 | B | 15.0 | 10YR 3/3 | 9 | 10.2 | 52.5 |
| LN5 | C | | 10YR 4/3 | 9 | 3.1 | 29.5 |
| | | | | | | |
| RS1 | B1 | 7.0 | 7.5YR 3/1 | 4 | 17.2 | 61.1 |
| RS1 | B2 | 6.0 | 7.5YR 3/2 | 8 | 13.0 | 54.8 |
| RS1 | B3 | 9.0 | 7.5YR 3/4 | 16 | 10.1 | 52.1 |
| RS1 | B4 | 8.0 | 10YR 3/3 | 9 | 5.7 | 29.9 |
| RS1 | C-ox/C | 10.0 + | 10YR 3/4 | 12 | 4.8 | 27.2 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| WT2 | A | 3.0 | 7.5YR 2.5/1 | 4 | 48.5 | 91.6 |
| WT2 | E | 2.5 | 10YR 3/2 | 6 | 18.7 | 77.6 |
| WT2 | B1 | 10.0 | 10YR 3/3 | 9 | 8.2 | 63.8 |
| WT2 | Cox | 10+ | 7.5YR 4/3 | 12 | 3.2 | 51.2 |
| WT2 | C | | No sample | n/a | n/a | n/a |
| | | | | | | |
| WT1 | OM | 4.0 | 10YR 3/2 | 6 | 25.0 | 88.7 |
| WT1 | B1 | 7.0 | 10YR 3/2 | 6 | 10.9 | 59.6 |
| WT1 | B2 | 12.0 | 10YR 3/3 | 9 | 8.3 | 66.4 |
| WT1 | Cox | 4.0 | 10YR 4/3 | 9 | 5.2 | 45.1 |
| WT1 | C | | 10YR 4/4 | 12 | 2.2 | 42.7 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| DP1 | A | 3.5 | 7.5YR 2/0 | 0 | 42.6 | 73.0 |
| DP1 | B1 | 6 | 5YR 3/1 | 5 | 14.8 | 74.0 |
| DP1 | B2 | 4 | 5YR 3/2 | 10 | 18.5 | 77.0 |
| DP1 | C-ox | 10 | 10YR 3/4 | 12 | 9.2 | 67.6 |
| DP1 | C | | 10YR 4/1 | 3 | 2.5 | 40.2 |
| | | | | | | |
| DP2 | A | 4 | 7.5YR 2/0 | 0 | 46.0 | 66.1 |
| DP2 | B1 | 3.5 | 5YR 3/1 | 5 | 25.6 | 66.6 |
| DP2 | B2 | 7 | 10YR 2/2 | 6 | 13.0 | 35.5 |
| DP2 | C | | 10YR 3/3 | 9 | 4.1 | 13.6 |
| | | | | | | |
| DP3 | A | 4 | 7.5YR 2/0 | 0 | 43.1 | 75.5 |
| DP3 | B | 9 | 10YR 2/2 | 6 | 25.3 | 69.0 |
| DP3 | C-ox | 16 | 10YR 3/3 | 9 | 7.8 | 35.8 |
| DP3 | C | | 10YR 3/4 | 12 | 4.5 | 37.9 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| PB1 | A | 5.0 | 10YR 2/1 | 3 | 34.2 | 66.4 |
| PB1 | E | 1.5 | 10YR 3/1 | 3 | 17.5 | 59.2 |
| PB1 | B | 7.5 | 10YR 3/2 | 6 | 6.5 | 30.9 |
| | | | | | | |
| PB2 | A | 5.0 | 10YR 2/1 | 3 | 42.0 | 68.8 |
| PB2 | E | 3.0 | 10YR 3/1 | 3 | 25.4 | 93.3 |
| PB2 | B | 8.0 | 10YR 3/2 | 6 | 8.8 | 46.5 |
| | | | | | | |
| PB3 | A | 4.0 | 7.5YR 2.5/1 | 4 | 42.5 | 86.1 |
| PB3 | E | 3.5 | 7.5YR 3/2 | 8 | 22.5 | 80.7 |
| PB3 | B | 14.0 | 7.5YR 4/3 | 12 | 14.6 | 77.3 |
| | | | | | | |
| PB4 | A | 5.0 | 10YR 2/1 | 3 | 35.8 | 85.3 |
| PB4 | B | 14.0 | 7.5YR 3/2 | 8 | 16.6 | 88.7 |
| | | | | | | |
| PB5 | A | 6.0 | 10YR 2/1 | 3 | 39.6 | 82.1 |
| PB5 | E | 5.0 | 10YR 3/1 | 3 | 19.3 | 84.2 |
| PB5 | B | 28.0 | 10YR 3/2 | 6 | 7.6 | 30.6 |

| Pit Number | Horizon | Thickness (cm) | Colour | CDE | LOI (%) | Silt/clay (%) |
|---|---|---|---|---|---|---|
| H1 | A | 4.0 | 7.5YR 2.5/1 | 4 | 43.5 | 59.5 |
| H1 | B1 | 6.0 | 10YR 3/1 | 3 | 19.8 | 73.1 |
| H1 | B2 | 3.0 | 7.5YR 3/2 | 8 | 15.7 | 67.1 |
| H1 | Cox | 3.0 | 10YR 4/4 | 12 | 6.4 | 69.1 |
| H1 | C | | 2.5Y 4/3 | 6 | 3.4 | 50.3 |
| | | | | | | |
| H2 | A | 4.0 | 10YR 2/1 | 3 | 80.5 | 47.6 |
| H2 | B1 | 9.0 | 10YR 2/2 | 6 | 51.4 | 48.1 |
| H2 | B2 | 4.0 | 7.5YR 3/2 | 8 | 7.6 | 47.4 |
| H2 | C | | 2.5Y 4/3 | 6 | 2.4 | 23.6 |
| | | | | | | |
| H3 | A | 4.0 | 7.5YR 2.5/1 | 4 | 87.6 | 64.4 |
| H3 | B1 | 4.0 | 7.5YR 2.5/1 | 4 | 64.0 | 59.8 |
| H3 | B2 | 5.0 | 7.5YR 3/1 | 4 | 42.3 | 44.4 |
| H3 | B3 | 8.0 | 7.5YR 3/2 | 8 | 9.7 | 44.7 |
| H3 | Cox | 3.0 | 10YR 3/4 | 12 | 9.7 | 54.6 |
| H3 | C | | 2.5YR 4/3 | 18 | 5.9 | 48.2 |
| | | | | | | |
| H4 | A | 8.0 | 10YR 2/1 | 3 | 47.5 | 58.4 |
| H4 | B | 11.0 | 10YR 2/2 | 6 | 30.0 | 77.8 |
| H4 | C | | 2.5 YR 4/3 | 18 | 4.8 | 35.4 |
| | | | | | | |
| H5 | A | 4.0 | 10YR 2/1 | 3 | 62.0 | 55.0 |
| H5 | E | 5.0 | 7.5YR 3/1 | 4 | 12.6 | 63.2 |
| H5 | B1 | 2.0 | 10YR 3/2 | 6 | 27.0 | 70.5 |
| H5 | B2 | 6.0 | 7.5YR 3/2 | 8 | 25.0 | 69.2 |
| H5 | C-ox | 13.0 | 7.5YR 3/4 | 16 | 16.2 | 63.5 |
| H5 | C | | No Sample | n/a | n/a | n/a |

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