

The glacial geomorphology of the Loch Lomond (Younger Dryas) Stadial in Britain: a review

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ABSTRACT: This paper systematically reviews the glacial geomorphological evidence of the Loch Lomond Stadial (LLS; 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 LLS 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 LLS glaciers and assess their synchronicity. Numerical models have allowed the glacier–climate relationships of the LLS 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 LLS in Britain and contribute to a more comprehensive understanding of glacier–climate interactions during the Younger Dryas. © 2018 The Authors. *Journal of Quaternary Science* Published by John Wiley & Sons Ltd

KEYWORDS: Britain; glacial geomorphology; Loch Lomond Stadial; palaeoglaciology; Younger Dryas.

Introduction

Following the Last Glacial Maximum (LGM) between 26.5 and 19 ka (Clark *et al.*, 2009) the Northern Hemisphere experienced abrupt cooling 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 $\delta^{18}\text{O}$ 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 (deMenocal *et al.*, 2000; Nakagawa *et al.*, 2003; Genty *et al.*, 2006), but particularly in the North Atlantic and surrounding regions (Benway *et al.*, 2010). However, global net cooling was only of approximately 0.6 °C, as cooling in the Northern Hemisphere was offset by warming in the Southern Hemisphere (Shakun and Carlson, 2010).

Although these effects were not experienced synchronously across all regions, the abrupt nature of the transition into cold conditions during a period of increasing insolation supports the long-held assumption that the YD was caused by disruption of the Atlantic meridional overturning circulation (AMOC), which moderates the region's climate (Carlson, 2010; Golledge, 2010). This disruption probably resulted from an influx of freshwater into the North Atlantic (Broecker, 2006; Carlson and Clark, 2012), caused either by: glacial outburst floods of meltwater northwards along the Mackenzie River system (Murton *et al.*, 2010); an extra-terrestrial impact causing destabilization of the Laurentide Ice Sheet (Firestone *et al.*, 2007); catastrophic break up and expulsion of palaeocrystic sea ice from the Arctic Ocean (Bradley and England, 2008); or retreat of the Laurentide Ice Sheet allowing meltwater to drain eastwards along the St Lawrence River

into the North Atlantic (Johnson and McClure, 1976; Teller *et al.*, 2002; Carlson, 2013). The resulting cooling caused the expansion of circum-Atlantic ice masses in North America (Mott and Stea, 1993; Lowell *et al.*, 1999; Occhietti *et al.*, 2004; Occhietti, 2007), Iceland (Ingólfsson *et al.*, 2010) and Scandinavia (Andersen *et al.*, 1995; Mangerud *et al.*, 2011).

Although it is difficult to match the timing of this event across global records during 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, and valley and cirque/niche glaciers in surrounding upland areas (Golledge, 2010) (Fig. 1). The existence of a phase of valley glaciation following the retreat of the British-Irish Ice Sheet was recognized 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 around the south-eastern margin of Loch Lomond, SW Scotland. Subsequently, Charlesworth (1955) reconstructed ice marginal positions, identifying a 'Moraine Glaciation' stage during the retreat of the last British-Irish Ice Sheet, in many cases the extent of which matches more recently proposed reconstructions of the LLS glaciers. However, it was only in the 1970s that systematic mapping of the glacial landforms attributed to this period was conducted, most notably by Sissons (1972, 1974, 1977a, b). From this, and other research, the extent of LLS glaciers in the Scottish Highlands and the English Lake District was synthesized by Sissons (1979a), alongside an overview of the types of landforms evidence and environmental conditions associated with the period.

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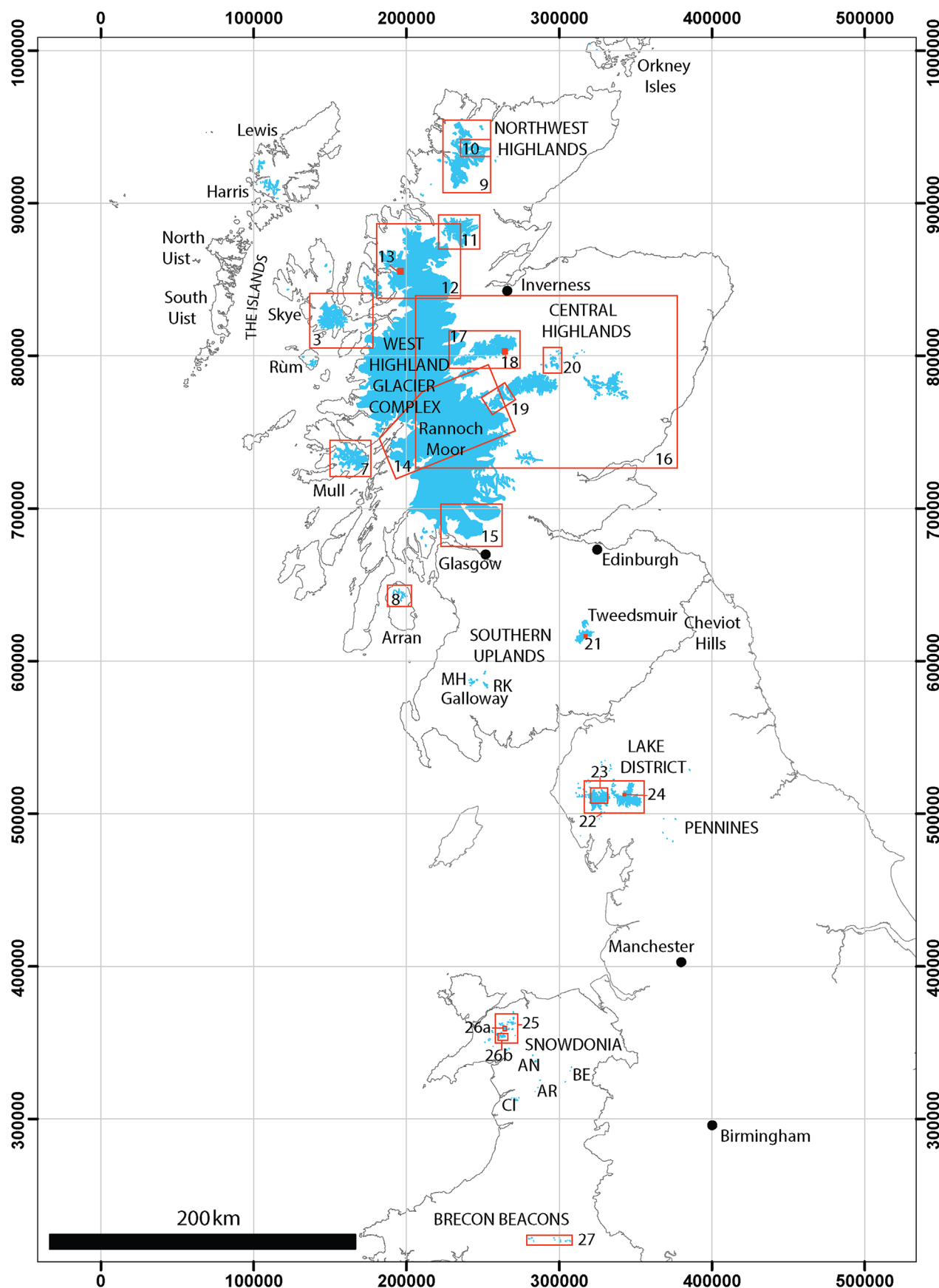


Figure 1. Location map of regions where LLS glacial geomorphology is present, as discussed in this paper, showing the approximate extent of the ice. The location of figures is indicated in red or is indicated later in the paper. MH: Merrick Hills, RK: Rhinns of Kells AN: Arenig Mountains, AR: Aran Mountains, BE: Berwyn Mountains, CI: Cadair Idris. Britain coastline reproduced from Ordnance Survey © Crown copyright and database Right 2015. Ordnance Survey (Digimap Licence).

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 published evidence for the glacial geomorphology of the LLS in Britain into a geographic information system (GIS) database and accompanying glacial map; this is a similar approach to that taken for the compilation of the evidence associated with the last (Dimlington Stadial) British Ice Sheet by Clark *et al.* (2004) and Evans *et al.* (2005). While 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.

The compilation of the geomorphological evidence based upon an extensive search of the published literature has enabled a systematic evaluation of a substantial LLS palaeogeographical 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 Fig. 1. An example from the complete map, as available in Bickerdike *et al.* (2016), is shown in Fig. 2. 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 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 regional compilation within the GIS database is used to reconstruct the extent of the LLS glaciation, based upon the literature available up to the census end date of June 2014. However, although in some areas the limits of LLS glaciation are well constrained by both geomorphology and absolute dates, in many locations the extent of ice is much less certain, despite advances made in recent decades to map, reconstruct, model and date these ice masses. This paper qualitatively assesses the level of confidence in the extent and chronological control of the LLS glaciers, highlights the implications of these limitations for the current understanding of LLS glaciation and makes recommendations for future research to remedy such shortfalls.

A wide variety of chronological evidence is discussed within this review. All quoted radiocarbon dates have been recalibrated for this review to allow comparison of ages from different sites, collected by different researchers, and are shown as calibrated years BP within two sigma confidence limits. Calibration was done using Calib 7.10 (Stuiver *et al.*, 2017), using the IntCal13 curve, with the exception of one sample from Menteith, which used the MARINE13 curve with a Delta *R*-value of -105 ± 42 . Similarly, cosmogenic exposure dates have been recalibrated using the CRONUS-Earth V2.3 online calculator (Balco *et al.*, 2008), using the Loch Lomond Production Rate (Fabel *et al.*, 2012) with a nominal sea-level high-latitude value of 4.00 ± 0.17 atoms $\text{g}^{-1} \text{a}^{-1}$, *Lm* scaling (Lal, 1991; Stone, 2000), and assuming an erosion rate of 1 mm ka^{-1} . Recalibrated dates are available as supplementary information. ^{36}Cl ages have not been recalibrated as there was insufficient information in the original publications to recalculate these with up-to-date production rates. Analytical uncertainties are shown in parentheses after each date. [Correction

added on 23 January 2018, IntCal13 in line 8 of this paragraph has been replaced with MARINE13.]

Wherever possible, we have used the terminology given in the relevant papers to ensure that the interpretations from the original research are communicated as accurately as possible. However, some generalization of these terms is required within figure captions (e.g. 'flutes', 'flutings' and 'fluted moraines' are all classified under 'flutings').

Geomorphological evidence in Scotland

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; 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, 2007a) and the Outer Hebrides (Ballantyne, 2006, 2007b), supported only valley glaciers while glaciation on the Orkney Isles was restricted to favourably situated cirques (Ballantyne *et al.*, 2007).

Orkney

Hoy, in the Orkney Isles, represents the most northerly site of glaciation in Britain during the LLS but ice was restricted to two small cirques, Enegars and Dwarfie Hamars, in the island's upland area (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 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 13.3 ± 1.3 (1.2) ka and 11.6 ± 1.8 (1.8) ka at Enegars and 13.8 ± 1.5 (1.3) ka and 13.0 ± 1.8 (1.8) ka at Dwarfie Hamars (Ballantyne *et al.*, 2007), supporting a LLS age in accordance with the significantly more developed scree slopes found only outside the proposed limits (Sutherland, 1996b).

While other authors previously argued for more extensive glaciation of Orkney (Charlesworth, 1955; Rae, 1976), it has subsequently been argued that the thick drift 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.

Outer Hebrides

Much of the generally low-lying rugged landscape of the Outer Hebrides was last glaciated by a locally nourished ice

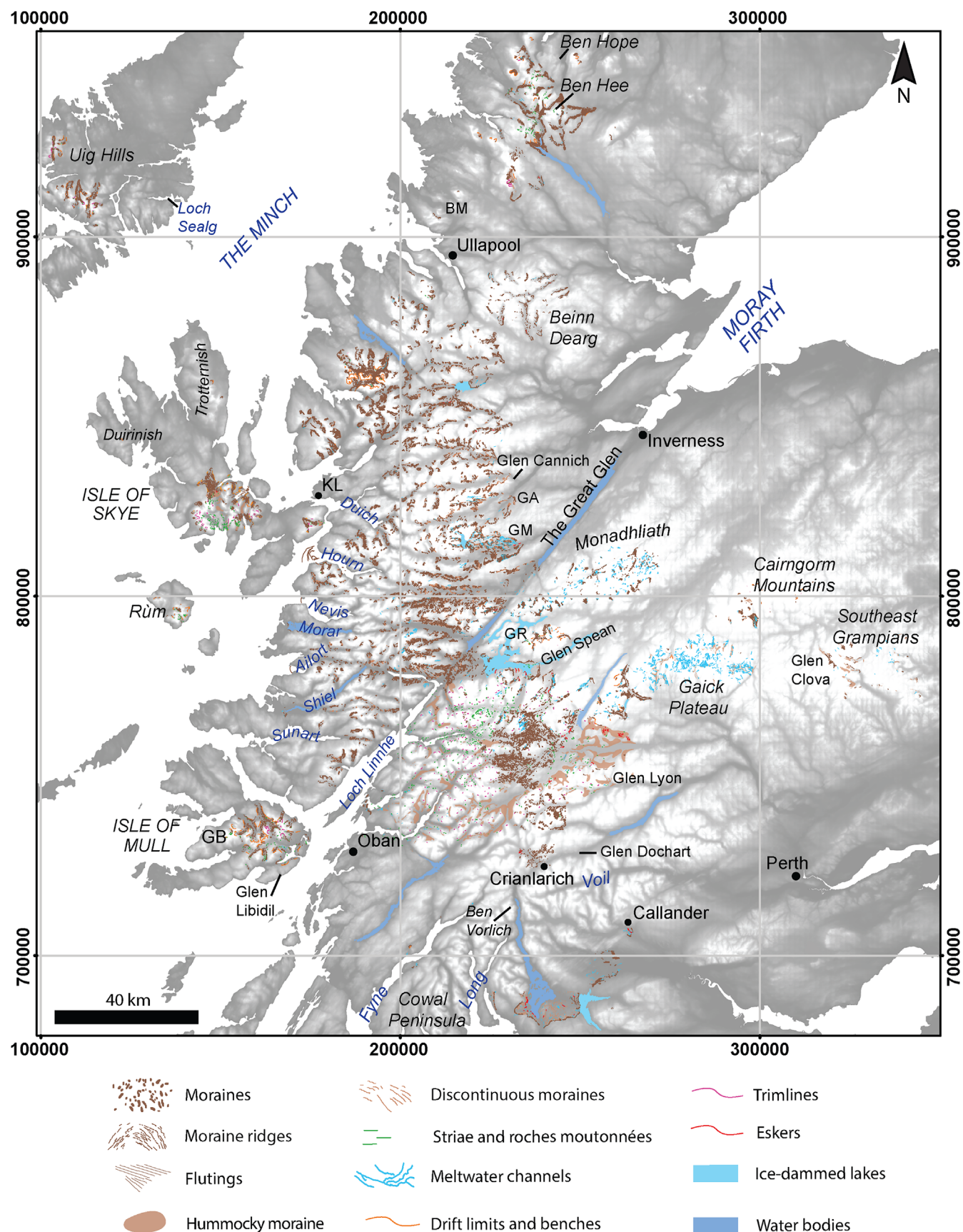


Figure 2. LLS glacial geomorphology of the Highlands and islands of Scotland, adapted from Bickerdike *et al.* (2016). The map shows the distribution of landforms relating to the West Highland Glacier Complex which covered most of the Western Grampian Mountains and was flanked by satellite icefields in neighbouring upland areas. BM: Ben Mor Coigach, KL: Kyle of Lochalsh, GB: Gribun, GA: Glen Affric, GM: Glen Moriston, GR: Glen Roy. 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.

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, 2007b).

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. They produced 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. The Dibadale and Suainaval glaciers drained from the eastern ridge, although 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. 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, citing trimlines and the alignment of lateral moraines as evidence that the Raonasgail was insufficiently thick to override the mountain ridge and coalesce with the Suainaval glacier.

LLS glaciers were more extensive in 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, 2007b). 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, 2007b; cf. incremental stagnation of Eyles, 1983). While the presence of moraines that extend to the glaciers' source area accords with the evidence found in the Uig Hills (Ballantyne, 2006), the organization 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 An Coire, Gleann Dubh and Gleann Bhearry cirque glaciers).

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 pre-date LLS glaciation (Ballantyne, 2007b). 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 geomorphological evidence has been used to infer a LLS age for the phase of valley glaciation (Ballantyne, 2006, 2007b). 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 after 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 area of moraines was 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 20.5 ± 1.0 (0.5) ka (Stone and Ballantyne, 2006), suggesting that the LLS represents the most probable cold period during which this readvance could have occurred.

Isle of Skye

The Isle of Skye (Fig. 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 phase of local glaciation during the LLS has long been recognized (Forbes, 1846; Harker, 1901; Charlesworth, 1955), the first detailed geomorphological mapping of the area was conducted by Sissons (1977a) who reconstructed nine cirque and valley glaciers in the Cuillins and four in the Eastern Red Hills (Fig. 4a). Subsequent 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) (Fig. 4b).

The LLS icefield occupied the mountainous area of the Cuillins and Red Hills, covering an area of approximately 150 km² (Ballantyne, 1989). The icefield showed pronounced north–south asymmetry, resulting from the low-gradient piedmont lobe which extended northwards from the Cuillins down Glens Sligachan, Drynoch and Varragill (Ballantyne, 1989) (Fig. 3). This area is underlain by a compact, sheared till that deformed under low shear stresses beneath the LLS glaciers, producing the low gradient ice lobe (Ballantyne, 1989; Benn, 1992). The hummocky moraines that mark the extent of the northern outlet glaciers are arranged into a complex pattern of transverse recessional moraines, chaotic mounds and flutings (Benn *et al.*, 1992) (Fig. 5a). These features were deemed by Sissons (1977a) to pre-date the LLS but pollen stratigraphies from within this area were found to date from the early Holocene onwards, confirming that these features were formed during the LLS (Benn *et al.*, 1992) (Chronological Site 1, Fig. 3).

The extent of glaciers which drained the eastern sector of the icefield is also demarcated by hummocky moraines. In Gleann Torra-mhichaig, 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). Intermittent recessional moraines are present along the sides and floor of Loch Ainort but these are replaced by isolated hummocks at the head of the loch and sedimentary exposures suggest that they are ice stagnation moraines (Benn *et al.*, 1992). Bathymetric surveys have revealed that a sequence of moraines is present at the opening of Loch Ainort (Fig. 3), suggesting

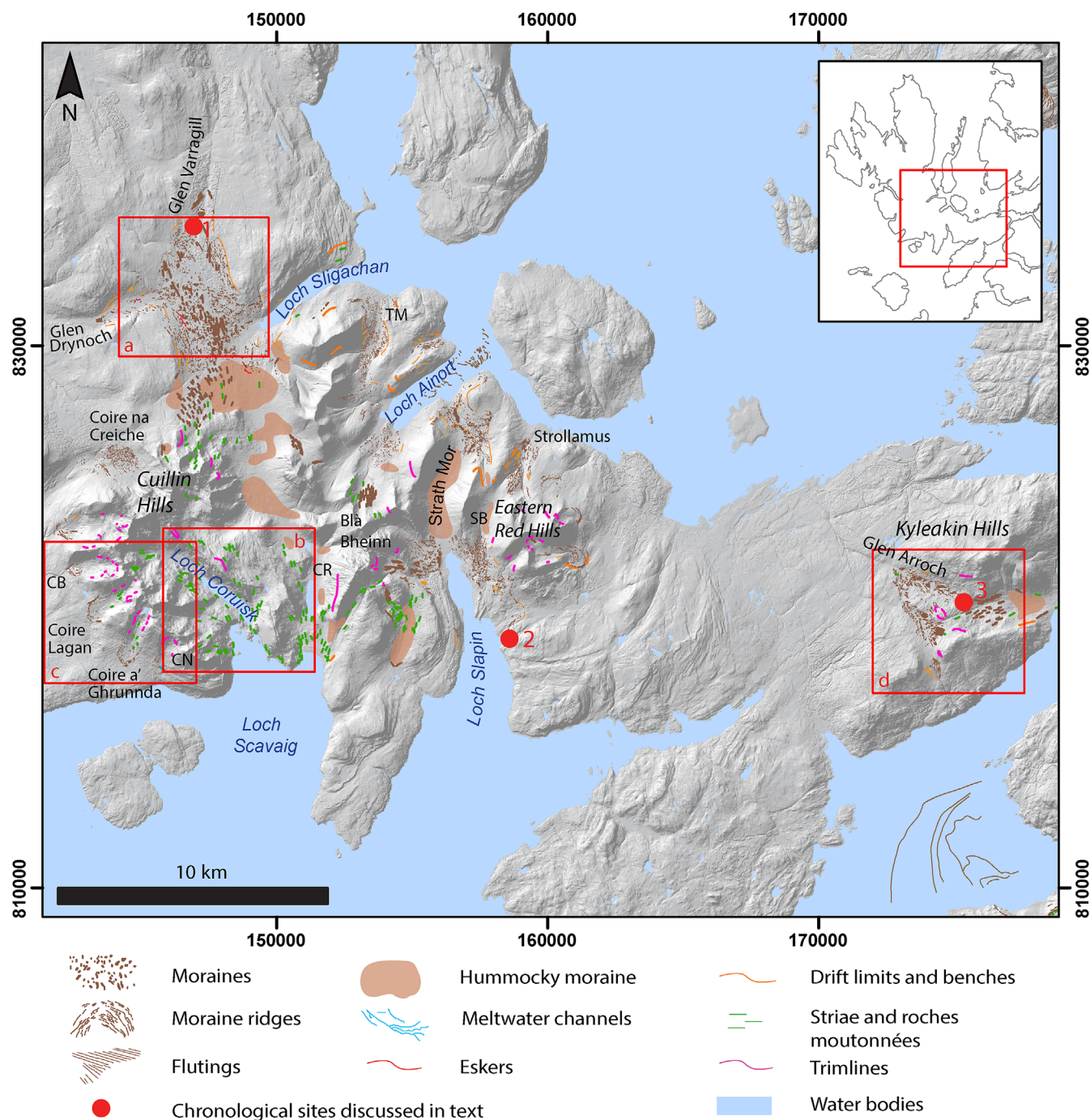


Figure 3. Geomorphological mapping compiled for the glacial map for the Isle of Skye (after Ballantyne, 1989; Benn, 1990, 1992, 1993; Benn *et al.*, 1992; Dix and Duck, 2000). Location boxes included for Fig. 5. TM: Gleann Torra-mhichaig, CB: Coire na Banachdich, CN: Coire nan Laogh, CR: Creitheach, SB: Strath Beag. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

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). Evidence of two-phase retreat is also present in Strath Mor; the lower valley displays clear moraines while the upper valley is occupied by a thick drift sheet of irregular hummocks and dissected kame terraces, suggestive of ice stagnation (Benn *et al.*, 1992).

Likewise, the limit of the southward flowing Slapin glacier is marked by a series of prominent end moraines on the eastern loch shore. These features have been dated to between 13.1 ± 0.9 (0.7) and 11.8 ± 0.9 (0.7) ka (excluding one anomalous age) (Small *et al.*, 2012; Ballantyne *et al.*, 2016) (Chronological Site 2, Fig. 3) and 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, but from approximately 1.5 km inside the limit, well-developed, sharp-crested moraine ridges occupy the slopes of Strath Beag (Fig. 6a), whereas the lower flatter ground is dominated by recessional moraines formed by

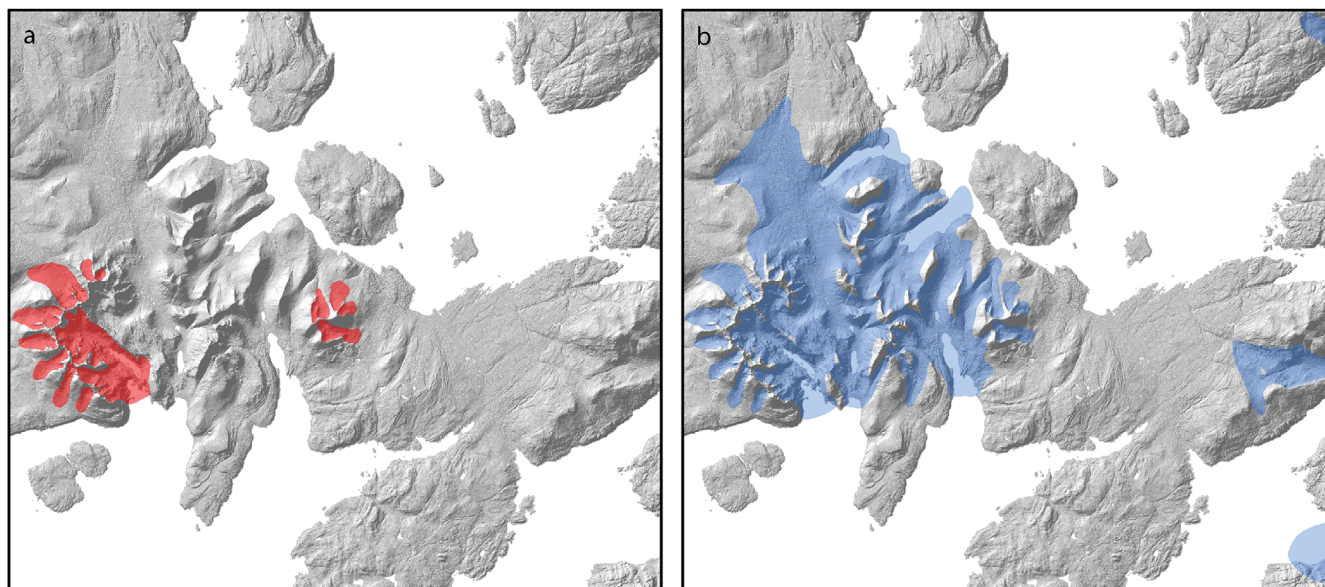


Figure 4. The extent of LLS glaciation on the Isle of Skye (a) after Sissons (1977a) and (b) as inferred from the subsequent geomorphological mapping, after Ballantyne (1989), Benn (1990, 1992, 1993), Benn *et al.* (1992) and Dix and Duck (2000). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

chains of large mounds and ridges (Benn *et al.*, 1992). These features continue up-valley until transitioning into undulating drift, again supporting the notion of two-phased retreat. Conversely, evidence of the neighbouring Coruisk and Creitheach glaciers consists primarily of ice-moulded bedrock and striae with a general absence of drift cover (Fig. 5b). In some locations, such as the north-west shore of Loch Scaavaig (Fig. 3), reconstruction of the margins of these glaciers has only been possible by identifying where LLS striations overprint older erosional features indicative of LGM ice flow (Ballantyne, 1989).

The main Cuillin icefield was flanked by nine cirque glaciers (Fig. 3). Most of these glaciers are characterized by chains of moraine ridges and boulders arranged in belts at each former glacier terminus and upper cirques dominated by ice-moulded bedrock and striations (Sissons, 1977a; Ballantyne, 1989; Benn, 1990) (Fig. 5c). Benn (1990) argued that both the Coire na Banachdich and the Coire Lagan glaciers were less extensive than had been proposed by Ballantyne (1989), suggesting that some moraine ridges within the area pre-date the LLS, and reconstructing glaciers that were similar to those proposed by Sissons (1977a). The arrangement of ridges in both the Cuillin and the 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 (Fig. 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 cross-valley pairs of chevron-shaped bands pointing down-valley (Benn, 1992; Benn *et al.*, 1992) (Fig. 5d). A sediment core from near

the head of the glen (Chronological Site 3, Fig. 3) revealed a transition from basal minerogenic to organic sediments containing a complete early Holocene 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 (Fig. 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 among the thick peat and landslide blocks of the surrounding area (Charlesworth, 1955; Ballantyne, 1990). A further small glacier has been reconstructed in Glen Osedale on the Duirinish Peninsula (Fig. 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 (1977a) 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 (1977a) 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

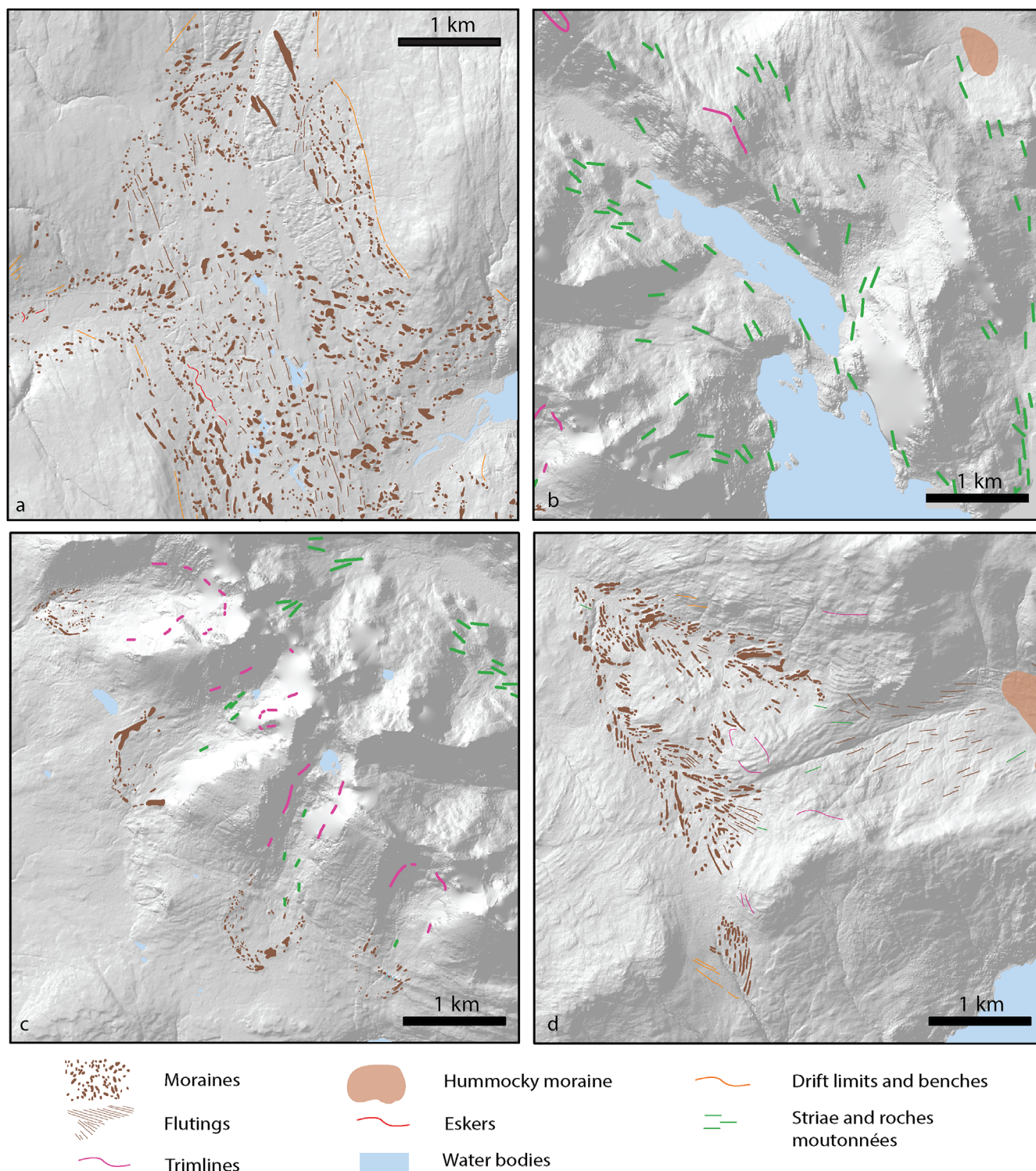


Figure 5. Examples of the varied LLS glacial geomorphology of the Isle of Skye. (a) Hummocky moraine in Glen Sligachan, comprising transverse recessional moraines, flutings and chaotically arranged mounds (after Benn *et al.*, 1992). (b) Erosional evidence, including striae, roches moutonnées and trimlines 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, 1992). (d) Recessional moraines marking the extent of the small Kylekin 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.

case everywhere around the island, although absolute dating is still required to confirm a LLS age.

Rùm

The glacial history of the Isle of Rùm (Fig. 2) has been the focus of relatively limited research. While 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



Figure 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. (c) Recessional moraines associated with the terminus of the Lomond Lobe at Gartness. 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 Fig. 25). This feature was formed during a phase of re-stabilization 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 moraine and a protalus rampart. Photograph (a) from H.L.B. and (b)–(f) from D.J.A.E.

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 impressive 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 (Ballantyne and Wain-Hobson, 1980). Thus, while there are

no absolute dates on the features demarcating the limits, it is reasonable to interpret these features as LLS in age.

Isle of Mull

The Isle of Mull (Fig. 7) supported an independent ice dome that deflected mainland ice around it during the LGM, before nourishing a mountain icefield during the LLS. Bailey *et al.* (1924) first recognized that, while the north and west of the island showed evidence of LGM glaciation,

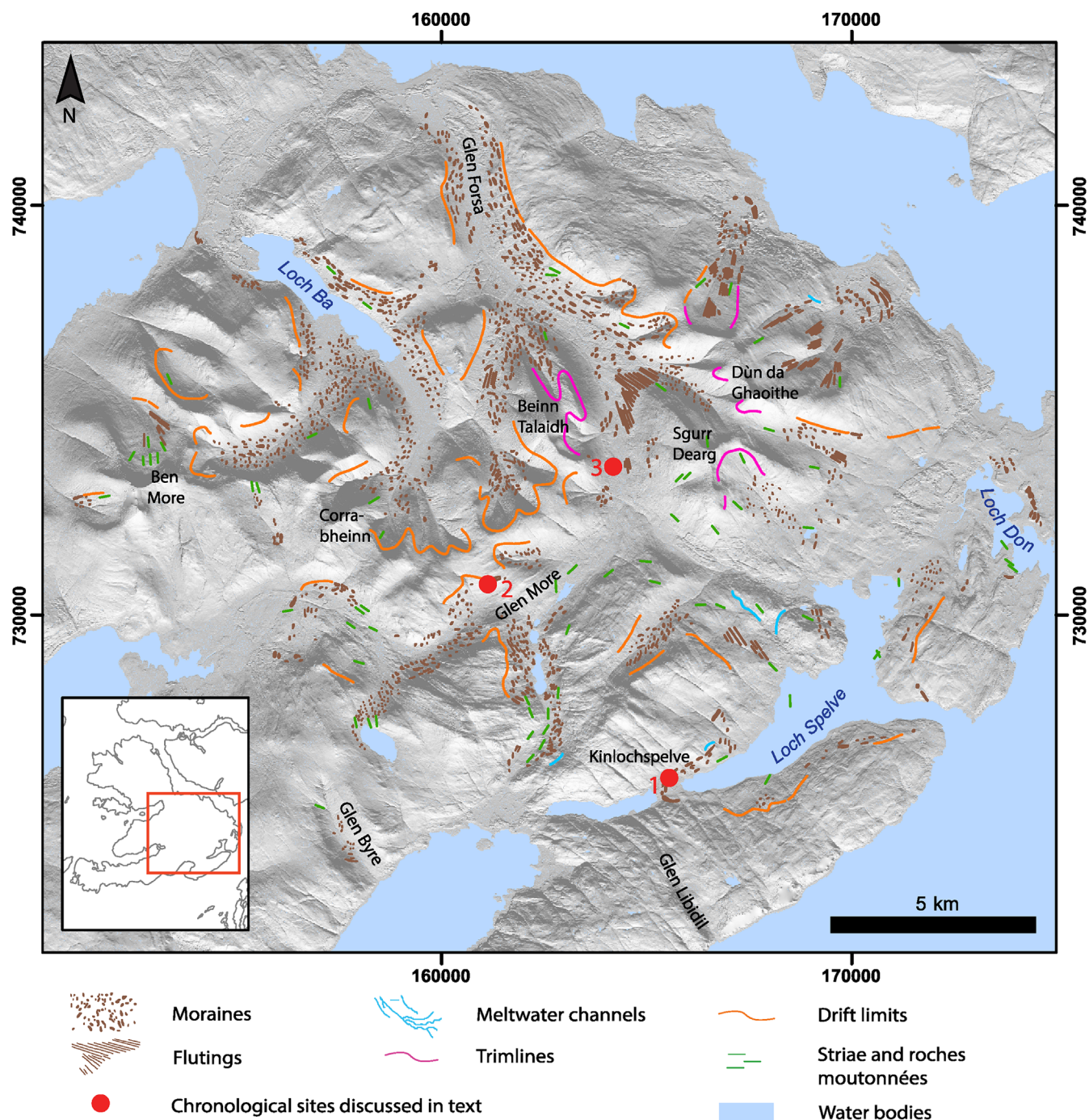


Figure 7. Geomorphological mapping compiled for the glacial map for the Isle of Mull (after Ballantyne, 2002). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

the valleys in the central and eastern mountainous area were characterized by hummocky deposits from a phase of valley glaciation, inferring 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, 2002), 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 north-west and north, respectively (Fig. 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, 2002) (Fig. 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, 2002), the latter being favoured herein. 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 (Fig. 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 1 km up-valley, as later confirmed by Ballantyne (2002). The neighbouring Spelve–Don lobe glacier appears to have

deposited few moraines, contrasting with the almost complete coverage of morainic material in the northern valleys, and much of the area is characterized by ice-moulded bedrock (Fig. 7). The outer limit of the glacier is marked by occasional fragmented end moraine complexes, and sections in the Loch Don end moraine (Fig. 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).

Six cirque glaciers flanked the Mull icefield (Fig. 7). The two largest cirque glaciers, which formed to the north-east of the Dun da Ghaoithe ridge, and the small Glen Byre glacier formed clear sequences of nested recessional moraines, whereas the three glaciers along the north-west periphery of the icefield produced only drift limits, striae and occasional hummocky moraines (Ballantyne, 2002). The contrast between the three small cirque glaciers on the flanks of the highest peak, Ben More, compared to the vast Spelve-Don lobe, which extended to the present coastline, has been attributed to the eastwards redistribution of windblown snow. This also accounts for the low equilibrium line altitudes (ELAs) of glaciers situated downwind of potential snow-contributing upland areas, such as the Glen Byre glacier (Ballantyne, 2002).

Further small cirque glaciers were proposed to have developed in Glen Libidil (Bailey *et al.*, 1924; Gray and Brooks, 1972) and at Gribun (Dawson *et al.*, 1987), approximately 6 km west of the area in Fig. 7. However, Ballantyne (2002) found no evidence for LLS glaciers at the former site and argued that the arcuate ridge at Gribun represents a rock-slope 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.

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 (Chronological Site 1, Fig. 7), substantiated by pollen analysis of the same sediments (Gray and Brooks, 1972). The presence of only Holocene sediments from infilled kettle hole basins within the glacier limits in Glen More (Chronological Sites 2 and 3, Fig. 7) 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, 2002).

Isle of Arran

Although relatively little research has been published regarding the LLS on the Isle of Arran, two contrasting styles of glaciation have been proposed for the island (Fig. 8). Evidence for a phase of local glaciation can be found in the valleys of the mountainous northern half of the island, which is characterized by granite peaks surrounded by cirques and glacial valleys. Gemmell (1973) noted that many of these valleys were occupied by locally derived grey till, which 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 (Fig. 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 (Fig. 8a,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 (Fig. 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 (2007a) argued that the moraines in the lower reaches of Glen Rosa, Glen Sannox, North Glen Sannox, Coire nam Meann, Gleann Easan Biorach, Glen Catacol and Glen Iorsa pre-date the LLS 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 (Fig. 8c). This glacial reconstruction contrasted with Gemmell's (1971, 1973) interpretation, most significantly for Glen Iorsa. In Gemmell's (1973) earlier reconstruction the Iorsa glacier reaches the lower valley, but Ballantyne (2007a) argued that it terminated approximately 7-km further up-valley in a series of subdued, often peat-covered, moraine ridges. Ballantyne (2007a) also 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 (2007a) argued is a bedrock feature. Gemmell (1973) also proposed that valley glaciers had extended down to approximately 30 m a.s.l. in Glen Sannox and Glen Rosa and had deposited the outwash at the mouths of these valleys. However, Ballantyne (2007a) 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 (Fig. 8a), where such moraines mark the limit of bouldery debris (Ballantyne, 2007a).

In the absence of chronological control it is not possible to refute the notion that the features beyond the limits suggested by Ballantyne (2007a) are of LLS age. It is possible that the features 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). 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 robust chronological control on its glacial landforms.

North-west Highlands

The geomorphology of the icefields and cirque glaciers that flanked the northern sector of the main icefield in the north-west (NW) Highlands has been relatively well recorded but remapping of these features (e.g. Lukas and Lukas, 2006a,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 (Lukas and Bradwell, 2010) (Figs 9 and 10), with a small icefield forming on the Applecross Peninsula and an ice cap occupying the Beinn

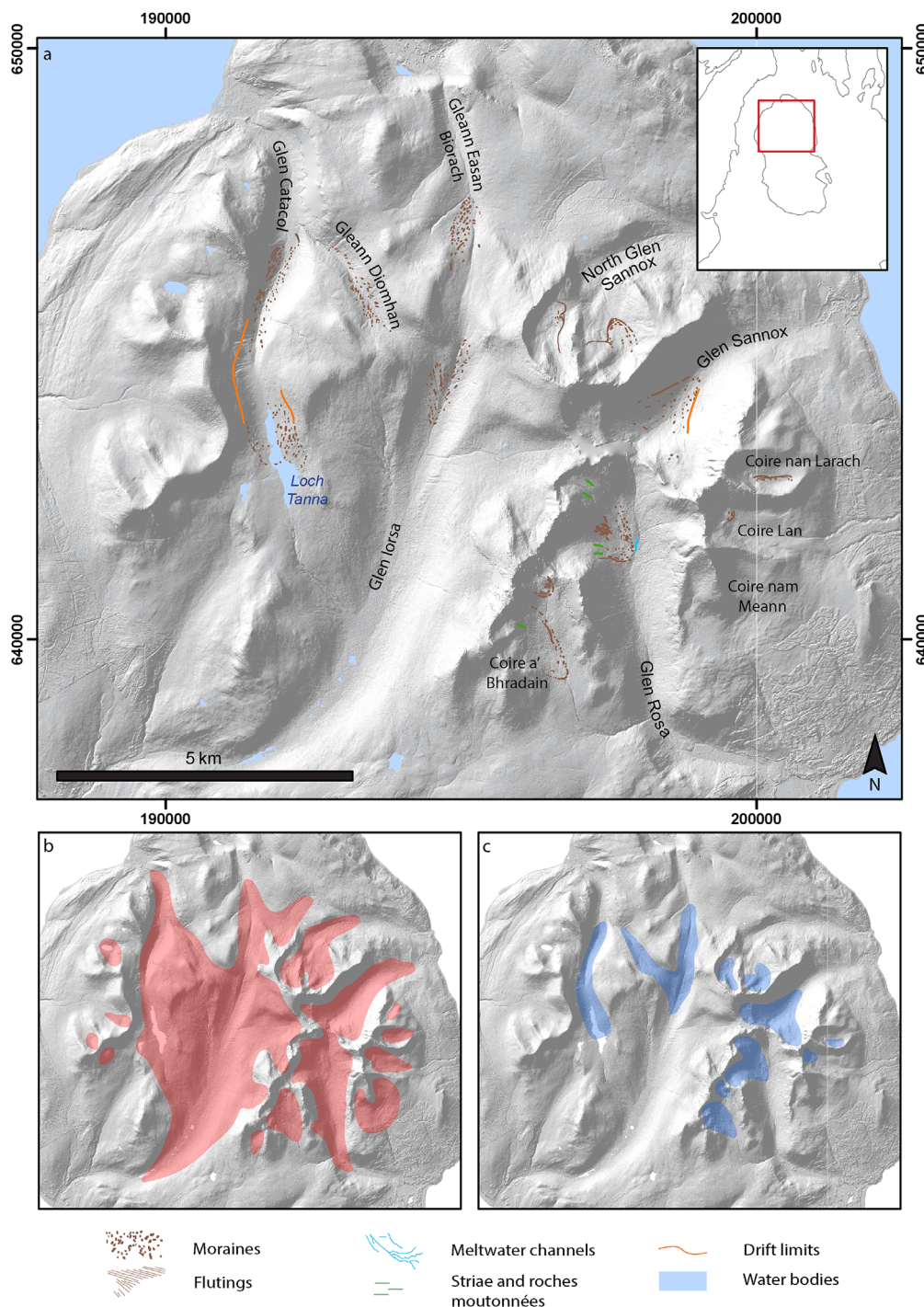


Figure 8. (a) Geomorphological mapping compiled for the glacial map for the Isle of Arran (after Ballantyne, 2007a). The extent of LLS glaciation on Arran as reconstructed by Gemmell (1973) (b) and Ballantyne (2007a) (c). Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC EarthObservation Data Centre.

Dearg massif (Fig. 11). Cirque glaciers formed in topographically favourable locations such as the cirques of An Teallach (Sissons, 1977b; Ballantyne, 1987) (Fig. 12). Although chronological control remains fairly sparse within this region, dating in specific locations has been used to determine the extent of LLS glaciation (Bradwell, 2006; McCormack *et al.*, 2011). Landforms relating to the Wester Ross Readvance, an ice sheet oscillation that occurred between approximately 18.0 and 16.5 ka (Everest and Kubik, 2006), are present at sites within this region.

Ben Hee

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 remote far NW Highlands was undertaken

(Sissons, 1977b). 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 (1977b) argued that the valley and cirque glaciers were contemporaneous and reconstructed a series of 70 LLS glaciers in the NW Highlands, 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-km² mountain icefield centred around Ben Hee (Lukas and Bradwell, 2010), covering five times the area of Sissons' (1977b) glaciers (Fig. 9).

The two models of glaciation of this area are in close agreement for the northern sector of the icefield, where both Sissons (1977b) and Lukas and Bradwell (2010) reconstructed glaciers draining the eastern cirques of Foinaven and coalescing in the valley below and then flowing northwards (Fig. 9).

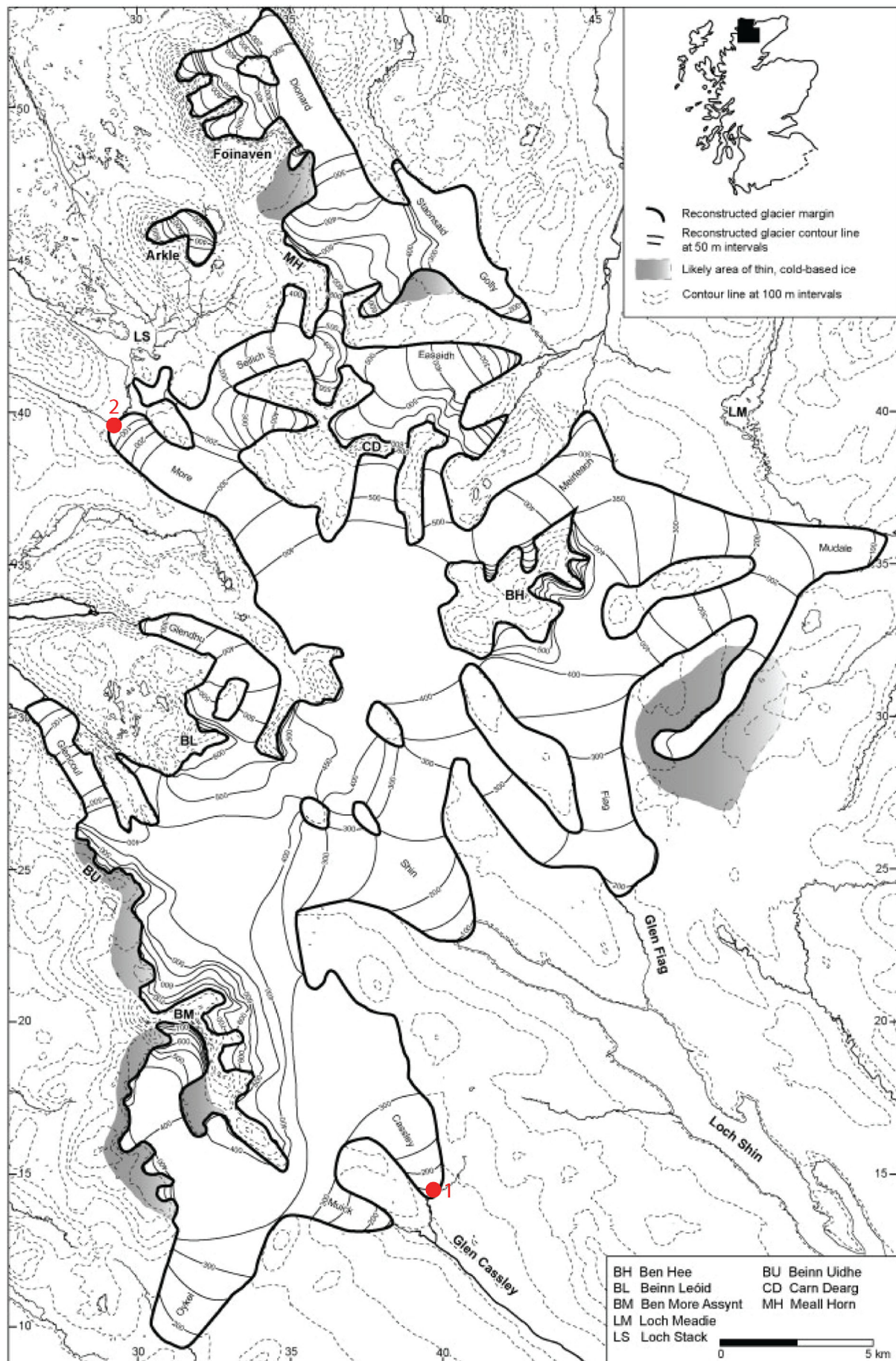


Figure 9. Three-dimensional reconstruction of the mountain icefield in the Ben Hee area. Chronological sites discussed in the text are indicated by the red dots. Adapted from *Journal of Quaternary Science* Vol. 25(4), Lukas, S. and Bradwell, T. Reconstruction of a Lateglacial (Younger Dryas) mountain icefield in Sutherland, north-western Scotland, and its palaeoclimatic implications, 567–580, Copyright (2010), with permission from John Wiley & Sons.

However, the presence of moraines beyond Sissons' (1977b) limits and the direction of ice flow inferred from roches moutonnées was thought 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 (Fig. 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 (1977b) to have been almost entirely ice-free. The area is dominated by clear sequences of moraines that often extend the entire length of the valleys and usually end in isolated terminal moraines. The mounds and ridges which fill these valleys are usually 5–15 m in height and are clearly arranged in arcuate chains which trend obliquely down-valley (Lukas and Benn, 2006; Lukas and Lukas, 2006a,b) (Fig. 10). Sections in these features show evidence of deformation from ice readvance after initial moraine formation in two-thirds of those features sampled (Benn and Lukas, 2006). 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 (1977b) used the extent of hummocky moraine and end moraine fragments to reconstruct four glaciers on the low ground north-east of the ridge formed by Ben More Assynt and Beinn Uidhe (Fig. 9). Lawson (1986) made only minor changes to these limits. However, subsequent mapping reported the presence of moraines beyond these limits, for example at the head of Loch Glencoul

(Bradwell, 2006) (Fig. 9). It was inferred 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 (Fig. 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 ages of 13.0 ± 1.1 (0.9) and 11.6 ± 1.0 (0.9) ka (Lukas and Bradwell, 2010) (Chronological Site 1, Fig. 9). This accords with radiocarbon dates of 11 247–10 698 cal a BP from the proglacial delta at the terminus of the More glacier (Chronological Site 2, Fig. 9) and with the presence of Lateglacial tripartite sequences in eight lochs just beyond the limits (Lukas and Bradwell, 2010).

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' (1977b) original account, but a LLS age for them has not been refuted in subsequent literature. These glaciers, which are marked by end and lateral moraines that enclose areas of hummocky moraine, are generally small, have northerly to easterly aspects and are 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; impressive

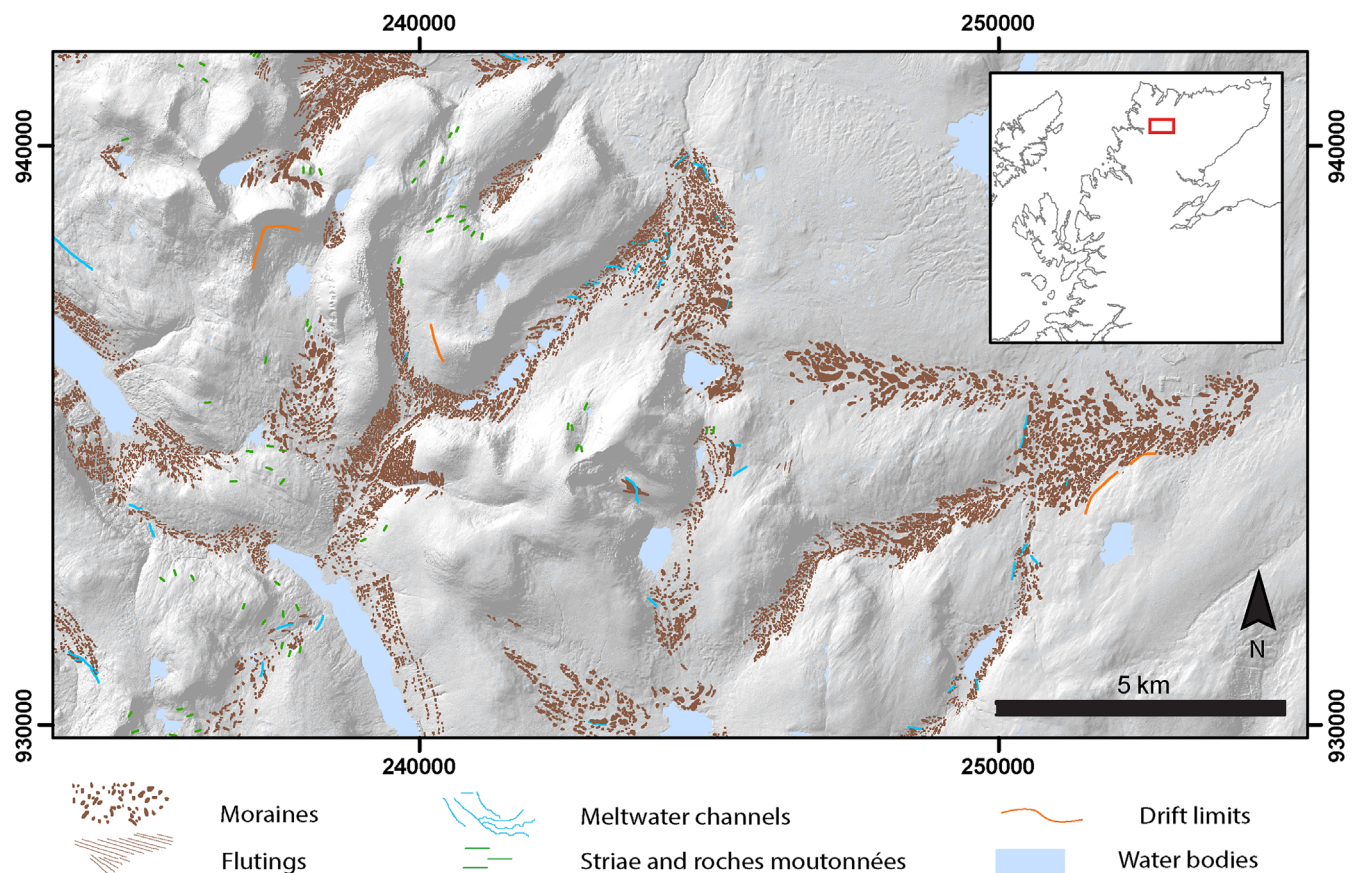


Figure 10. Geomorphological mapping of the central section of the Ben Hee icefield (after Lukas and Lukas, 2006a,b) showing the dense coverage of recessional moraines formed by extremely active retreat of these glaciers. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

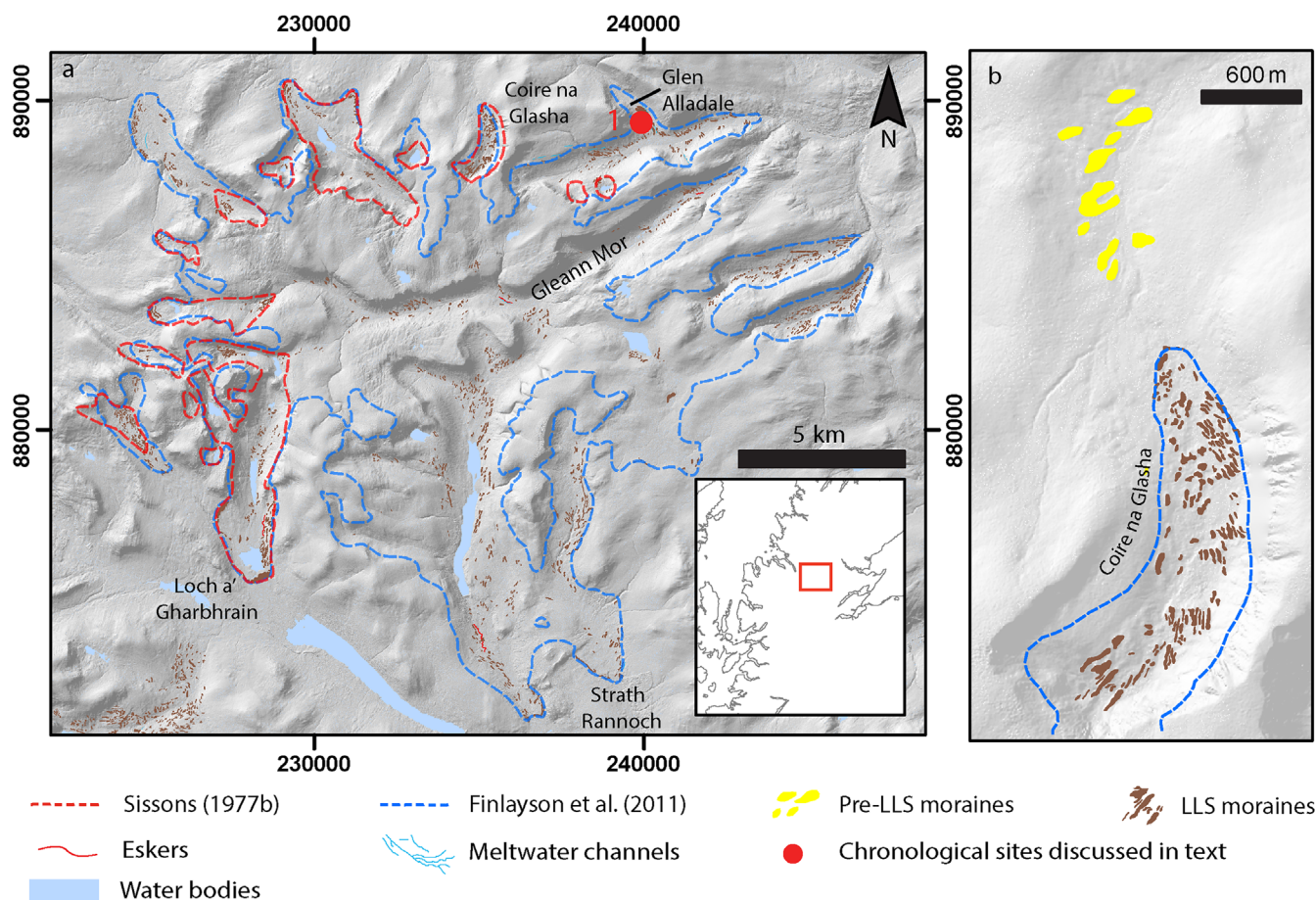


Figure 11. (a) The glacial landforms of the Beinn Dearg massif (after Finlayson *et al.*, 2011). The presence of moraines beyond the limits proposed by Sissons (1977b), as indicated in red, has led to reconstruction of a much more extensive ice cap and outlet glaciers (Finlayson *et al.*, 2011) (in blue). (b) The contrast between the smaller, more sharp-crested recessional moraines attributed to the LLS and the larger more subdued features thought to pre-date the stadal. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

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, 1977b; Lawson and Ballantyne, 1995; Lukas, 2006) (Fig. 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 the origin of landforms has 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.

Beinn Dearg

During the LLS, an ice cap developed on the Beinn Dearg massif (Fig. 11). 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 (1977b) suggested that just 12 cirque glaciers and a further two small valley glaciers had formed in the area (Fig. 11a). Recent geomorphological mapping of Glen Alladale (Fig. 11a) suggested that this earlier reconstruction was an underestimation (Finlayson and Bradwell, 2007), and Finlayson *et al.* (2011) reconstructed a highly asymmetrical LLS ice cap drained by extensive outlet glaciers to the north and south-east, but with only small cirque glaciers in the western sector (Fig. 11a).

The valleys occupied by ice from the Beinn Dearg ice cap are characterized by an abundance of closely spaced, recessional hummocky moraines, usually between 2 and 5 m high (Finlayson *et al.*, 2011). The coverage of moraines is not continuous throughout the valleys and the termini of the glaciers are frequently marked by an abrupt transition between these moraines and larger, more subdued features in the lower valleys (Fig. 11b). An exception occurs at Loch a' Gharbhrain (Fig. 11a), which is dammed by an impressive terminal moraine, Cnoc a' Mhoraire, some 800 m long, 200 m wide and up to 25 m high (Sissons, 1977b; 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 (Fig. 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 (Finlayson *et al.*, 2011). Cosmogenic isotope dating of boulders on the moraine complex in a northern tributary of Glen Alladale (Chronological Site 1, Fig. 11a) gave ages between 13.9 ± 0.8 (0.5) and 12.0 ± 0.8 (0.6) ka and indicated that the whole valley was glaciated during the LLS (Finlayson and Bradwell, 2007).

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 probably an important control on glacier build-up (Finlayson *et al.*, 2011).

An Teallach

Six small LLS glaciers developed in the north- and east-facing cirques of the An Teallach massif (Fig. 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, 1977b; 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 Fig. 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 confirm that the cirque glaciers represent a later readvance of probable LLS age (Ballantyne, 1987).

Applecross

Independent glaciers existed in the hills of the Applecross Peninsula but the extent of these glaciers has been debated. Robinson (1977, 1987) reconstructed three valley glaciers flowing south-eastwards from the Applecross Hills into Loch Kishorn (Fig. 12), the largest of which deposited an arcuate boulder belt in the loch, and possibly reached the southern shore of the loch. 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 that trimlines in the valley heads showed 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 between 13.4 ± 0.8 (0.5) and 11.0 ± 0.6 (0.4) ka (Chronological Site 1, Fig. 12), indicating that the site was glaciated 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 pre-LLS ages 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 pre-date 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.

West Highland Glacier Complex

The West Highland Glacier Complex (Figs 1, 2 and 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 while it is possible to reconstruct the lateral extent of the glaciers in the Rannoch Moor area, it is difficult to infer ice dynamics from Thorp's (1984) smaller-scale maps. Generally, mapping coverage of the southernmost section remains poor and although the termini of the Lomond and Menteith lobes (Figs 2 and 15) 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.

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 (Fig. 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 (Fig. 2). Sissons (1967a) presented a much smaller LLS icefield confined to the Western Grampian Mountains. Geomorphological data for much of the area between Torridon and Loch Linnhe were 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.

The 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 (Fig. 12). The greatest difference between Bennett's (1991) reconstruction and previous interpretations was the inclusion of LLS glaciers at Achnasheen (Fig. 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). Sissons (1982) argued that these features pre-dated the LLS because the required configuration of ice did not match reconstructed LLS glacier limits to the west. However, Bennett (1991) suggested that the Achnasheen moraines could be traced continuously westward to those limits proposed by Sissons (1977b) 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, 1993a,b) (Fig. 12). This interpretation is further supported by the absence of Lateglacial sediments from within the limits of the Achnasheen glaciers (Sissons, 1982).

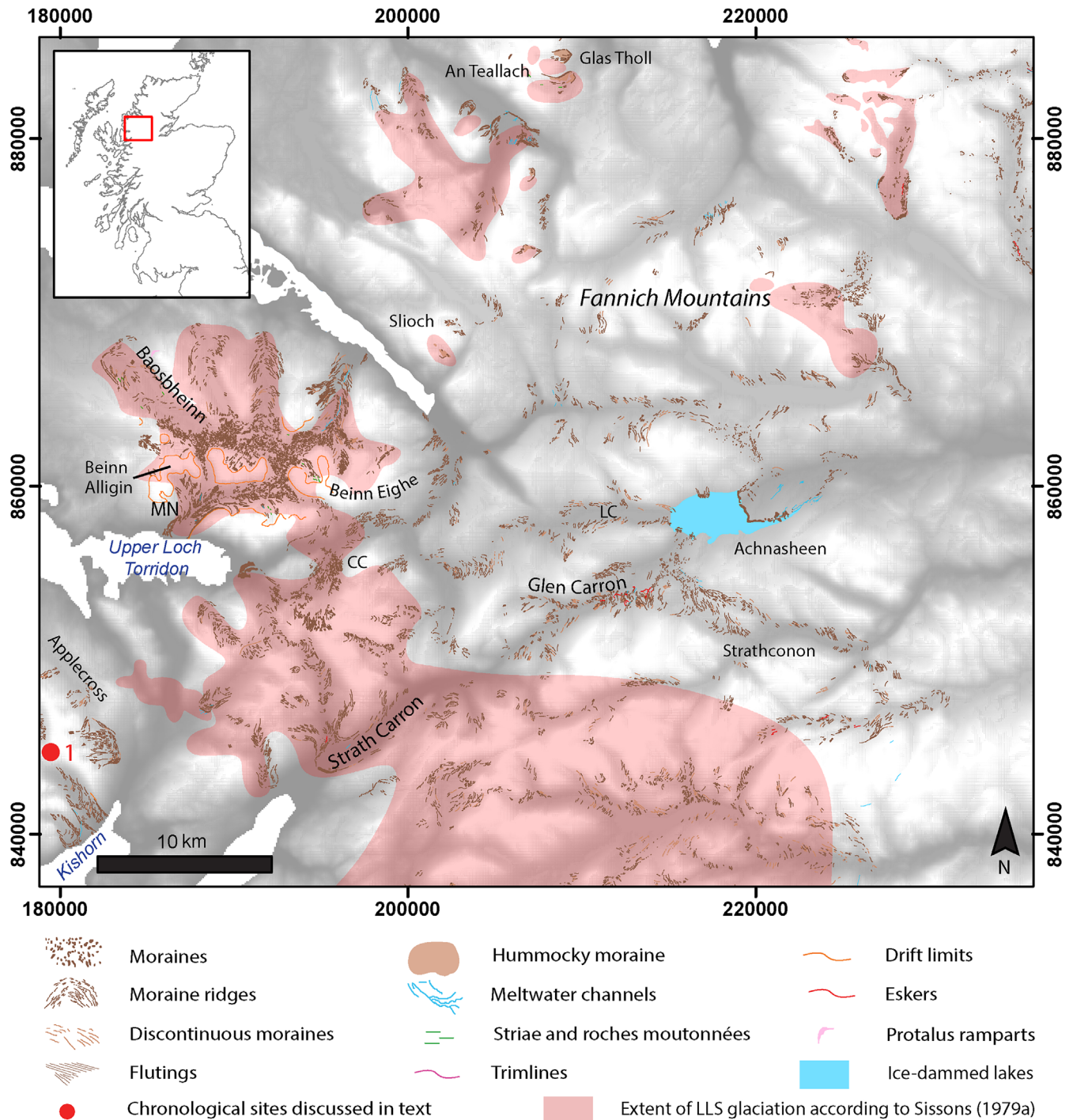


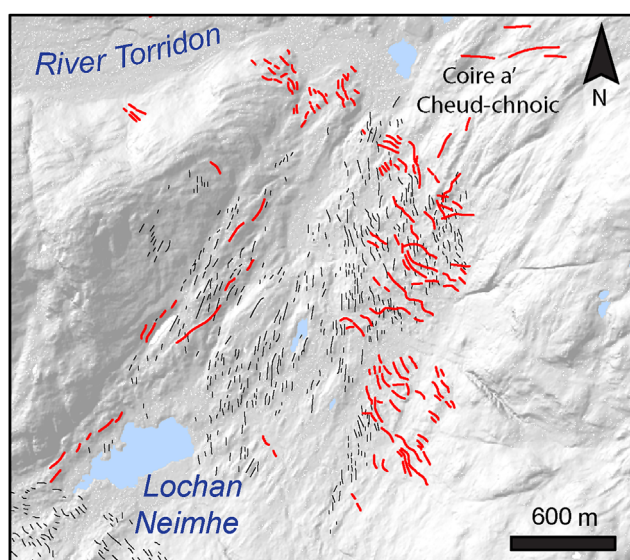
Figure 12. The northern sector of the West Highland Glacier Complex (after Sissons, 1977b, 1982; Ballantyne, 1986, 1987; Bennett, 1991; Bennett and Boulton, 1993b; Wilson and Evans, 2000; Wilson, 2005; Finlayson *et al.*, 2011; McCormack, 2011). Sequences of recessional moraines extend well beyond the 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. LC: Loch a' Chroisg, MN: Coire Mhic Nobuil, CC: Coire a' Cheud-chnoic. GB SRTM Digital Elevation Model from ShareGeo, available at www.sharegeo.ac.uk/handle/10672/5. Original dataset from NASA.

The reinterpretation of the Achnasheen features as being of LLS age led to landforms in the valleys south of this site also being assigned to the stadial. Bennett (1991) and Bennett and Boulton (1993a,b) added a large lobe of ice in the Strathconon area (Fig. 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 (Fig. 2). The extents of these eastern glaciers were reconstructed predominantly from sequences of recessional moraines on the valley sides and floors, although Tipping *et al.* (2003) attributed the limited

depositional evidence in Glen Affric to fast flowing ice and highlighted the conflicting nature of trimline evidence in the valley. The limits of the Glen Moriston glacier are well preserved, comprising end and lateral moraine ridges, meltwater channels and associated areas of 'water-washed' 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) (Fig. 12). Both the north- and the south-flowing glaciers in this area formed clear sequences of recessional moraines (Bennett and Boulton, 1993b), with particularly clear examples in Coire Mhic Nobuil (Fig. 12) where the contribution of debris from fresh rock 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') (Fig. 13) is covered by flutings parallel to the valley axis, the asymmetrical 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



- Hummocky ridge (fluting)
- Moraine ridge

Figure 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 pre-date 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.

ridges. Although these features have previously been attributed to ice stagnation processes (Sissons, 1967b; Eyles, 1983), the relationship of the flutings and transverse ridges suggests that the features were formed by a LLS glacier overriding older cross-valley, ice sheet retreat moraines (Hodgson, 1987; Wilson and Evans, 2000). The spread of recessional moraines in the north-western sector of the West Highland Glacier Complex implies that active retreat continued throughout deglaciation (Bennett and Boulton, 1993a). Evidence of glacial stagnation is uncommon; esker fragments and irregular mounds of sand and gravel are present in Glen Ling and Strath Carron but suggest that ice stagnation only occurred in small areas at the margins of actively retreating glaciers (Bennett, 1991).

Further south, particularly along the west coast, the margins of the West Highland Glacier Complex are less clear (Fig. 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 (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 Hourm (Fig. 2) extending 3.5 km beyond the proposed LLS limit. They suggested that the outermost ridges had only been occupied briefly by a tide-water glacier margin, with the glacier then retreating rapidly to stabilize at the main moraine sequence within the basin (McIntyre *et al.*, 2011). The age of these features is not made explicit but it is emphasized that onshore evidence may underrepresent the extent of these outlet glaciers, and suggested that the ridges 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, although these features remain undated.

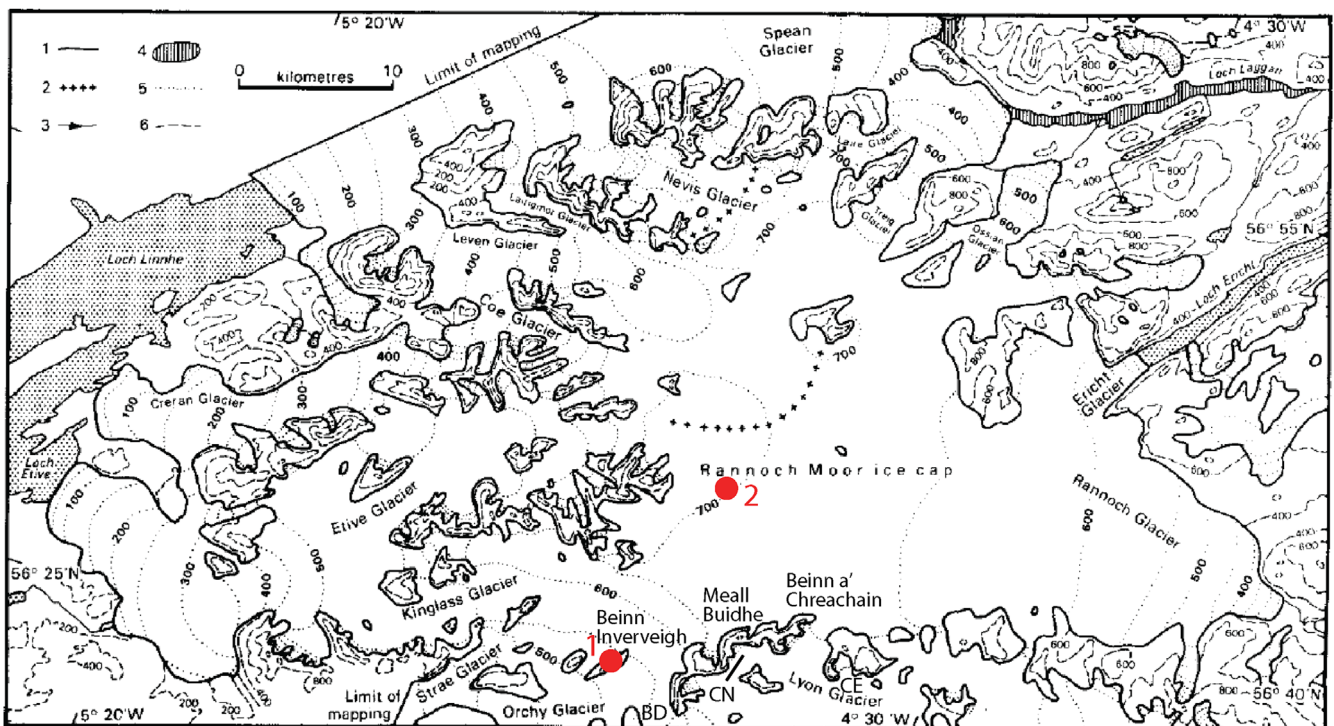
The patchy nature of the evidence along the west coast and the potential that many of these glaciers extended offshore complicates reconstructing the maximum extent of this section of the West Highland Glacier Complex. While numerical modelling suggests that the Hourm and Nevis glaciers terminated at 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.

Rannoch Moor

The glaciers in the north-west highlands coalesced across the Great Glen with ice nourished in the western Highland mountains and Rannoch Moor to the south (Fig. 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 (Thorpe, 1984; Turner *et al.*, 2014) although much of the area, particularly in the west, has not been mapped in detail since the 1980s.

The vertical extent of the glaciers in this area has been debated 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 900 m a.s.l., although this value was extrapolated from the limits of moraines in the valleys. Horsfield (1983) suggested that the ice cap had extended above 1000 m a.s.l. 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 a.s.l. (Fig. 14).

To resolve this uncertainty, 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 (Fig. 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 pre-dates the LLS as the neighbouring summit of Meall Buidhe (970 m) has clearly been frost-weathered (Golledge and Hubbard, 2005). Similarly, the presence of erratics from Beinn Heasgarnich (approximately 4 km south of Fig. 14) 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- to 3-m-high boulder moraines were observed in Coire nan Clach at 820–830 m a.s.l. (Golledge and Hubbard, 2005), supporting an ice surface elevation of approximately 900 m a.s.l. in this area. Cosmogenic exposure dating of glacial erratics and bedrock on Beinn Inverveigh (Chronological Site 1, Fig. 14) between 623 and 581 m a.s.l. gave recalibrated ages between 15.0 ± 1.1 (0.8) and 12.8 ± 1.0 (0.9) ka (Golledge *et al.*, 2007). These dates are older than would be expected if this area had been glaciated during the LLS, despite giving apparent LLS ages in the original publication when different production and erosion rates were used (Golledge *et al.*, 2007).



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In the area between Glen Lyon and Glen Lochy, Golledge (2007) observed several landsystem elements diagnostic of topographically unconstrained ice caps. These included the oblique alignment of moraines across valleys, asymmetrical 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 (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. 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).

During the LLS, outlet glaciers from the main icefield advanced into Glen Roy and Glen Spean (Fig. 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. Recalibration of cosmogenic nuclide dates from bedrock on the 325-m shoreline gave formation ages of between 12.9 ± 0.8 (0.6) and 12.3 ± 1.3 (1.2) ka (Fabel *et al.*, 2010). These dates suggest that the Spean glacier may have reached its maximum position during the mid-stadial, rather than at its end, as suggested by Fabel *et al.* (2010) when using different production rates. However, Palmer *et al.* (2010) used the Lochaber Master Varve Chronology to estimate, by adding the number of varves (515) to the date of the onset of Holocene warming, 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. 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. 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) (Chronological Site 2, Fig. 14), the proposed centre of the icefield. The oldest replicable age for the colonization of the area by vegetation was 12 613–11 821 cal a BP, indicating that ice reached its maximum extent in this area 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 north-west Highlands, or advanced before the onset of the stadial and retreated under warmer summer conditions during the stadial. This interpretation conflicts with all previously published research on the LLS in Britain, presents limited geomorphological mapping and relies on a relatively small number of radiocarbon dates. More recently, cosmogenic isotope dating of boulders situated on the crest of a moraine which impounds Bromley *et al.*'s (2014) core sites gave best estimate ages between 11.7 ± 0.6 (0.3) and 11.3 ± 0.6 (0.3) 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 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). 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.

Loch Lomond and the Trossachs

The south-west sector of the main icefield remains understudied and little published literature on the LLS glaciation of this area is available. Sutherland (1981) mapped glacial features at specific sites within the area, but focused primarily on the raised shorelines and much of the area still lacks detailed geomorphological mapping. At Ardentinny, on the shores of Loch Long (Fig. 2), and at Furnace and Strathlachlan, on Long Fyne, hummocky moraine and outwash mark the termini of glaciers in these sea lochs (Sutherland, 1981). However, 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 these limits. Sutherland (1981) identified periglacial features above approximately 750 m a.s.l. on Beinn Ime and 800 m a.s.l. on Beinn Vorlich, suggesting that these summits formed nunataks above the surrounding ice. Small cirque glaciers formed at Black Craig, Corrachave, Corarsik, Stronlonag and Creag Mhor on the Cowal Peninsula (Fig. 2), as inferred from end moraines and cross-cutting striations. However, these features were not mapped by Sutherland (1981) and no glacial reconstruction was presented.

With the exception of the Menteith and Callander lobes, the glaciers which drained the south-east sector of the West Highland Glacier Complex have been largely overlooked in the published literature. LLS glaciers occupied Glens Lochay and Dochart and almost coalesced at the head of Loch Tay (Wilson, 2005). 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 been proposed that a glacier developed in Glen Almond, based on the 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 mean this reconstruction is equivocal. Indeed, the glacial reconstruction for much of the area between the Rannoch and Callander lobes relies entirely on the limits proposed by Thompson (1972) and represents a region which 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 shoreline in Gare Loch and Loch Long (Fig. 15). The maximum extent of the Lomond glacier is demarcated by a continuous belt of well-developed moraine ridges which enclose an area of widespread till hummocks (Rose, 1981) (Figs 6b and 15). Along the loch's western shore, large areas of outwash and eskers indicate glacier thinning and the production of large volumes of

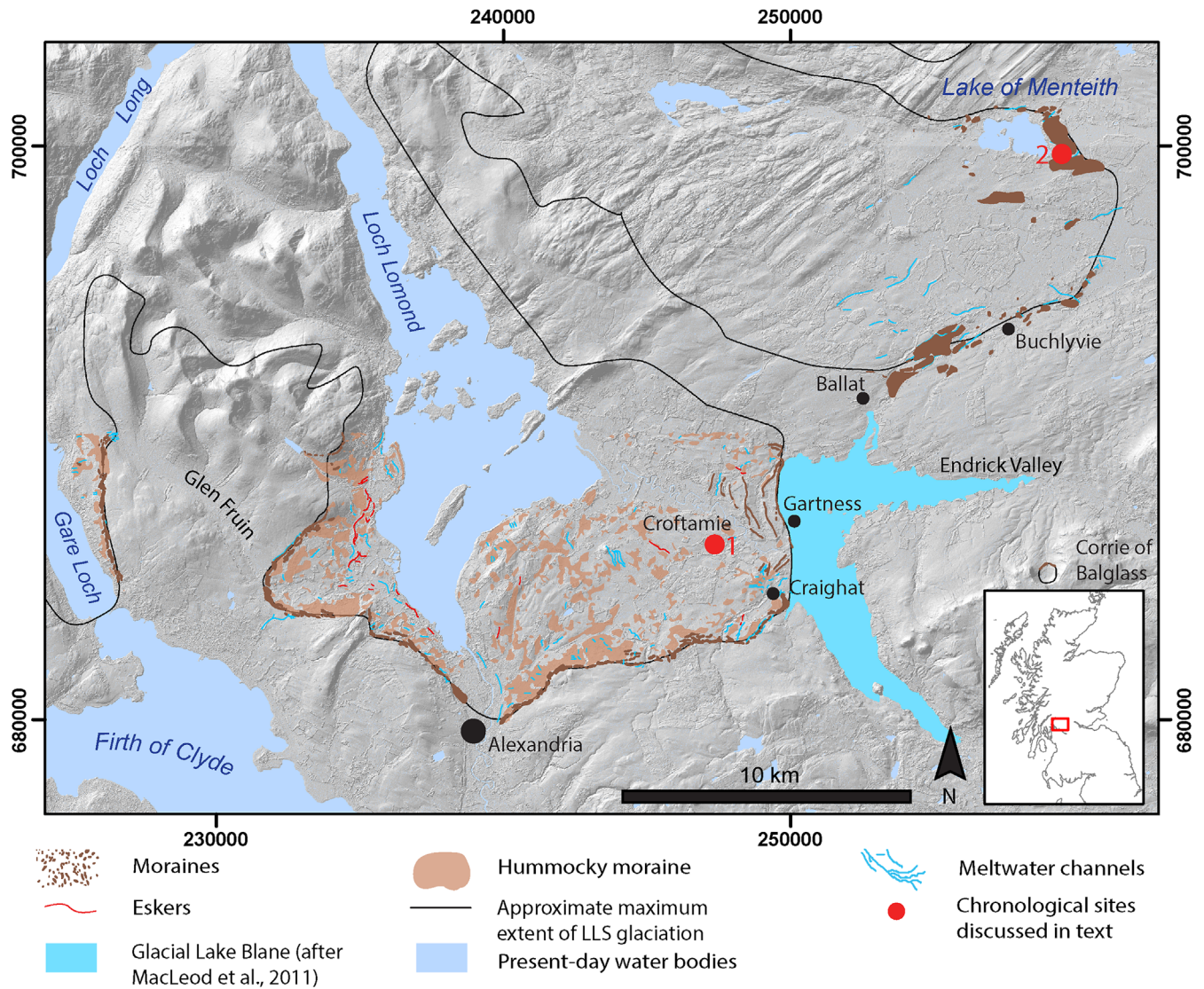


Figure 15. The geomorphology of the Gare Loch, Lomond and Menteith ice lobes (after Rose, 1980, 1981; Wilson, 2005). The approximate LLS glacier extent is included to demonstrate the restriction of mapped geomorphological evidence to the termini for these glaciers. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

meltwater and glaciofluvial sediment (Rose, 1980). This terrain contrasts with that found outside the limit, which is characterized by north-west to south-east aligned drumlins formed by the Dimlington Stadial ice sheet and periglacial features on steeper slopes. Near the terminus of the Lomond glacier (Fig. 15), the ice flowed a short distance down Glen Fruin while the majority inundated the lowland area south of the loch, terminating in a belt of moraine ridges which runs between Alexandria and Craighat. The limit then turns sharply north towards Gartness, comprising a series of six recessional moraine ridges (Rose, 1981). The outermost ridge changes along its length from a flat-topped feature formed of laminated silts and glaciofluvial sands and gravels in the south to a sharp-crested ridge of till in the north (Rose, 1981) (Figs 6c and 15). The southern part of this moraine was formed subaqueously in a proglacial lake formed when the advancing 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).

The glacial history of the area has been well recorded in several stratigraphic exposures, including the official type site for LLS glaciation at Croftamie (Rose *et al.*, 1988; Rose and Lloyd-Davies, 2003) (Chronological Site 1, Fig. 15). The

stratigraphic succession for the whole basin (Table 1) comprises: the basal LGM Wilderness Member/Till; overlain sequentially by the Killearn Member (glacier lake sediments); Lateglacial glaciomarine Clyde Beds (Linwood and Paisley Members); a layer of organic detritus (Croftamie Member), radiocarbon-dated at 12 127–11 624 cal a BP (MacLeod *et al.*, 2011); topped by Loch Lomond Stade 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 which was formed during deglaciation by ice damming the drainage of the Endrick and Blane valleys (Fig. 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 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).

Table 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 30 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 the Quaternary Research Association.

Rose (1980)	Hall et al. (1998)	Sutherland (1999)	
Blane Valley Silts	Blane Water Formation	Blane Member	Clyde Valley Formation
Rhu Sands and Gravels	Drumbeg Formation	Drumbeg Member	
Gartocharn Till	Gartocharn Till Formation	Gartocharn Member	
Blane Valley Silts	Blane Water Formation	Blane Member	
Organic detritus and palaeosol		Croftamie Member (12127–11624 cal a BP)	
	Killearn Formation	Killearn Member	
Clyde Beds	Linwood & Paisley Formation	Linwood & Paisley Members	
Gartness Silts	? Killearn Formation (part)	? Killearn Member (part)	
Wilderness Till	Wilderness Till Formation	Wilderness Member	

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 dates are consistent with those from Lochaber, and potentially suggest that larger glaciers, such as the Spean and 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).

A small glacier developed in Corrie of Balglass, a northern-facing cirque in the Campsie Fells (Brown, 2003; Brown and Rose, 2003; Brown *et al.*, 2007). The limits of this glacier are marked by an outer ridge, interpreted as a terminal moraine, inside of which discontinuous parallel ridges, thought to be recessional moraines, are present (Brown, 2003). Coring of basins within the moraine belt revealed a 'diamicton, overlain by minerogenic and organic lake sediments and finally peat' (Brown, 2003, p. 179) and only one period of warming was represented in the pollen record, suggesting a LLS age for the glacier. Glaciological modelling by both Brown (2003) and Carr and Coleman (2007a) suggested the reconstructed glacier could have existed under LLS climate conditions, although the latter suggested this glacier was only marginally viable. It is likely that the north-facing aspect of the cirque protected the glacier from insolation and that both snowblow and avalanching from the plateau behind nourished the cirque glacier (Brown, 2003).

The extent of the Menteith glacier is marked 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) (Fig. 15). 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). 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). Smith (1993) suggested that the ridge was a push moraine and Evans and Wilson (2006b) proposed that the 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 (Wilson, 2005). Such an explanation is entirely consistent with a glacier flowing over deformable sediments, such as the marine beds that 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 13 764–13 006 cal a BP (Sissons, 1967a) (Chronological Site 2, Fig. 15), indicating a LLS age for the formation of the moraine.

The apparent limit of LLS glaciation in the upper Teith Valley is marked by an arcuate terminal moraine approximately 2.5 km down-valley of Callander (Thompson, 1972) (Fig. 2). Stratigraphic analysis from a borehole 200 m outside the terminal moraine on the valley's southern side revealed 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 typical 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 15 454–14 910 cal a BP, indicating that the overlying till was deposited during the LLS (Merritt *et al.*, 1990, 2003). This notion is further supported by the transition from Lateglacial to LLS deposits in both the pollen and the coleopteran records. The former indicates that a substantial volume of sediment accumulated in this area after the onset of LLS conditions but before the readvance glacier overran the site (Merritt *et al.*, 1990). 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). The glacier then retreated slightly and stabilized to form the terminal moraine ridge. 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.

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 (Fig. 1). Although glacial landforms are widespread in this region, in the absence of reliable chronological control, the age of these landforms has been disputed. This has prompted 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 that 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 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) (Fig. 16).

Monadhliath

The glacial history of the Monadhliath Mountains (Fig. 17) remained largely unstudied until the last decade. Numerical modelling (Golledge *et al.*, 2008) indicated that ice had been nourished on the upland plateau, but the extent of glaciation remained unsupported by empirical evidence until the first systematic geomorphological mapping of the region by Boston (2012a,b). 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) (Fig. 17). 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 moutonnées 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, fresher ridges, or larger, more subdued features) that were then used to define the LLS glacier limits (Fig. 18).

Before 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 (Fig. 17). However, Boston *et al.* (2013) suggested that this feature and the moraines in the lower valley pre-dated 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 south-east of the icefield, Young (1978) reported that Gleann Lochain, Ballach and Chaorainn (Fig. 17) 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. However, it

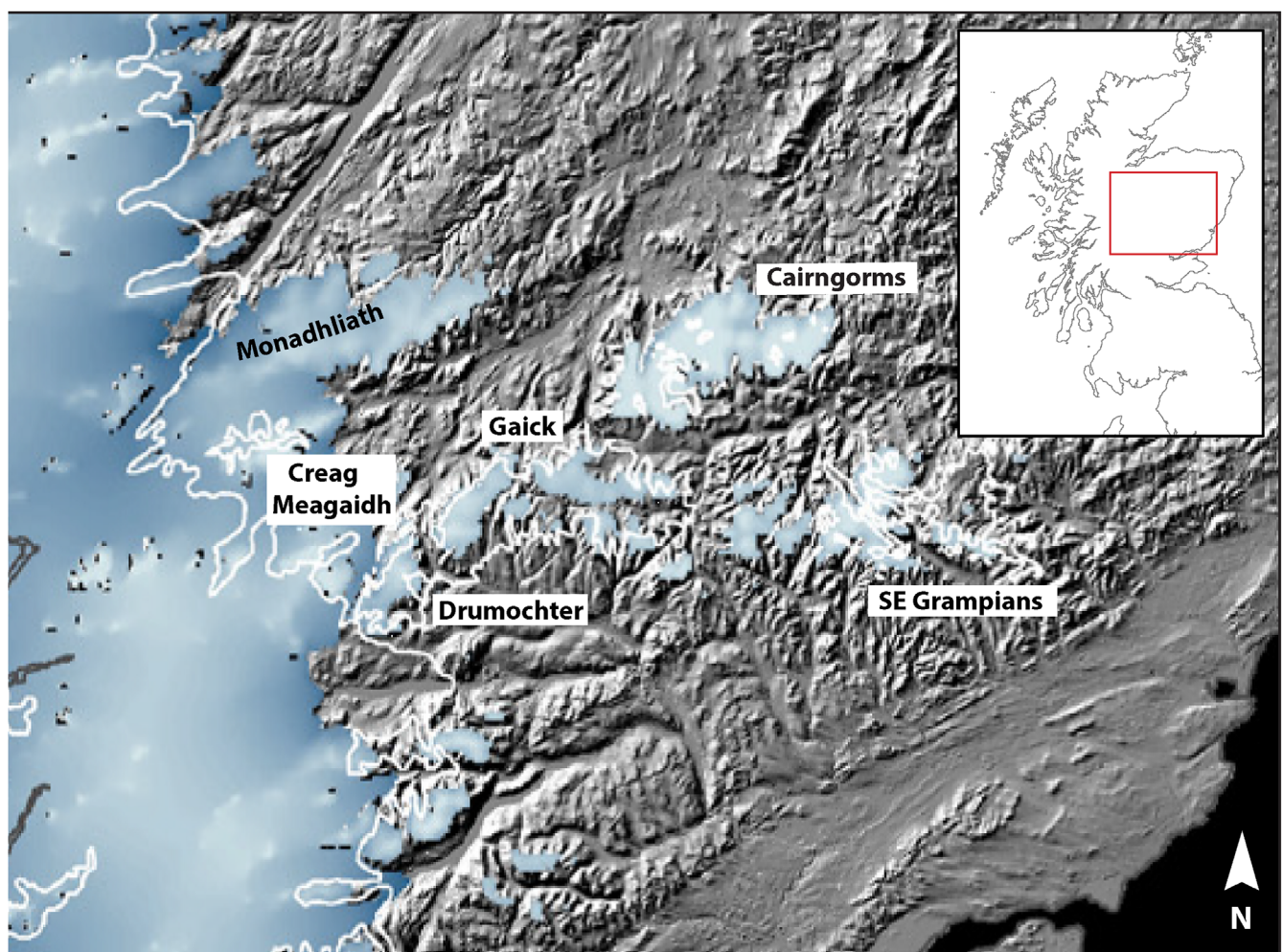


Figure 16. The extent of LLS glaciation in the Central Grampians, adapted from Golledge *et al.* (2008). The solid white line indicates the extent of empirically based reconstructions, superimposed over the results of numerical modelling (from Golledge *et al.*, 2008). The modelling supports a more extensive coverage of ice, although since publication an extensive icefield has been reconstructed on the Monadhliath Mountains which accords well with the modelled extent.

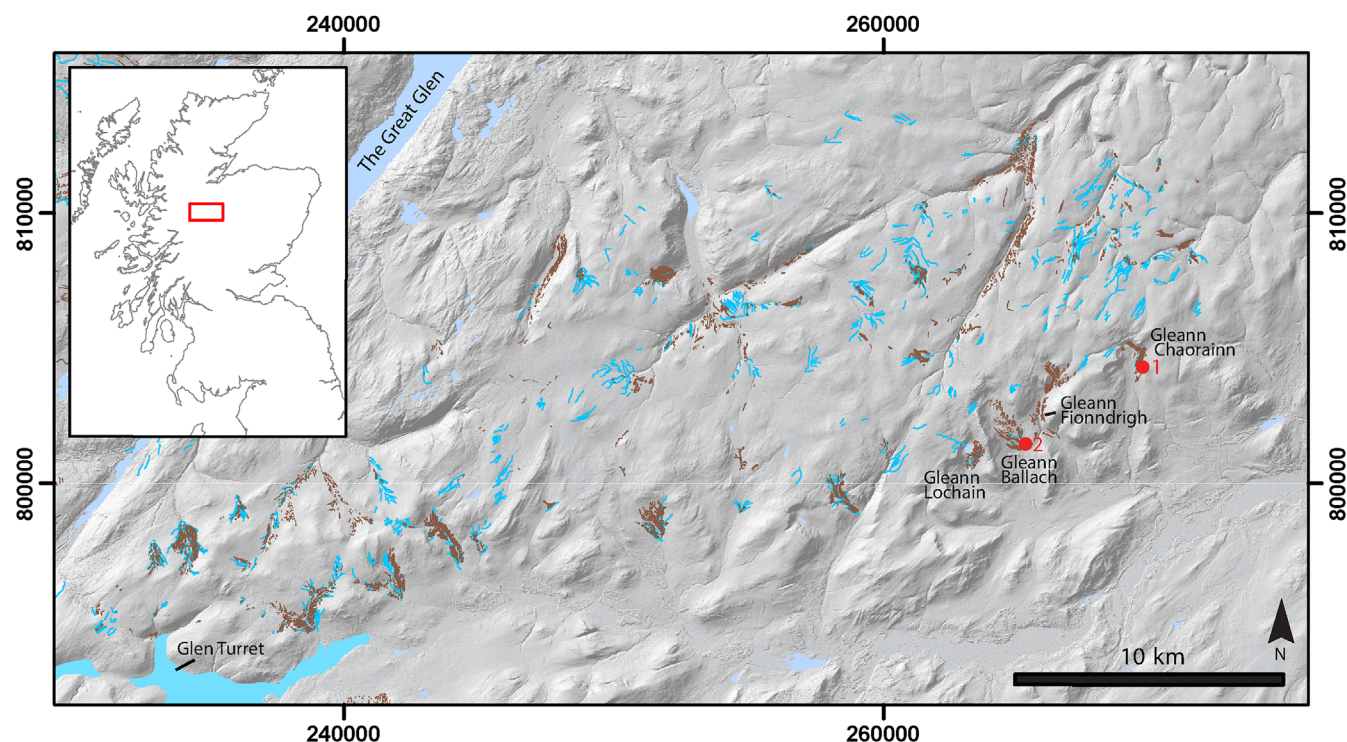


Figure 17. Geomorphological mapping of the Monadhliath Mountains (after Boston, 2012a). While recessional moraines are apparent in around one-third of the valleys, most of the icefield was reconstructed from ice-marginal meltwater channels. Legend in Fig. 15. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

was not possible 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 Gollidge, 2012).

The reconstructions presented by Boston *et al.* (2013), Trelea (2008) and Trelea-Newton and Gollidge (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 18.7 ± 0.6 (0.6), 16.3 ± 0.5 (0.5) and 13.4 ± 0.7 (0.4) ka (Gheorghiu *et al.*, 2012; Gheorghiu and Fabel, 2013) (Chronological Site 1, Fig. 17). These dates suggest that the valley was last glaciated before 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 Gleann Ballach

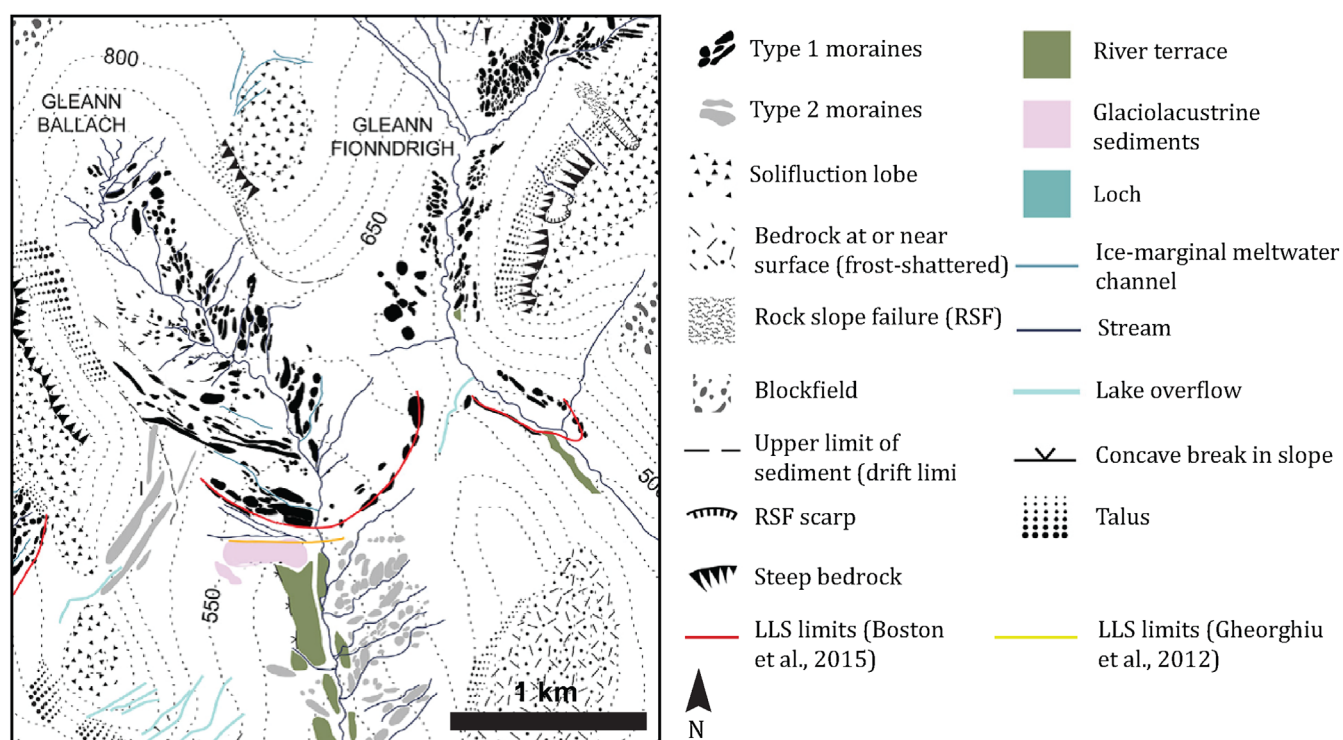


Figure 18. Geomorphological mapping of Gleann Ballach and Gleann Fionndrigh (adapted from Boston *et al.*, 2015). A morphostratigraphic approach was used to determine the limits of LLS glaciation.

(Figs 17 and 18) two concentric, arcuate end moraine ridges 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 with mean ages of between 11.2 ± 0.6 (0.4) and 12.7 ± 0.7 (0.4) ka, respectively (Gheorghiu and Fabel, 2013) (Chronological Site 2, Fig. 17), potentially indicating that LLS glaciation here occurred in two phases.

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 and future work to systematically date the landforms in these valleys is required.

Creag Meagaidh

The glacial geomorphology of the Creag Meagaidh massif (Fig. 16) 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 that 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-rai 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.

Drumochter and Gaick

Both the extent and the timing of LLS glaciation in the Gaick and Drumochter areas (Figs 16 and 19) have been debated. Charlesworth (1955) and Horsfield (1983) proposed that the entire area was covered by an extension of the West Highland Glacier Complex. Subsequently, Sissons (1974) mapped the area and used both the geomorphological evidence and glaciological principles to produce a detailed three-dimensional reconstruction of the ice mass, the first such reconstruction of LLS glaciers in Britain. Sissons (1974) reconstructed a more restricted ice cap on the Gaick Plateau (Fig. 16), the south-west outlet glaciers of which flowed down Coire Mhic-sith, and the neighbouring Allt a Stalacir', into the Drumochter Pass and, ultimately, coalesced with an icefield nourished on the West Drumochter Hills (Sissons, 1980a).

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 characterized 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 (Sissons, 1974). The features inside the limits contrasted with the much larger channels and major glaciofluvial deposits, including kames, found beyond the limits. Furthermore, large periglacial features were present around the peripheries of the plateau but are absent from within, suggesting that a substantial ice mass developed on the plateau during the LLS (Sissons, 1974). End moraines are very rare within the area, leading Sissons (1974) to infer that the glaciers reached their maximum extents after climatic amelioration had commenced and thus remained at these positions only briefly.

Merritt *et al.* (2004b) argued that Sissons' (1974, 1980a) reconstruction had implausible ice surface contours and accumulation centres and instead argued that the evidence around Drumochter represented south-westward ice sheet retreat. Since 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 pre-date 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). Ice in the Drumochter Pass probably dammed a lake in Coire Mhic-sith (Fig. 19), where glaciolacustrine sediments have been discovered (Lukas and Merritt, 2004; Benn and Ballantyne, 2005), and this is difficult to reconcile with Sissons' (1980a) model of coalescent ice between Drumochter and Gaick.

The valleys of the West Drumochter Hills are dominated by hummocky, recessional moraines that trend obliquely downslope, for example in Glen Garry and Glen Truim (Benn and Ballantyne, 2005) (Fig. 19). In each of these valleys, moraines and/or trimlines suggest that ice retreated up-valley towards a central source rather than to the south-west, as favoured by Merritt *et al.* (2004a,b). Moraines are less apparent in the south-west-facing valleys, a pattern which Benn and Ballantyne (2005) attributed to snow-blow away from these valleys. Terraces show the existence of a lake dammed by the Garry, Shallainn and Easan glaciers (Fig. 19), indicating that these lobes reached their maximum positions contemporaneously and that the Shallainn glacier began to retreat before the Garry glacier (Benn and Ballantyne, 2005). Such a retreat pattern indicates the deglaciation of an independent icefield in the area, rather than progressive south-westerly retreat of the last ice sheet. Absolute dates are limited in this area to a single kettle hole in the Drumochter Pass, where Holocene sediments with a basal date of between 11 387 and 9908 cal a BP were found (Walker, 1975) (Chronological Site 1, Fig. 19). The absence of mature periglacial features from inside the limits of the West Drumochter glaciers supports a LLS age.

Given the strength of the evidence for a LLS icefield in the Drumochter Hills, it seems implausible that glaciation on the Gaick would have been as restricted as suggested by Merritt (2004b). Likewise, given the likelihood that an extensive plateau icefield developed on the neighbouring Monadhliath, it is difficult to imagine that ice would not have accumulated on the similar topography of the Gaick. Numerical modelling suggests that substantial glaciers would have developed on the plateau (Golledge *et al.*, 2008), although less extensively than proposed by Sissons (1974). Given these uncertainties

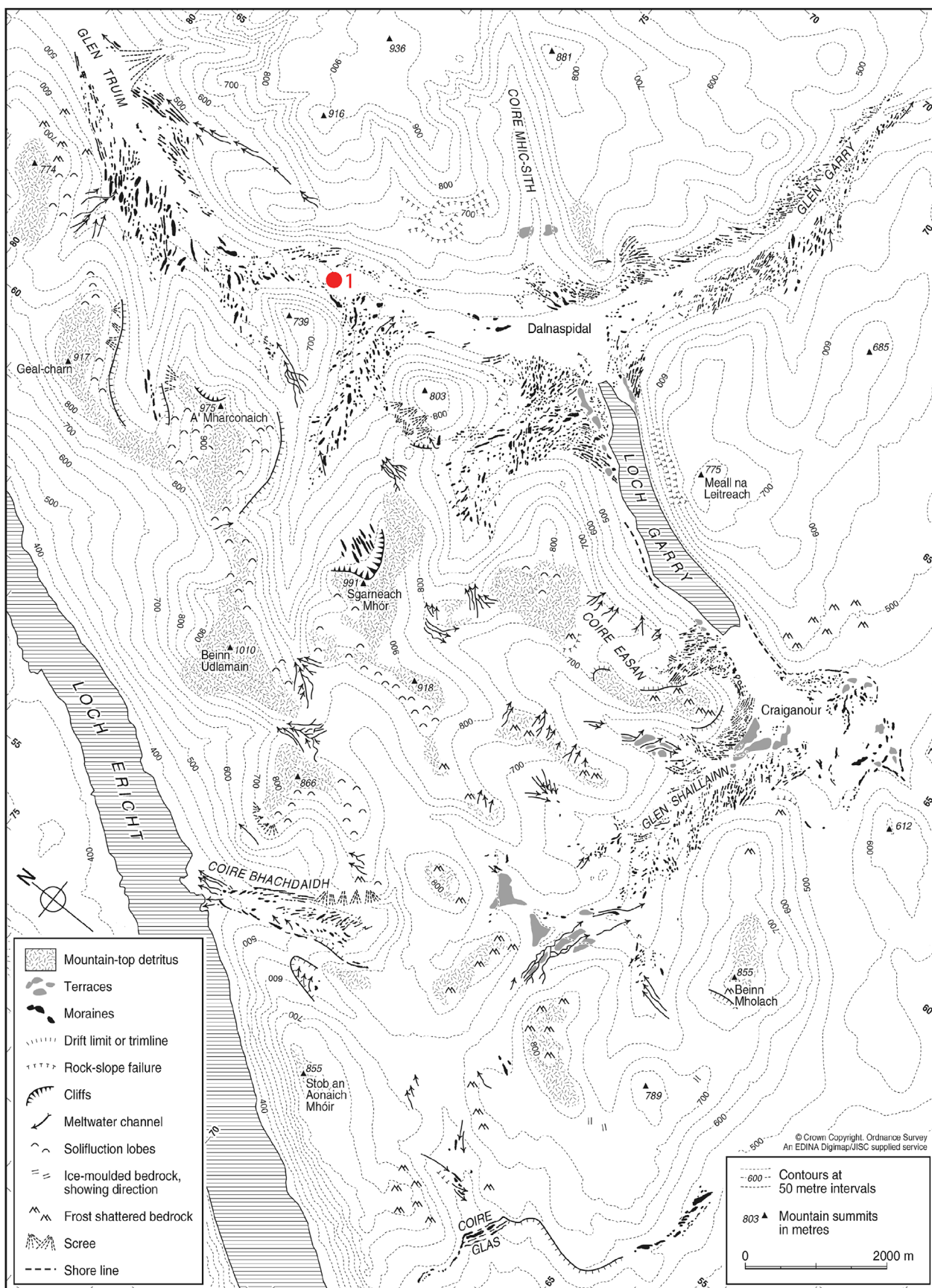


Figure 19. Geomorphological mapping of the West Drumochter Hills. Adapted from *Journal of Quaternary Science*, Vol. 20(6), Benn, D.I. and Ballantyne, C.K., Palaeoclimatic reconstruction from Loch Lomond Readvance glaciers in the West Drumochter Hills, Scotland, 577–592. Copyright (2005) with permission 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 south-westwards retreat of the last ice sheet.

and the advancements in glacial research in the 40 years since Sissons' (1974) study, the Gaick is a prime candidate for future mapping and dating work.

Cairngorms and south-eastern Grampians

In the Cairngorm mountains (Figs 16 and 20) it has proved difficult to distinguish between LLS and older glacial landforms and consequently the extent of LLS glaciation is uncertain. Sugden (1970) mapped the geomorphology, noting that many valleys contain topographically constrained drift deposits, usually arranged in distinct ridges that trend obliquely down the valley. He identified these features as eskers, based on their sandy composition and attributed them to ice stagnation at the close of the LGM, suggesting that during the late stages 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 (Fig. 20), where moraines were identified, before undergoing widespread stagnation (Sugden, 1970). Arcuate terminal moraines in eight favourably sited cirques within the area were suggested to represent a later readvance. The presence of Lateglacial sediments within an area of hummocky moraine at Loch Builg (18 km east of the area shown in Fig. 20) in the eastern Cairngorms (Clapperton *et al.*, 1975) demonstrated that this area had escaped LLS glaciation and suggested that other areas characterized by these landforms in the Cairngorms may have also remained unglaciated (Sugden and Clapperton, 1975).

The notion of LLS glaciation in the Cairngorms being restricted to eight cirques contrasts with Sissons' (1979c) subsequent remapping and glacial reconstruction (Fig. 20). The two proposed ice extents are broadly comparable for the eastern and central Cairngorms but major differences exist between reconstructions for the western hills. Sissons (1979c) proposed that valley glaciers had occupied Garbh Coire, Glen Geusachan and Glen Eidart (Fig. 20), which Sugden (1970) suggested nourished only small cirque glaciers. Arcuate terminal moraines mark the cirque glaciers, while sequences of hummocky moraines demarcate the three valley glaciers (Sissons, 1979c; Standell, 2014). Since 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 recessional moraine ridges that the Glen Geusachan glacier had undergone active retreat (Fig. 20).

Cosmogenic nuclide dating of boulders on lateral moraines at the outermost margin of the Glen Geusachan glacier (Chronological Site 1, Fig. 20) yielded ages between 18.6 ± 1.6 (1.4) and 13.6 ± 1.4 (1.3) ka (Everest and Kubik, 2006), indicating that this area escaped LLS glaciation. However, Standell (2014) argued that the features dated by Everest and Kubik (2006) were flat-topped pre-existing surfaces, through which meltwater channels had eroded, rather than moraines, and that the true LLS limit was slightly further up-valley, marked by the hummocky moraine below Devil's Point (Chronological Site 2, Fig. 20). Dates on these moraines, of between 12.1 ± 0.7 (0.4) and 10.8 ± 0.6 (0.3) ka, indicate 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 also reconstructed a previously unmapped glacier in Coire Etchachan (Fig. 20) 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 is supported by cosmogenic dates from the boulder sheet in Coire an Lochain (Kirkbride *et al.*, 2014) but this small glacier is not necessarily coeval with the larger valley glaciers.

Standell (2014) added a further three cirque glaciers on the flanks of Braeriach and a substantial volume of ice on the north-eastern slope of Ben Macdui (Fig. 20) but acknowledged that the cirque and valley glaciers could have been fed by additional plateau ice on Monadh Mor, Braeriach, Ben Macdui and Beinn a' Bhuid (approximately 7 km east of the area in Fig. 20). The cirque glaciers were predominately north to east facing, minimizing 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 south-east Grampian Mountains (Figs 1 and 16) have received minimal attention in the published literature. Sissons (1972), Sissons and Grant (1972) and Sissons and Sutherland (1976) reconstructed 27 LLS glaciers in this area, comprising an ice cap on the plateau area south-west of Mount Keen, valley glaciers in Glens Callater, Doll, Clova, Muick, Effock and West Water, and a collection of cirque glaciers. The extent of the valley glaciers is relatively well constrained by abundant hummocky moraine, particularly in the lower reaches of the valleys. Similarly, the majority of the cirque glaciers are clearly represented by terminal moraine ridges that often enclose areas strewn with large boulders or 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 to tentatively reconstruct its dimensions. This ice configuration represents a substantially larger ice mass in the south-eastern Grampians than in the Cairngorms, which Sissons (1979c) attributed to the dominant south-easterly winds causing a decrease in snowfall to the north-west.

Numerical modelling predicts substantial LLS glaciation in the Cairngorms [far exceeding even Standell's (2014) most extensive reconstruction], and overestimates glacier extent in the western hills, but fails to produce the large Glen Mark ice cap in the east (Golledge *et al.*, 2008). Given Standell's (2014) proposal that plateau glaciation may have been more extensive in the Cairngorms, the south-eastern Grampians merit further investigation to better constrain glacial reconstructions for this area, and to determine how it fits with updated understandings of regional trends.

Southern Uplands

LLS glaciation in the Southern Uplands occurred in the Galloway and Tweedsmuir Hills. The former was characterized 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.*, 2013, 2014).

Galloway

The extent of LLS glaciation in the Galloway Hills was restricted to 11 cirques in the area, seven in the Merrick Hills and four in the Rhinns of Kells (Cornish, 1981) (Fig. 1).

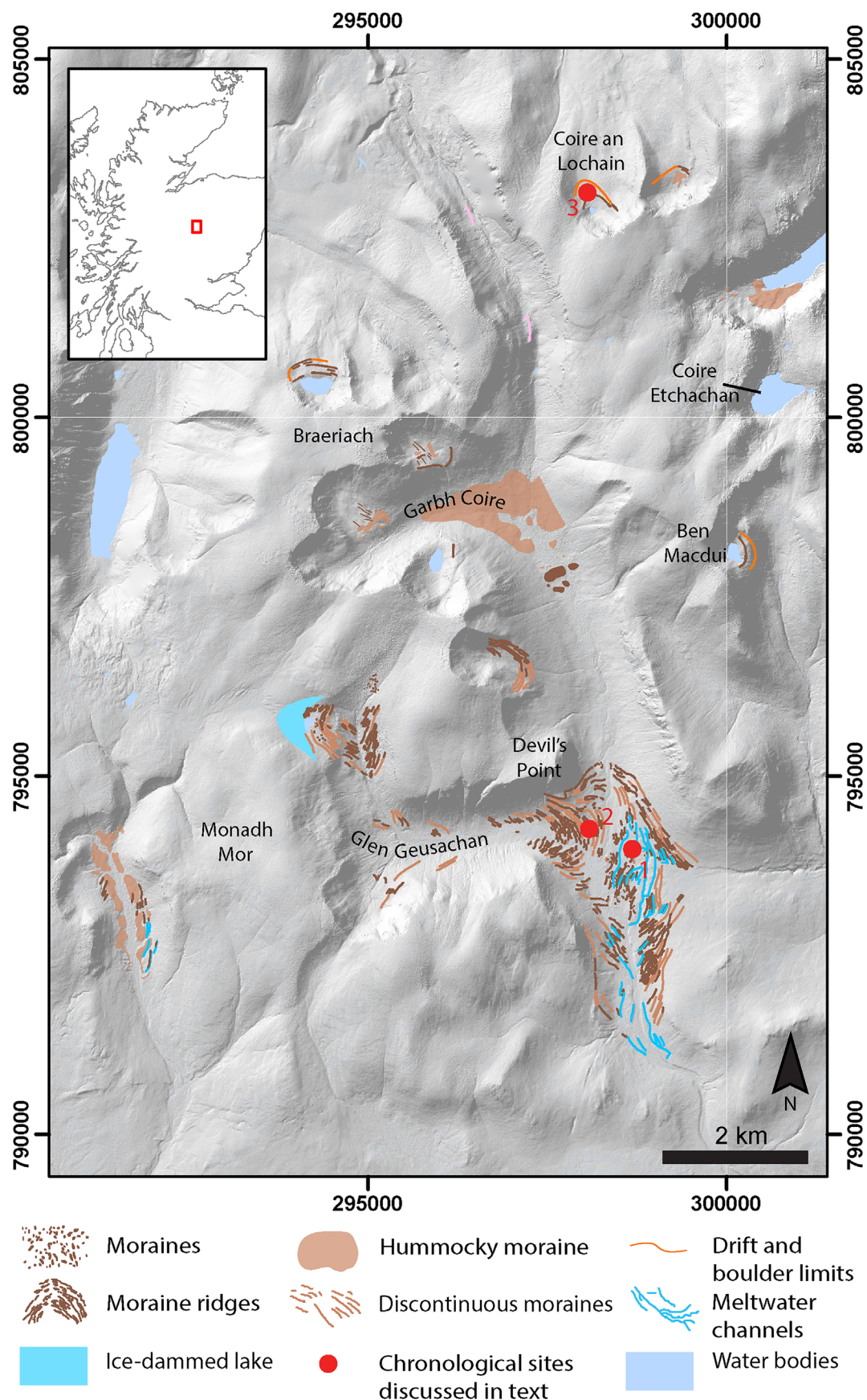


Figure 20. Geomorphological mapping of the western Cairngorms (after Sissons, 1979c; Bennett and Glasser, 1991). LLS glaciers were far less extensive in this region owing to the steep west to east reduction in precipitation. Whilst Standell (2014) suggested that plateau glaciers on Monadh Mor, Braeriach 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.

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 period of restricted glaciation. End or lateral moraine ridges mark at least part of the margins of five glaciers. The largest of these features, the Tauchers moraine complex, comprises a large outer ridge up to 200 m wide and 12 m high, inside of which sits a much narrower 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 termination of hummocky moraine (Cornish, 1981). The absence of periglacial debris lobes within the limits of these glaciers but their appearance immediately outside the proposed limits support 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. Cosmogenic isotope dating of boulders on the Tauchers moraine ridge gave ages between 13.1 ± 1.3 (1.2) and 11.5 ± 0.9 (0.7) ka (excluding one anomalous age) (Ballantyne *et al.*, 2013). These ages strongly support a LLS age for all the cirque glaciers and potentially indicate that deglaciation occurred before the end of the stadial (Ballantyne *et al.*, 2013). Several of the glaciers with particularly low ELAs, including the Tauchers glacier, had large potential snow-contributing areas to the south-west (Cornish, 1981). While generally glaciers facing between north and east were larger than those facing other aspects, the snow-blow seems to have offset this effect, as observed for the south-east-facing glacier on the east slope of Meikle Millyea (Cornish, 1981). Similarly, the largest glacier, the Loch Dungeon glacier, which accounted for almost one-third of the total ice in the area, was nourished by snowblow from the large upland area of Meikle Millyea and Milldown to the south-west.

Tweedsmuir

Early research on LLS glaciation in the Tweedsmuir Hills 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, p. 327). 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 these features 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 (1967b) reconstructed 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.

Pearce *et al.* (2014) systematically mapped the area and identified two types of moraines on the valley floors and lower slopes. Type 1 moraines occupy the upper valleys and comprise closely spaced chains of sharp-crested ridges that trend obliquely down-valley, as in Talla Valley and at the foot of Loch Skene (Fig. 21). Most of the valleys in this region lack backwalls and in some instances these moraines extend up onto the plateau. Such features are found in Fruid Valley, where they are associated with lateral meltwater channels,

strongly indicating a plateau icefield style of glaciation (Rea and Evans, 2003; Pearce *et al.*, 2014). Type 2 moraines are found sporadically at lower elevations and have more subdued morphology and no apparent pattern. Reconstruction of the extent of the icefield suggested that Type 1 moraines mark the extent of LLS glaciation, with Type 2 moraines pre-dating the stadial (Pearce *et al.*, 2013). The consistency of reconstructions based on the empirical evidence and those informed by numerical modelling of the ice surface profiles (Pearce *et al.*, 2013) inspires confidence in these results.

Geomorphological evidence in northern England

The area of England that was glaciated during the LLS is relatively small and almost entirely lies in the Lake District region in north-west England (Figs 1 and 22), 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. 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).

Lake District

The signature of LLS glaciation in the Lake District has been recognized for almost 150 years. Ward (1873) observed that after ice sheet retreat most of the 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, 1961). The first detailed geomorphological mapping of the area was conducted by Sissons (1980b) who reconstructed a series of 64 alpine style valley and cirque glaciers in the area during the LLS (Fig. 22). 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 pre-date the LLS. A LLS age for Sissons' (1980b) reconstructed glaciers was supported by the work of Pennington (1978, 1996), who identified 19 Lateglacial pollen sites beyond the proposed LLS limits, and 14 sites showing only Holocene sediments inside the limits.

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 (Fig. 22). These icefields were drained by outlet glaciers in the valleys which, in some cases, had similar extents to Sissons' (1980b) reconstructions, but in other areas were more extensive. This ice configuration explains the variations in glacier extent and the ELAs of Sissons' (1980b) 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

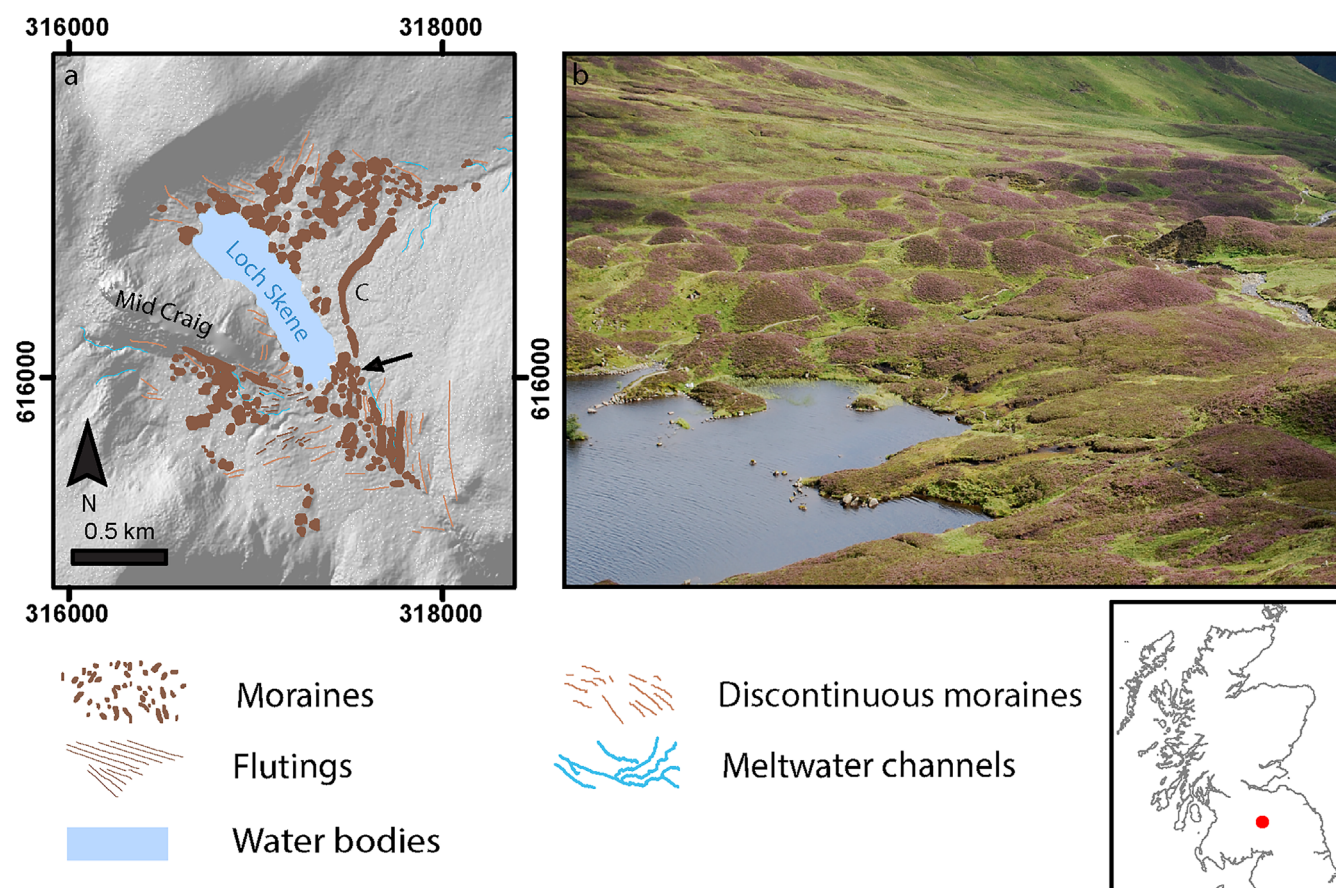


Figure 21. (a) Geomorphological mapping of the Loch Skene area in the Tweedsmuir Hills (after Pearce *et al.*, 2014). C: The Causey moraine ridge. Arrow indicates those moraines photographed in Fig. 21b. Underlying hill-shaded images were derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre. (b) Recessional moraines at the foot of Loch Skene, taken from Mid Craig looking south-east. Photograph from H.L.B.

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.

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 that range between 1 and 10 m in height and often enclose patches of hummocky moraine (Sissons, 1980b). Particularly impressive examples are found in Mosedale (Fig. 22) where two clear nested arcuate end moraines contrast starkly with the larger more subdued moraine mounds in the lower valley, which are thought to pre-date the LLS (Brown *et al.*, 2011, 2013). Coring of a peat-filled depression (Chronological Site 1, Fig. 22) in the lower valley revealed 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). Cosmogenic dating of a boulder on a ridge identified as an end moraine (Sissons, 1980b; Ballantyne and Harris, 1994) of the Keskdale glacier (Chronological Site 2, Fig. 22) gave an age of 14.0 ± 0.9 (0.7) ka, while the bedrock lip higher up in the cirque was not exposed until 13.2 ± 0.8 (0.5) ka (Hughes *et al.*, 2012). In the original publication, both dates fell within the LLS, suggesting that there were two phases of glacier retreat, but recalibration for this paper indicates that the outer ridge potentially pre-dates the LLS.

Glaciation was more widespread in the central Lake District and much of the region's most impressive glacial geomorphology is found in these valleys (Fig. 23), which Sissons (1980b) believed were occupied by alpine style valley glaciers during the LLS. In light of recent advancements of glaciological theory and remapping of the geomorphology (McDougall, 1998, 2001; Rea *et al.*, 1998; 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 (Fig. 22). Clear sequences of recessional moraines are found in many of the valleys of the central Lake District, particularly those which radiate out from the summits of High Raise, Ullscarf and Thunacar Knott (McDougall, 1998, 2001; Rea *et al.*, 1998; 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 (Fig. 6d) indicate a complex deglaciation pattern of retreat up onto the plateau rather than to the head of the valley, while meltwater channels on the slopes show that ice from Langstrath probably inundated Langdale Combe (McDougall, 2001) (Fig. 23). 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 (Fig. 23) a chaotic complex of moraines suggests localized ice stagnation, but some features indicate retreat out of the valley and onto the plateau. This evidence

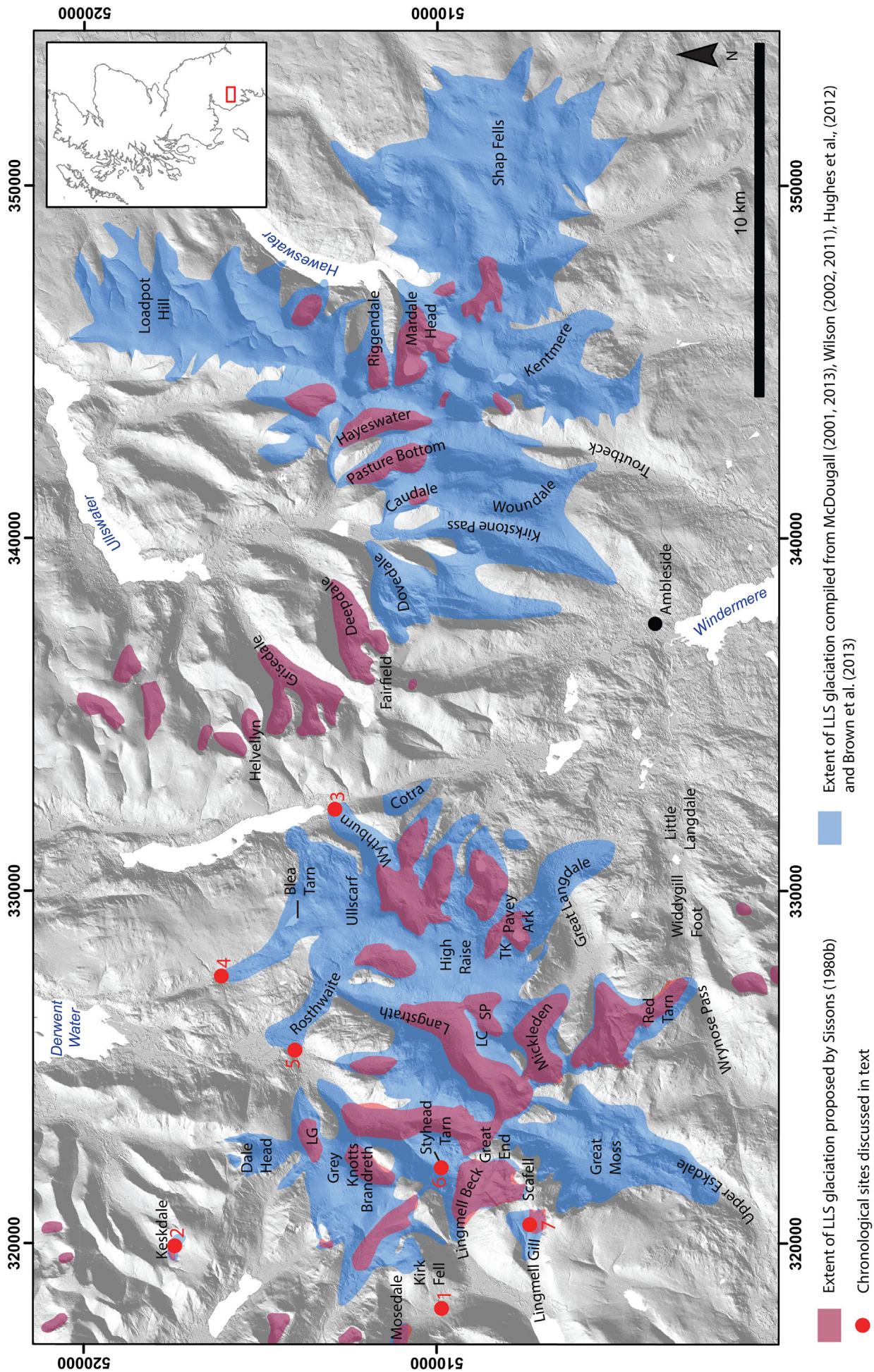


Figure 22. The extent of LLS glaciation in the Lake District. Sissons' (1980b) proposal of alpine style glaciers having developed in these hills was challenged (Rea *et al.*, 1998; McDougall, 2001, 2013) in favour of a plateau style of glaciation. Plateau icefields developed on the central and eastern fells, surrounded by cirque glaciers at topographically favourable sites. LG: Little Galesgarthdale, LC: Langdale Combe, SP: Stake Pass, TK: Thunacar Knott. Underlying hill-shaded images were derived from NEXTMap DSM from Internap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

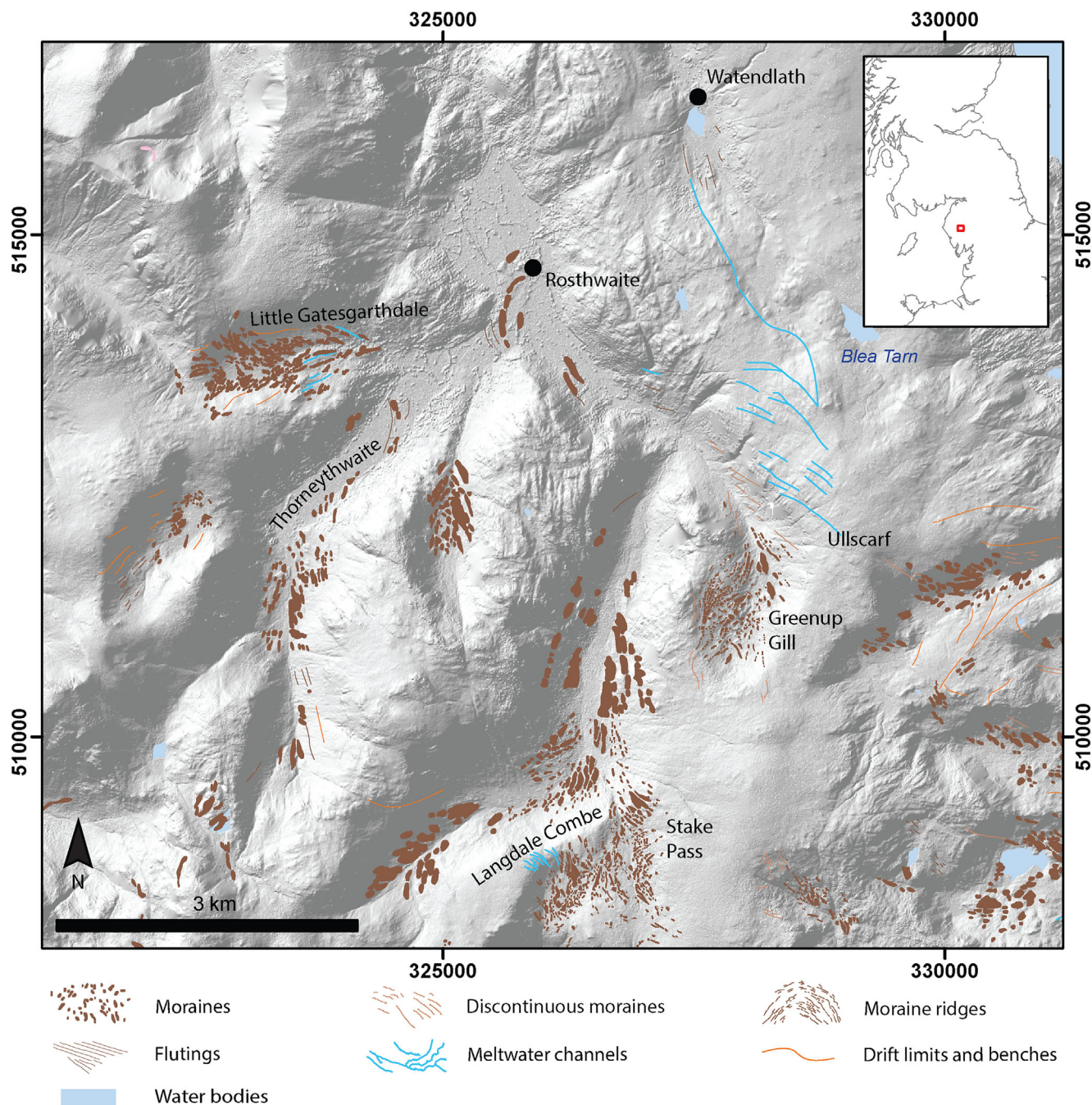


Figure 23. Geomorphological mapping of the central Lake District (Sissons, 1980b; McDougall, 1998, 2001). The sequence of recessional moraines, which are preserved in many of the valleys within this area, shows the retreat of ice up onto the plateau at Langdale Combe and Greenup Gill, indicating that 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.

points towards a much more extensive plateau icefield that contrasts with Sissons (1980b) 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,

strongly supporting a LLS age for the moraines (Walker, 1965).

The age of the Rosthwaite, Watendlath and Wythburn lobes, which mark the northern extent of the High Raise icefield, remains uncertain (Clark and Wilson, 1994). Pennington (1978) argued that the presence of Lateglacial sediments at Blea Tarn (Fig. 23) 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 break-up 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 most dates from boulders on the Wythburn, Watendlath and Rosthwaite moraines (Chronological Sites 3, 4 and 5, Fig. 22) indicated

pre-LGM ages (Wilson *et al.*, 2013). These ages seem implausible given that the last British Ice Sheet completely overran this area and they 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. An alternative approach has been used to establish the age of the moraine ridges at Cotra (Fig. 22). Carr and Coleman (2007a) argued that modelling of the reconstructed glacier showed an implausibly high component of basal motion and an implausibly low ELA and thus favoured a pre-LLS age for the moraines, but this interpretation has been challenged (Wilson, 2008) and the age remains uncertain.

In the south-eastern sector of the icefield, both Sissons (1980b) 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 (Fig. 22). Numerical modelling supports the notion that these features mark the limit of LLS glaciation (Brown *et al.*, 2013). A further glacier at Widdigill Foot (Fig. 22) has previously been afforded a LLS age (Manley, 1961; Pennington, 1978; Wilson, 2002), but remains undated. 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 Widdigill moraines pre-date the LLS.

Further west, Brown *et al.* (2011, 2013) confirmed the presence of ice in Lingmell Beck and Lingmell Gill (Fig. 22) as proposed by Sissons (1980b), but suggested that the ice had been more extensive in the valleys and that the Lingmell Beck glacier was coalescent with the main icefield. Although this conflicts with the presence of undisturbed Lateglacial deposits near Styhead Tarn (Pennington, 1978) (Chronological Site 6, Fig. 22), Brown *et al.* (2013) argued that these could have been preserved beneath cold-based ice. The Lingmell Gill glacier has been dated to the LLS by a minimum ^{36}Cl exposure age of 12.5 ± 0.8 ka (Ballantyne *et al.*, 2009) (Chronological Site 7, Fig. 22). Like Mosedale, Lingmell Beck and Lingmell Gill, moraines are present along the length of Upper Eskdale. Sissons (1980b) thought that these features pre-dated the LLS but Brown *et al.* (2013) have subsequently argued that they were formed by a large LLS glacier that initially retreated actively before stagnating below Great Moss, where it became decoupled from its accumulation centre on Scafell (Fig. 22) (Wilson, 2004). Such a retreat pattern has been replicated by numerical modelling (Brown *et al.*, 2013), suggesting this interpretation of the geomorphology is plausible. A complete absence of ice from Upper Eskdale during the LLS seems unlikely given the favourable aspect and elevation of the cirques on eastern Scafell, the presence of LLS ice on the north-western side of the ridge, and the orientation of the moraines, which could only relate to either the LLS or a very late stage of ice sheet deglaciation (Wilson, 2004).

As with the central Lake District, understanding of the extent of LLS glaciation in the eastern Lake District (Fig. 22) has shifted radically from an alpine to a plateau style of glaciation. Manley (1961) proposed that 10 glaciers developed in the area east of the Kirkstone Pass, including four valley glaciers in Hayeswater, Riggendale, Mardale Head and Kentmere. Subsequently, Sissons (1980b) reconstructed 11 cirque and valley glaciers in the area (Fig. 22), inferring a LLS age only for glaciers marked by clear, sharp-crested moraines and

consequently substituting the large Kentmere glacier for two small niche glaciers. McDougall (2013) identified that moraines and drift limits can be traced from lower in the valleys up onto the high ground, including in Hayeswater (Fig. 24), Kent Valley, Troutbeck and Caudale Beck, indicating that these glaciers were fed by ice on the plateau. Even where direct evidence of plateau ice is lacking, such a configuration explains the apparent presence of ice in unfavourable south-facing valleys within the Shap Fells. The reliability of using freshness of appearance to determine moraine age has recently been questioned (Wilson, 2002) and McDougall's (2013) reconstructed LLS plateau icefield encompasses both valleys with sharp-crested recessional features and those hosting much more subdued ridges. However, given the absence of absolute dates, the subdued nature of some features and the absence of clear terminal moraines, McDougall's (2013) reconstruction was speculative.

Between the reconstructed plateau icefields of the central and eastern Lake District, much of the Helvellyn range has not been mapped since Sissons (1980b) proposed that valley glaciers had occupied Grisedale and Deepdale and that a further seven cirque glaciers had existed in the northern hills of the range. A further valley glacier in Dovedale (Fig. 22) was reconstructed by Wilson (2011). McDougall *et al.* (2015) suggested that the Deepdale glacier may have been fed by an icefield on Fairfield, but presented no reconstruction. A range of retreat styles appear to be represented by moraines in this area; the clear recessional moraines in Deepdale indicate active retreat throughout deglaciation, whereas the single, large end moraine at Wolf Crag (just north of the area in Fig. 22) suggests that the cirque glacier remained at its maximum for a sustained period, while being nourished by snow blown onto the glacier from the plateau area to the south-west (Manley, 1961; Sissons, 1980b). A similar large end moraine is present at Bowscale Tarn (Evans, 1994), one of seven small cirque glaciers reconstructed in the Skiddaw Range (approximately 10 km north of Derwent Water). All seven glaciers (three proposed by Sissons, 1980b, three by Wilson and Clark, 1999 and one by Clark and Wilson, 2001) would have had favourable aspects, facing between north-west and east and would have received snow from plateau areas to the south-west.

The Lake District is extremely significant in the study of LLS glaciation in Britain as the first region where the notion of a primarily alpine style of glaciation was challenged in favour of a plateau icefield configuration (McDougall, 1998, 2001, 2013; Rea *et al.*, 1998), and a paradigm shift has since occurred for other LLS icefields in Britain (Boston, 2012a,b; Boston *et al.*, 2013; Pearce *et al.*, 2013). Sandwiched between the two plateau icefields, the Helvellyn Range remains largely unmapped since Sissons' (1980b) 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.

Pennines and Cheviots

LLS glaciation in the English Pennines (Fig. 1) has been the focus of comparatively little research. The first account of the

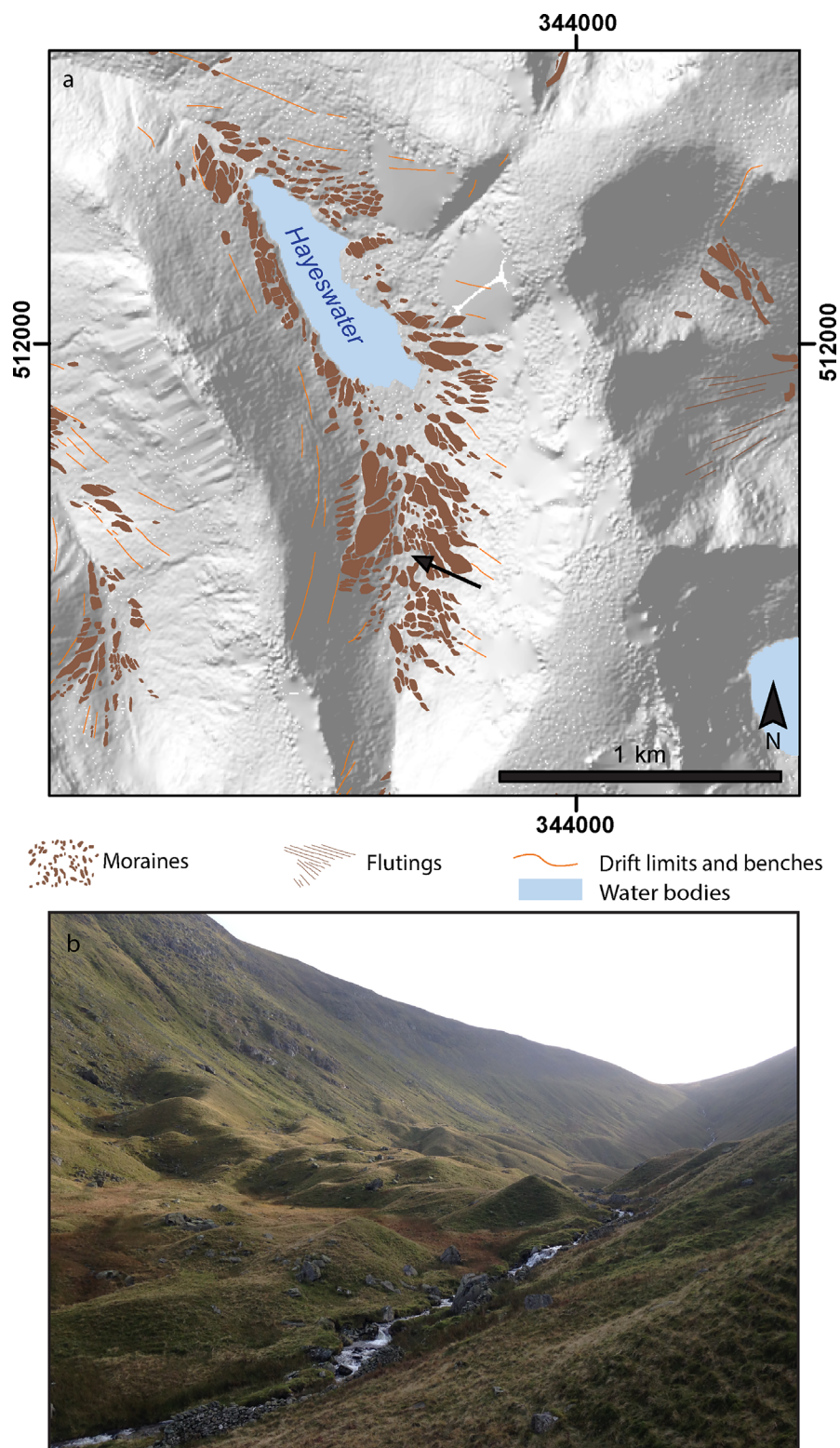


Figure 24. (a) Sharp-crested recessional moraines at Hayeswater (after McDougall, 2013). At the valley head the moraines show glacier retreat back onto the plateau, refuting the alpine style of glaciation Sissons (1980b) proposed for this site. Image b was taken looking south from the point indicated 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.B.

LLS in the area was by Rowell and Turner (1952), but their interpretation of ridges as lateral and terminal moraines was challenged by Manley (1961), who reclassified them as snow-bed features. Six small LLS cirque glaciers have been proposed in this area, five identified by Mitchell (1996), at Great Coum, Swath Fell, Combe Scar, Whernside and Cautley Crag, and a sixth at Cronkley Scar (Wilson and Clark, 1995). A unifying feature of all six glaciers is their location downwind of large areas of potential snow-blow 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 sediments from inside the glacier limit at Combe Scar support a LLS age for these glaciers (Gunson, 1966).

In the Cheviots (Fig. 1), the only site proposed to have nourished a LLS glacier 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). The Bizzle Cirque is situated downwind of a

large plateau area and it seems highly likely, given the absence of LLS ice elsewhere in the Cheviots, that snowblow from this plateau was essential for the nourishment of the 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).

Geomorphological evidence in Wales

During the LLS, glaciation in Wales was restricted to upland sites, particularly Snowdonia and the Brecon Beacons, where cirque glaciers developed at favourable sites (Figs 25 and 26) (Carr, 2001; Hughes, 2002, 2009; Coleman and Carr, 2008; Bendle and Glasser, 2012). 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 snow-beds rather than glaciers (Unwin, 1970; Shakesby and Matthews, 1993; Carr *et al.*, 2007a).

Snowdonia and northern Wales

Evidence of LLS glaciation in Snowdonia (Fig. 25) 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). Seddon (1957) mapped the 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, 11 of which he attributed 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). Nineteen were sub-classified as protalus ramparts. Unwin's (1970, 1975) Younger Series glaciers were predominantly north-easterly facing and showed 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 geomorphological mapping was first conducted by Gray (1982), who mapped evidence for 35 former glaciers and 16 semi-permanent snow-beds marked by 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 (Fig. 25).

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 Lateglacial tripartite sequences of sediment in two lake basins beyond the cwm/cirque moraine limits, at Llyn Dwythwch (2 km west of the area in Fig. 25) and Nant Ffrancon (Fig. 25), but only Holocene sediments, showing a single transition from cold to warm conditions, inside the limits at Cwm Idwal and Cwm Cyngorion (approximately 1 km west of Afon Arddu, Fig. 25). Similarly, Lateglacial sediments were found in lake basins from outside the limits at Capel Curig (Crabtree, 1972) while only Holocene sediments were found within the moraine limits at Llyn Clyd, Llyn Glas and Melynlyn (Fig. 25)

(Evans and Walker, 1977; Walker, 1978). At Cwm Llydaw and Cwm Cywion (Chronological Sites 1 and 2, respectively, Fig. 25) the onset of organic sedimentation was radiocarbon dated to about 11 000 cal a BP (Ince, 1981, 1983), indicating a LLS age for these glaciers. Taken together, this body of evidence makes a LLS age for the cirque glaciers highly likely. This assertion is further supported by cosmogenic isotope dating of the moraine damming Llyn Idwal (Chronological Sites 3 and 26, Fig. 25), which gave ^{36}Cl 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).

Recessional moraines are present in many of the Snowdon cirques but in many cases are concentrated in a belt near the former glacier terminus, with the accumulation areas often being characterized by ice-moulded bedrock (as at Cwm Dulyn, Cwmglas Bach and Cwmglas Mawr; Fig. 25). 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 Melynlyn (Gray, 1982) (Fig. 25). Protalus ramparts are present at sites that were unfavourable for glacier formation, but that nourished snowbeds during the LLS (Bendle and Glasser, 2012). The Snowdonian LLS glaciers are relatively small, with only four of the 38 reconstructed by Bendle and Glasser (2012) measuring over 1 km². The reconstructed glaciers in the Carneddau and Snowdon massifs are variable in both size and aspect. This contrasts with the consistent aspect and size of the Glyderau glaciers, the outermost moraines of which rarely exceed 1 km in distance from the cirque backwall (Bendle and Glasser, 2012) (Fig. 25). The size and shape of these cirques has been attributed to resistant igneous side and backwalls and weak mudstone floors acting as geological controls while the north-easterly aspects of the glaciers would provide protection from insolation and nourishment by windblown snow from the plateau to the south-west (Addison, 1997).

Of the Glyderau cirques, Cwm Idwal is perhaps the most famous and intensively researched site in Snowdonia (Darwin, 1842; Seddon, 1962; Phillips *et al.*, 1994; Addison, 1997; Bendle and Glasser, 2012). An impressive series of recessional moraines cover the valley floor for approximately 1 km, the outermost damming Llyn Idwal (Bendle and Glasser, 2012), and have been dated to the LLS (Phillips *et al.*, 1994). The moraines in the inner cirque are much more pronounced in appearance than the outer moraines, with particularly large ridges, up to 15 m high, being present on the western side of the lake. The most substantial feature (Fig. 25, feature X) comprises a 450 m-long, sharp-crested ridge (Graham and Midgley, 2000), which has been interpreted as a protalus rampart with moraine below (Unwin, 1975), an ice-marginal lateral moraine (Seddon, 1957) and subglacial flutes (Gray, 1982). It is also possible that the Idwal glacier was fed by an icefall from Cwm Cneifion and was deflected north-west, suggesting that the ridge represents frontal moraines (Addison, 1997). Graham and Midgley (2000) favoured formation of the ridges by englacial thrusting within a composite Idwal–Cneifion glacier, although they acknowledged the difficulty in substantiating this idea.

The LLS glaciers within the Snowdon massif 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, the largest LLS glacier reconstructed in the area, which accounted for 4.48 km² of the cumulative area (20.74 km²) of the 38 Snowdonia glaciers (Bendle and Glasser, 2012). This glacier

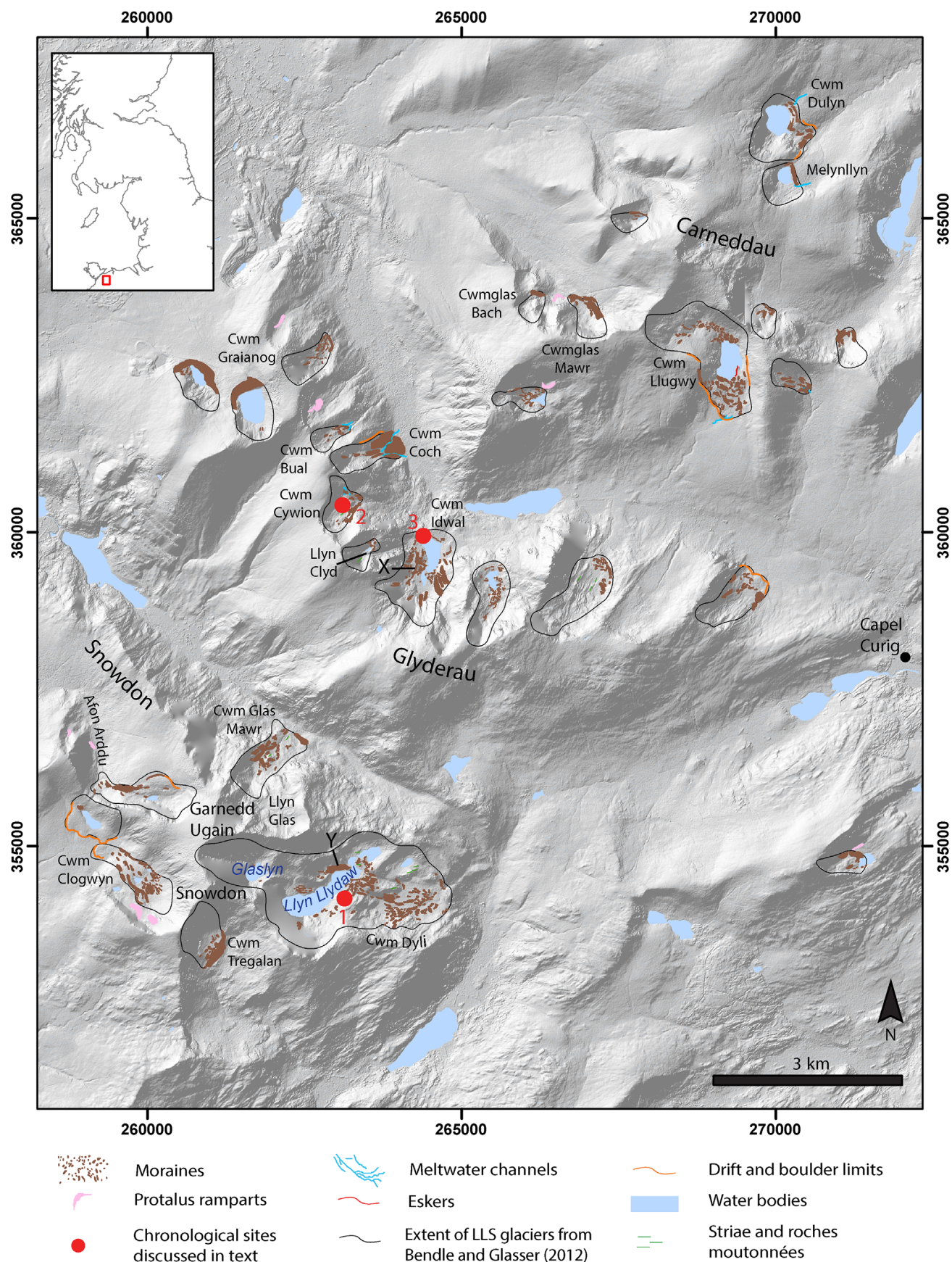


Figure 25. The LLS glacial geomorphology of Snowdonia and reconstructed ice extents (after Bendle and Glasser, 2012). Underlying hill-shaded image derived from NEXTMap DSM from Intermap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

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 (Gray, 1982; Addison, 1997; Bendle and Glasser, 2012)

(Fig. 25). Glaslyn is dominated by ice-moulded bedrock and glacially transported boulders with moraines only becoming abundant at the eastern end of Llyn Llydaw. The large steep-sided ridge on the lake shore (Figs 6e and 25, feature Y) is

thought to represent re-stabilization of the Llydaw glacier after it separated from ice in Cwm Dyli (Gray, 1982). 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 and carrying little englacial debris, and that its flow was topographically constrained (Sharp *et al.*, 1989). Further evidence of LLS glaciation, predominately moraines, is found in the larger central cirques around Snowdon (Cwm Du'r Arddu, Cwm Clogwyn, Cwm Tregalan and Cwm Glas Mawr) and in the smaller north-east-facing cirques that flank the Nant Ffrancon glacial trough (Cwm Graianog, Cwm Bual and Cwm Coch) (Fig. 25). The reconstructions of these glaciers by Gray (1982) and Bendle and Glasser (2012) are very similar, although the latter proposed a slightly greater extent of ice in Cwm Du'r Arddu.

Research into LLS glaciers in the mountain areas peripheral to Snowdonia is somewhat limited. Hughes (2002) reconstructed three small cirque glaciers in the Arenig mountains at Llyn Arenig Fach, Llyn Arenig Fawr and Cwm Gylchedd [although the last is not included in Bickerdike *et al.* (2016) as mapping could not be georeferenced with sufficient accuracy]. A LLS age for these glaciers is consistent with the absence of mature periglacial features inside the reconstructed limits and with the absence of pre-LLS sediments inside the Cwm Gylchedd limits (Hughes, 2002) and Llyn Arenig Fach (Lowe, 1993). A further six glaciers have been reconstructed in the Aran and Berwyn Mountains (Hughes, 2002, 2009). All nine of these reconstructed cirque glaciers 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 (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). The largest of these glaciers occupied Cwm Cau, although the maximum extent of this glacier has been disputed by Lowe (1993) and Sahlin and Glasser (2008), the latter favouring a more extensive limit. Ballantyne (2001) reported that the 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. Larsen (1999) suggested that this area was probably not exposed to severe periglacial conditions during the LLS and thus was glaciated. Lowe (1993) suggested that two cirque glaciers had developed along the northern escarpment face. Sahlin and Glasser (2008) proposed that moraines were found in four sites, suggesting that a boulder-strewn lobe just west of Cwm Gadair was a series of recessional moraines from a debris-strewn extension of the Gadair glacier, rather than an ice-cored rock glacier, as proposed by Lowe (1993). The presence of only Holocene sediments in basins within the limits of the Cwm Cau and Cwm Gadair glaciers supports a LLS age for these glaciers (Lowe, 1993).

Brecon Beacons

The Brecon Beacons represent the southernmost location in Britain where LLS glaciers formed (Shakesby and Matthews, 1993) (Figs 1 and 26). Environmental conditions in this area were only very marginally suitable for glacier development (Ellis-Gryffydd, 1977) and, consequently, landform evidence consists mainly of single depositional ridges at the foot of the predominantly north-east-facing sandstone escarpments (Fig. 26). Uncertainties about the mode of formation of the depositional ridges have been the focus of research for over a century (Reade, 1894; Lewis, 1970; Shakesby and Matthews,

1996; Carr, 2001) and are compounded by a lack of absolute dating in the area.

Based on the distribution of glacial lineations, Jansson and Glasser (2008) proposed that the depositional ridges in the valley heads (namely Fan Hir and Cwm Llŵch) were formed by regional ice sheet flow deflected into valleys during the LGM. This suggestion has been undermined by the presence of only Holocene and younger sediment within the limits of these ridges, for example at Craig y Fro and Craig Cerrig-leisiad (Preston *et al.*, 2007; Walker, 2007a,b). Furthermore, Shakesby and Matthews (2009) argued that the morphology of the features at Fan Hir and Cwm Llŵch (Fig. 26a,c) precludes formation by LGM ice deflected up the valleys. Lewis (1970) proposed that two phases of cirque glaciation had occurred in the Brecon Beacons and that during the second, taken to represent the LLS, only three sites (Craig Cerrig-leisiad, Cwm Llŵch and possibly Craig y Fro) nourished glaciers. Re-evaluation of many sites has since shown that the depositional 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.

Along the foot of the Mynydd Du (the Black Mountain) escarpment (Fig. 26a) 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 (Fig. 26a). This linear ridge alternates between sharp-crested, up to 25 m high, and more subdued sections, before curving towards the scarp at its southern end (Shakesby and Matthews, 1993). Previously thought to be a proglacial rampart because of its linearity (Ellis-Gryffydd, 1972), subsequent studies (Shakesby and Matthews, 1993; Carr and Coleman, 2007a) have favoured a glacial origin for the feature. The abraded and striated clasts within the ridge, its curvature towards the scarp and the potential presence of two small recessional moraine ridges between the ridge and the scarp support the interpretation that the Fan Hir ridge formed at the margin of a small, locally nourished glacier (Shakesby and Matthews, 1993). If the ridge were a proglacial rampart, an implausibly high rate of escarpment erosion would have been necessary to produce sufficient debris to form the feature and the depression behind the ridge is such that sufficient snow could have accumulated to form a glacier (Shakesby and Matthews, 1993). 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).

The origins of other ridges along the north-facing section of Mynydd Du (Fig. 26a) have also been debated, for instance the easternmost feature in Cwm Sychlwch, 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 also been used to confirm whether glaciers could have developed in these sites under LLS conditions (Carr *et al.*, 2007a). Curry *et al.* (2007) differentiated between sites that had been glaciated during the LLS (such as Llyn y Fan Fach) and those which probably remained

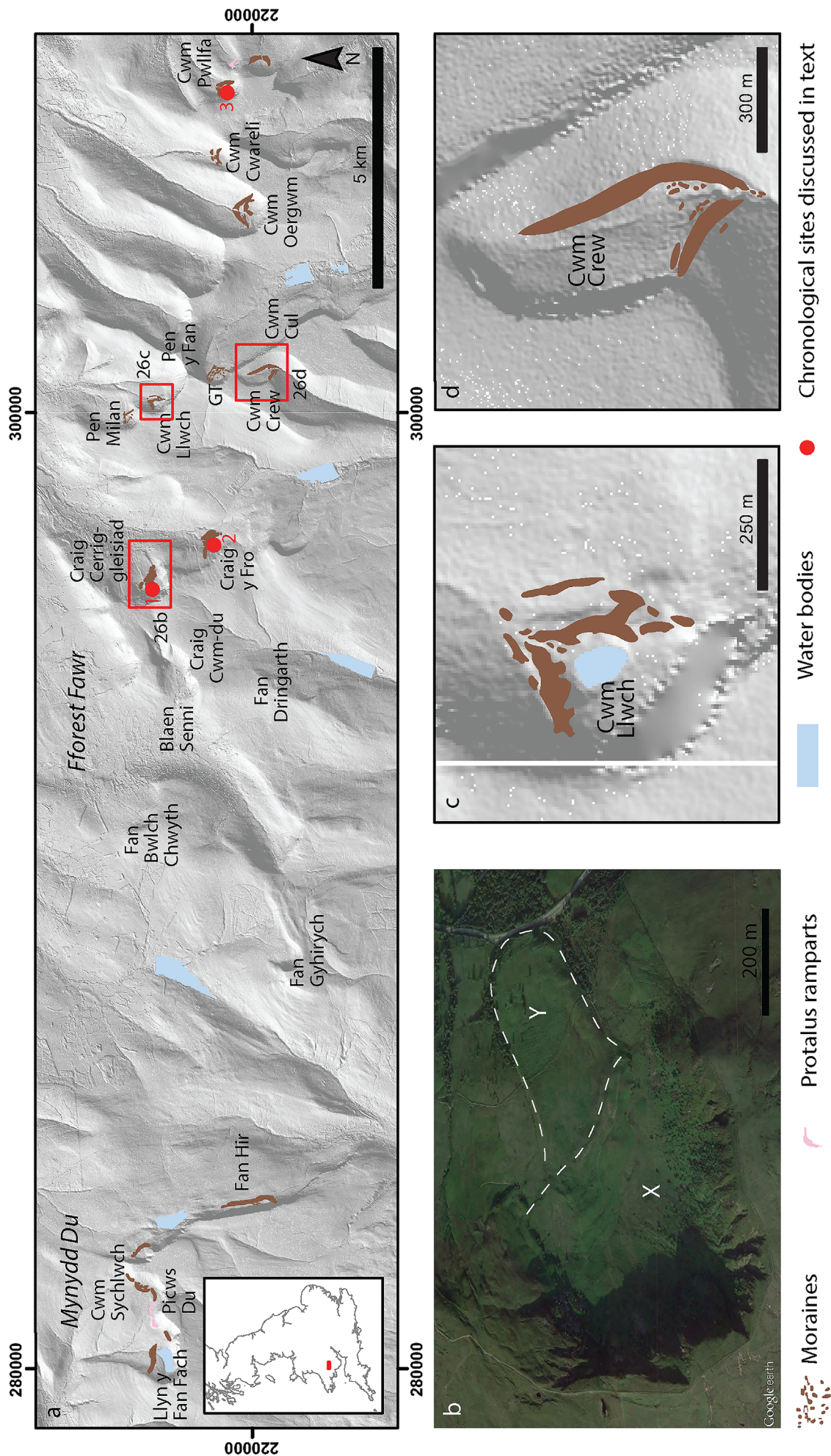


Figure 26. (a) The glacial geomorphology of the Brecon Beacons (after Shakesby and Matthews, 1993, 2007a,b; Carr, 2001; Shakesby, 2007; Carr and Coleman, 2007c; Coleman and Carr, 2008). The style of glaciation in this marginal region is one of cirque and niche glaciers and permanent snowbeds. GT: Cwm Gwaun Taf. (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 December 2013. (c) End moraine ridges at Cwm Llwch. (d) The impressive terminal moraine in Cwm Crew. Underlying hill-shaded images were derived from NEXTMap DSM from Internmap Technologies, Inc. provided by the NERC Earth Observation Data Centre.

exposed to severe periglacial conditions (such as Picws Du) based on the maturity of talus at these sites.

Similar difficulties are associated with identifying the origin of depositional ridges in the Fforest Fawr area in the central Brecon Beacons. Although small LLS glaciers are thought to have existed at Fan Gyhirych, Blaen Senni, Craig Cwm-du and Fan Bwlch Chwyth (Shakesby, 2002), there is a lack of detailed and systematic mapping of these features. The most extensively studied site within the Fforest Fawr area is Craig Cerrig-gleisiad (Fig. 26b), which gives an insight into the relationship between glacial and paraglacial activity in the Brecon Beacons. At this site, a complex of ridges, between 10 and 20 m high and aligned north-west to south-east, sits closest to the scarp (Fig. 26b, X), from which a long tongue of deposits extends to 1 km north-east of the escarpment head-wall (Y) (Shakesby and Matthews, 1996, 2007a). The contrasting morphology, sedimentology and extent of the two features led Lewis (1970) to believe they represented two phases of cirque glaciation. Shakesby and Matthews (1996) refuted a wholly glacial origin for the features, arguing that it would be unusual for a glacier with such a small accumulation area to have responded to variable snowblow from the plateau with such different extents. They also refuted Ellis-Gryffydd's (1972) suggestion that the tongue was formed during ice sheet wastage, arguing that its morphology indicates movement away from the scarp. Instead Shakesby (2002) and Shakesby and Matthews (1996) proposed that the tongue of debris represents a landslide triggered by debuttressing of the scarp during ice sheet deglaciation, and that the inner ridges represent modification of the deposits during a subsequent phase of cirque glaciation. This explanation accounts for both the restriction of striated clasts to the inner ridges, and the similarity of the debris tongue to the landslide at nearby Fan Dringarth (Shakesby and Matthews, 1996, 2007a; Shakesby, 2002) (Fig. 26a). Coring of the peat bog enclosed by the moraines indicated that the transition from minerogenic sediments occurred between 12 910 and 12 665 cal a BP (Walker, 2007a) (Chronological Site 1, Fig. 26a). Matching of 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, while the pollen sequence supports a LLS age for this limit (Walker, 2007a). The contribution of windblown snow from the plateau to the south-west was almost certainly an important factor in nourishing this glacier and may have accounted for up to 45% of accumulation (Carr and Coleman, 2007b).

A glacial origin has been suggested for the ridges at Craig y Fro, based on the presence of abraded and striated clasts, the distance from the headwall (which refutes a protalus rampart origin) and the absence of evidence of headwall instability (which precludes a landslide origin) (Shakesby and Matthews, 2007b). 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 from the adjacent plateau (Carr *et al.*, 2007b). Furthermore, Carr *et al.* (2007b) argued that the reconstructed glacier was inconsistent with the parameters used to verify other LLS glaciers in the region, favouring a rockslope failure rather than glacial origin for the features. 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. Sediment cores taken proximal to the ridges revealed the onset of organic sedimentation between 11 964 and 11 245 cal a BP (Walker, 2007b) (Chronological Site 2, Fig. 26a) but the origin of the features remains uncertain.

In the eastern Brecon Beacons, depositional ridges have been recognized at about a dozen sites at the heads of the glacial valleys and indentations into the escarpment (Fig. 26). At the head of Cwm Llwh (Fig. 26c) 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 Llwh glacier could have formed during the LLS, but would have been slow moving, leading Carr (2001) to infer that a large component of debris fell directly onto the glacier from the valley walls. This style of a small glacier accumulating in (usually) a north-east-facing cirque and forming an arcuate end moraine, which in some instances encloses inner fragmented ridges, is typical for the Brecon Beacons. Using glaciological modelling, Carr (2001) and Coleman and Carr (2008) asserted that glaciers reconstructed from these ridges at Cwm Llwh, Pen Milan, Cwm Gwaun Taf, Cwm Crew, Cwm Oergwn and Cwm Cwareli 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. The Cwm Crew glacier (Fig. 26d), which is marked by a sharp-crested ridge up to 15 m in height (Shakesby, 2002), is exceptional in its unfavourable southerly aspect which would have afforded no protection from insolation and, consequently, the contribution of windblown snow from the high ground to the south-west must have been substantial to offset ice lost by melting.

Although moraine ridges have been found at other sites within the area, these are thought to pre-date the LLS. For example, reconstruction of a glacier from moraine ridges at Cwm Cul produced an implausibly thin glacier that required almost all movement to be driven by basal motion, leading Carr (2001) to reject a LLS age for the glacier. The same was true of four glaciers along the northern flank of Mynydd Llangattwg (approximately 14 km south-east of Cwm Pwllfa), which were also deemed to pre-date the LLS (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 (Chronological Site 3, Fig. 26a) where a radiocarbon date of 9425–9139 cal 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.

Discussion

Assessment of evidence

A key issue regarding 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 south-western Highlands, where mapping is limited to four localized sites along the shores 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. 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, has been attributed to very active, potentially streaming, glaciers stripping deposits from the valley sides followed by a rapid retreat that left little depositional evidence (Bennett, 1991). The same processes might have been responsible for the apparent lack of

evidence in the south-west Highlands along the banks of Lochs Fyne and Long. 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 Houran and Loch Nevis, are present.

For the eastern outlet glaciers, the evidence 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 appear to be widespread within this area, although future mapping of these features could be inhibited in some areas by 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 suggests that in some areas LLS glaciers extended beyond the extent inferred from the onshore evidence (Dix and Duck, 2000; McIntyre *et al.*, 2011). 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. 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 and may reflect the imagery used to produce it (Smith *et al.*, 2006). While some researchers have mapped individual moraine mounds and hummocks as defined by the break of slope around the landform (Benn *et al.*, 1992; Ballantyne, 2002; Lukas and Lukas, 2006a,b), others have mapped moraine ridge crests or, simply areas of 'hummocky moraine'. For 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 the dynamics of past ice masses are not misinterpreted.

Use 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, thought to be recessional moraines. Where detailed mapping records the position of individual moraine mounds or ridge crests, researchers have identified the positions of palaeo-ice fronts and inferred the dynamics of these actively retreating glaciers (Bennett and Boulton, 1993a,b). 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 that formed them. This is currently the case for much of the Central Highlands, including the south-east 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. 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. While 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 (Ellis-Gryffydd, 1977; Shakesby, 2007). There has also 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, 2002; 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 LLS palaeoclimatic conditions. It is hoped that highlighting the ambiguous genesis of these landforms in this review will prompt future targeted research to determine their origins.

Given that geomorphological mapping forms the first step in reconstructing both the dimensions and the dynamics of former glaciers, it is crucial that these data are both detailed and accurate. Analysis of retreat dynamics (cf. Lukas and Benn, 2006) is not possible from low-resolution data, such as where areas of moraines, rather than the individual landforms, are recorded. Thus, 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.

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-dependent 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 a 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 YD scenario, which closely accorded with 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 to 12 ka and a low component of basal sliding. The closeness of the modelled output with the empirical limits at a national scale inspires confidence and the small parameter space within which this match occurs suggests that these conditions could closely represent those during the LLS.

However, there are discrepancies between the modelled and empirical limits in several locations, particularly 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 result from the reduction of basal drag by a deformable sediment bed or water body, allowing the glacier to advance further, which the model does not account for. The anomalously extensive limits may also reflect glacier surging (Thorp, 1991; Evans and Wilson, 2006b). Incorporating bed characteristics into future numerical models could facilitate more accurate representation of glacier extents and allow inferences about the controls on ice flow dynamics to be made. Similarly, Golledge *et al.*'s (2008) model assumes LLS glaciers developed from ice-free conditions. Bradwell *et al.* (2008) argued that ice caps existed in the north-west Highlands throughout the Lateglacial Interstadial and into the LLS, although this notion has since been challenged (Ballantyne and Stone, 2012). 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 north-west Highlands and other sites. Further work to determine whether ice was inherited is needed and, if this is 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 cap (Fig. 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 south-westerly 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 YD glaciation was characterized 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, p. 1022). 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.

Style of glaciation

Advances in glaciological theory and mapping technology have influenced the prevailing understanding of LLS glaciation in Britain. Charlesworth's (1955) study was the first attempt to detail the extent and retreat patterns of LLS ice throughout Scotland and, although much of the work has subsequently been revised, represents a landmark study in this body of literature. The first systematic and detailed mapping of many locations was conducted by Sissons (1977a,b, 1979b, 1980b) from aerial photographs and fieldwork. This research gave rise to the understanding that the LLS was characterized in most areas by alpine style valley and cirque glaciers, with the exception of the Gaick plateau icefield (Sissons, 1974). Following the proposal that LLS glaciers in the central Lake District were probably nourished by a plateau icefield (McDougall, 1998; Rea *et al.*, 1998), many of the upland areas thought 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; Brown *et al.*, 2011, 2013; Finlayson *et al.*, 2011). Such reinterpretation has implications for glacial reconstructions and for ELAs calculated from them; incorrectly assuming cirque occupation only produces an artificially lower ELA than if the mass of ice on the plateau is accounted for (McDougall, 2001). As several studies have assessed regional trends in ELAs (e.g. Ballantyne, 2007b) or used them to infer palaeoclimatic conditions, such as temperature and precipitation (e.g. Ballantyne, 2002, 2007a, b; Benn and Ballantyne, 2005), accurately calculating ELAs and therefore correctly interpreting of 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 thought. 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; Bickerdike *et al.*, 2018).

Snowblow onto glacier surfaces was critically important in ice accumulation and accounts for the presence of cirque glaciers in topographically favourable locations (Bickerdike *et al.*, 2018). This applied especially to marginal areas where glaciers formed downwind of large potential snow-contributing areas, despite conditions generally being unsuitable for glaciation. These small glaciers were generally north to east facing, as in the Pennines (Mitchell, 1996). However, in some cases the

effect of snowblow was sufficient to offset the increased insolation received by southerly facing cirques, for example the cirque south-east of Meikle Millyea in the Southern Uplands (Cornish, 1981) and Cwm Crew in the Brecon Beacons (Carr, 2001).

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 (Fig. 27). This builds upon earlier studies (e.g. Gollledge, 2010; Sissons, 1967a) that concentrated on Scotland by also incorporating glaciers in England and Wales. Empirically based glacial reconstructions derived from the 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). 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. Where geomorphological evidence is absent, the resolution of this reconstruction is, by necessity of its scale, reasonably coarse 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, 2002; 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. 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 certainty with which the glacier limits presented on Fig. 27 have been identified varies across Britain. Figure 28 represents a preliminary attempt to qualitatively identify which limits are well-constrained, both by geomorphology and by chronology. Chronological sites discussed in the review are included, highlighting how few of the LLS glacier limits are robustly dated. Each site is labelled according to its calibrated age in this review: 'LLS' where chronological evidence supports a LLS age, or 'LD' where dating has indicated formation during the retreat of the last British–Irish Ice Sheet. Sites where the extent of LLS glaciation is well-constrained by both geomorphology and absolute dating are comparatively rare (Fig. 28), including the termini of the Lomond and Menteith glaciers. Elsewhere, the extent of ice is well-constrained by high-resolution geomorphological mapping but chronological control on the limits may either be indirect (e.g. sections of the Mull icefield) or has been contradicted by subsequent work (e.g. Gleann Chaorainn in the Monadhliath). In some cases, chronological control is completely absent.

The extents of the glaciers that comprised the West Highland Glacier Complex are moderately well-constrained in the northern and central sectors. However, because it has been suggested that glaciers that flowed beyond the present-day coastline may have extended further than inferred from onshore evidence (Dix and Duck, 2000; McIntyre *et al.*, 2011), the ice limits along the western coast to the north of

the Great Glen are here viewed as speculative. Likewise, the limits of glaciation on the Gaick Plateau and in the Cairngorms and south-eastern 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 geomorphological evidence 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, and in the northern Tweedsmuir Hills.

The reconstruction shapefile, available as supplementary material with this paper, provides a framework for future research, Fig. 28 highlighting regions of uncertainty where further work is required. 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. Use 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, with Fig. 28 indicating which limits models need to fit closely and which are more speculative. While modelling was used to identify a narrow parameter space of palaeoclimatic and basal conditions that matched well with the extent of glaciation in Scotland (Gollledge *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.

Timing and dating of LLS glaciation

Much of the uncertainty surrounding the extent and nature of LLS glaciation in Britain results from the currently limited number of absolute dates for glacial features associated with this period (Fig. 28). 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 assigned to glacier limits based on the presence of hummocky moraine, which was believed to be diagnostic of the stadial (Sissons, 1974). However, the presence of Lateglacial 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 (Fig. 3) (Benn, 1990) 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 have observed a mutually exclusive relationship between the LLS glacial landforms and relict periglacial features, including frost-weathered bedrock, solifluction lobes and mature talus slopes (Sissons, 1974; Ballantyne and Wain-Hobson, 1980; Ballantyne, 1989, 2007b; Benn and Ballantyne, 2005; Finlayson, 2006). Given the absence of periglacial features within the reconstructed glacier limits, but

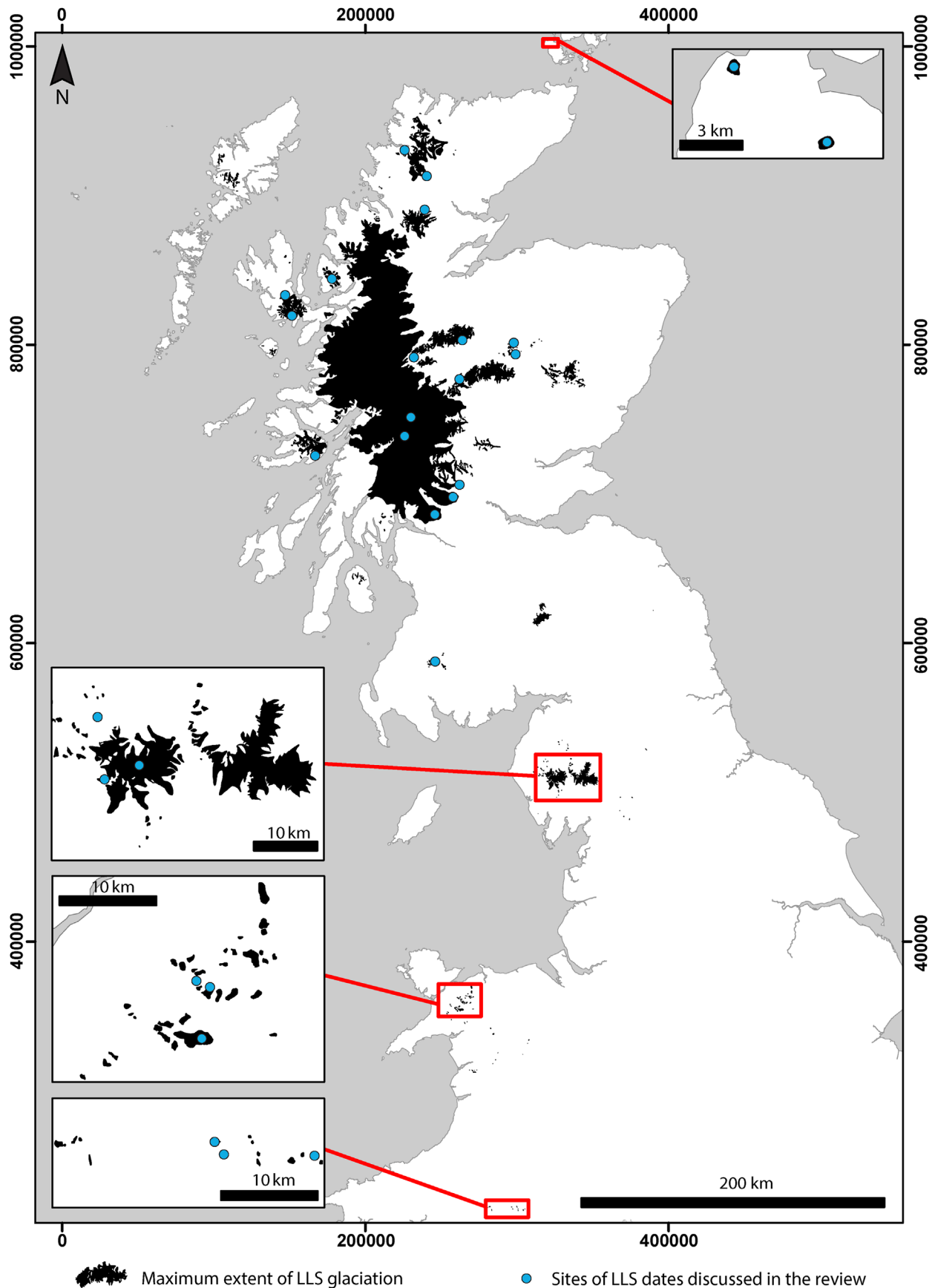


Figure 27. Approximate reconstruction of the maximum extent of LLS glacial limits in Britain. The West Highland Glacier Complex is shown flanked by satellite icefields, while in marginal areas glaciation is limited to topographically favourable valleys and cirques. Britain coastline and present-day water bodies reproduced from Ordnance Survey © Crown copyright and database Right 2015. Ordnance Survey (Digimap Licence).

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, Holocene shorelines were present within them, suggesting that the older shorelines formed during ice sheet deglaciation were destroyed during a

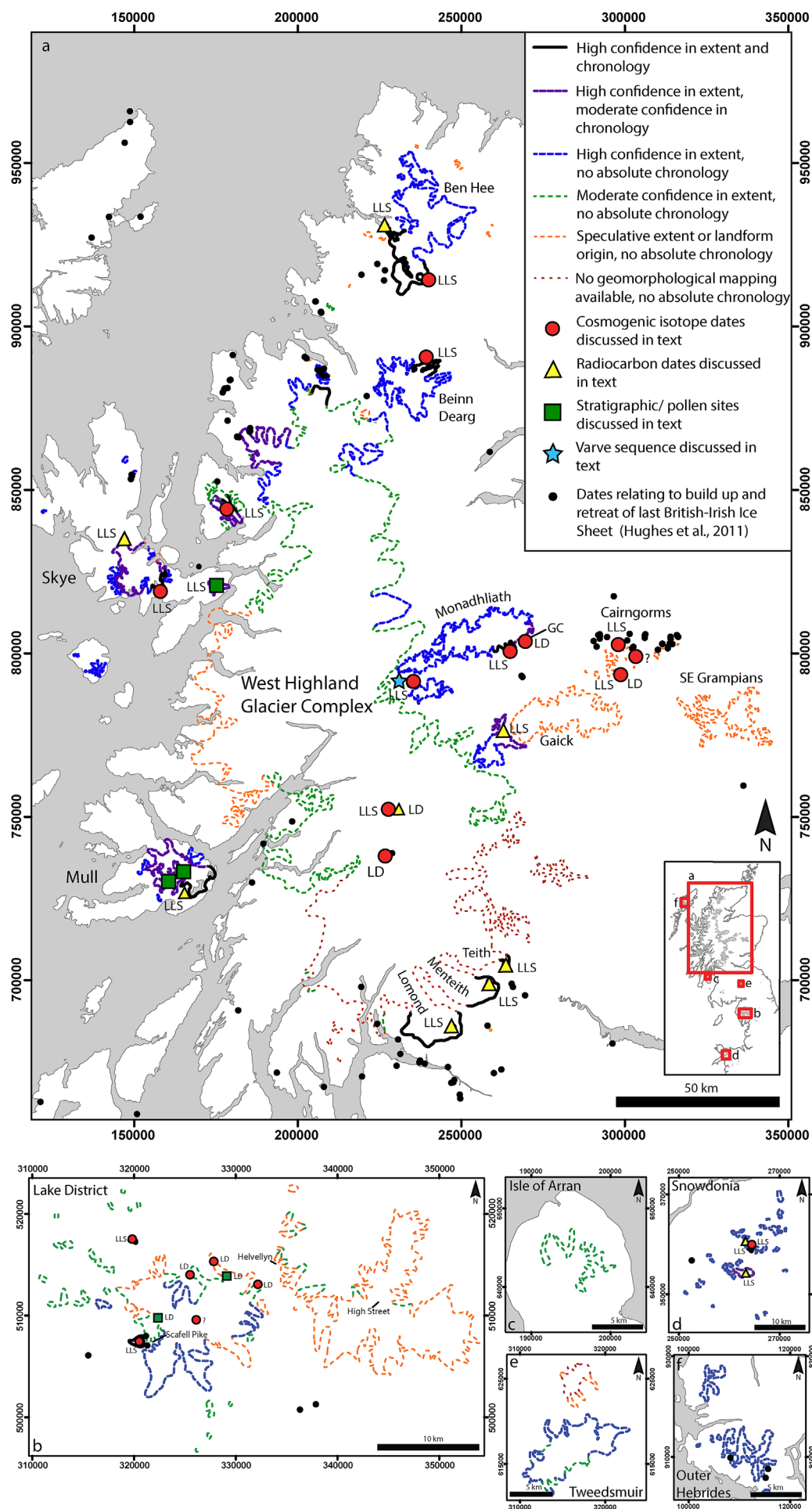


Figure 28. (Continued).

Figure 28. 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) Tweedsmuir, and (f) the Outer Hebrides. Across Britain are few sites where both the extent and the 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).

later phase of glacial readvance, *i.e.* the LLS (Ballantyne, 1989, 2002). At several sites, extensive river terrace sequences and large glaciofluvial features, such as kames or eskers, may be present in the lower valleys but are absent within the inferred glacier limits (Sissons, 1974; Lukas, 2006; Boston *et al.*, 2015). Lukas (2006) formalized these lines of evidence into a series of morphostratigraphic principles to identify landform assemblages of specific ages. This approach stresses the importance of using multiple lines of evidence when inferring a LLS age for landforms. These criteria have been applied in the north-west Highlands (Lukas and Lukas, 2006a, b), Monadhliath and Creag Meagaidh mountains (Finlayson, 2006; Boston, 2012a) and Tweedsmuir hills (Pearce *et al.*, 2014) 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 Holocene organic sediments are present. The age of these sediments has been supported by pollen analysis and radiocarbon dating (Pennington, 1977; Walker *et al.*, 1988; Benn *et al.*, 1992). However, Lowe and Walker (1976, 2015) highlighted that radiocarbon dates may be contaminated by younger or older carbon from a variety of sources, the latter being suggested by Small and Fabel (2016) to have affected the dates collected by Bromley *et al.* (2014). Furthermore, this technique can only be applied to locations where organic material is preserved and thus has been largely unsuitable for areas such as the north-west Highlands where bedrock remains close to the surface (Lukas and Bradwell, 2010).

Cosmogenic isotope dating is an alternative technique used 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 (Fig. 20) 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.

The use of tephrochronology has potential to resolve some of these dating issues. Microtephra layers can act as marker horizons within sediment sequences and thus can be used to precisely correlate palaeoenvironmental records between different locations and provide independent chronological control (Lowe *et al.*, 2016). Currently, 11 distinct tephra layers that span the Lateglacial and early Holocene have been detected in sediment sequences from various locations across Scotland (Pyne-O'Donnell, 2007; Lowe *et al.*, 2016), including the mid-LLS Vedde Ash (Lowe and Turney, 1997)

and the Abernethy Tephra that is associated with the end of the stadial (MacLeod *et al.*, 2015). Dating of these layers has produced a series of isochrons and has led to creation of a tephrostratigraphic framework that can then be applied to other (potentially undated) sites where the same tephras are present to provide precise, independent chronological control.

Determining the age of glacial landforms is particularly important when reconstructing the extent of former glaciers and numerous cases exist where changing age attributes have radically altered the palaeoglaciological reconstruction. For example, reinterpretation of the moraines and terraces at Achnasheen (Fig. 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 north-west Highlands (Bennett and Boulton, 1993a, b) than previously proposed (Sissons, 1982). 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, indicate a substantially greater ice extent than if they pre-date the stadial (McDougall, 2001). Where conflicting interpretations of geomorphological evidence exist, systematic dating of particular landforms can determine which reconstruction is most realistic. For example, cosmogenic isotope dating was used by Golledge and Hubbard (2005) 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. When assigning ages to landforms, studies now commonly use multiple lines of geomorphological evidence and a morphostratigraphic approach, often in conjunction with absolute dates which can be extrapolated to adjacent valleys displaying the same landform assemblages. However, as illustrated by Fig. 28, the limit of LLS glaciation has been robustly dated in very few locations and there remain several locations where the age of landforms remains poorly constrained, particularly where absolute dating is lacking. Even where dating has been attempted, the reliability of chronological evidence may remain controversial. For example, in the Lake District (Wilson *et al.*, 2013) and Gleann Chaorainn in the Monadhliath (Gheorghiu and Fabel, 2013; Boston *et al.*, 2015), dating of features on, or immediately inside, the empirical LLS limit returned pre-LLS ages that were subsequently challenged because 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. 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 ice limits are not clearly traceable from one valley to the next. Thus, in Fig. 28, only sections of the limit in the immediate vicinity of dates are deemed to be chronologically constrained.

While the stratigraphic and chronological evidence from Croftamie and Callander makes it very difficult to refute that

at least some glaciers underwent a significant readvance during the LLS, such sites are exceptional. Most of the 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 difficult to prove definitively 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, which was punctuated by major readvances (such as the Wester Ross Readvance) and 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.

Given the limitations of the current body of chronological evidence associated with LLS glaciation, further targeted research is required before the total extent of glaciation during the stadial can be inferred reliably. There are several locations where robust absolute dating could distinguish between very different hypotheses of the extent of LLS glaciers. For example, cosmogenic isotope dating of the outermost 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 Sissons' (1974, 1980a) reconstruction of valley glaciers nourished by ice accumulating on the plateau. Conversely, if these features pre-date the LLS, this could support the model of glaciation restricted to a few cirques favoured by Merritt (2004b). Similarly, absolute dating of the geomorphologically well-constrained limits in the West Drumochter Hills would help to determine whether these features were formed by a LLS icefield (Benn and Ballantyne, 2005) or during the retreat of the last British–Irish Ice Sheet (Merritt *et al.*, 2004a,b).

The limited chronological evidence in the Lake District poses probably the most significant barrier to establishing the extent of LLS glaciers in this region. Cosmogenic isotope 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). Similarly, work by Pennington (1978, 1996) supported the presence of LLS glaciers in the Lake District but the sites sampled cannot be used to test whether the style of glaciation was alpine (Sissons, 1980b) or plateau icefield (McDougall, 1998, 2001, 2013). Dating of landforms at the limits of the icefields reconstructed by McDougall (1998, 2001, 2013) would help to distinguish which of these two glaciation styles more likely occurred during the LLS. However, given that several cosmogenic isotope dates within this region seem to have been impacted by nuclide inheritance (Wilson *et al.*, 2013), it is possible that alternative dating techniques may be required. Chronological control 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, 2007a) and those generated by otherwise broadly accurate numerical modelling (Golledge *et al.*, 2008). 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 (2007a) and the more extensive icefield supported by

Gemmell (1973). Such dates would be 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, and the much more extensive glacial reconstruction that resulted from this (Bennett, 1991; Bennett and Boulton, 1993a), was based on the Achnasheen moraines appearing to form a continuous sequence with moraines attributed to the LLS further west (Bennett and Boulton, 1993a,b). It was also 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 absolute chronological control on the limits at Achnasheen would represent a significant step in constraining the overall extent of LLS glaciers within this region. Likewise, while recent bathymetric surveying (McIntyre *et al.*, 2011) suggests that LLS glaciers extended further into the lochs along the west coast of Scotland than previously thought, only by dating the outermost of these features will it be possible to confirm whether they 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 synchronicity 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 when others peaked much later (Ballantyne, 2012). The conflicting dates from Rannoch Moor (Bromley *et al.*, 2014; Small and Fabel, 2016) further complicate this picture of retreat. Proglacial lake sediments containing annually deposited varved layers can provide extremely high-resolution evidence of glacier behaviour. Using the Lochaber Master Varve Chronology, Palmer *et al.* (2010) were able to reconstruct the advance and retreat of the Roy and Spean glaciers and the duration of the resulting ice-dammed lakes, although Devine and Palmer (2017) suggested that only the overall duration of the lake system, rather than that of individual lakes, is reliable. However, there are only a minimal number of locations where this type of analysis is possible and it is still affected by uncertainty, either during varve counting (Lowe *et al.*, 2008) or when used in conjunction with existing absolute dates.

Only through a systematic and comprehensive programme of absolute dating, such as that currently being undertaken for the BRITICE-CHRONO project, will it be possible to determine the timing and synchronicity of LLS glaciation. Most of the existing dating has been conducted near the termini of the LLS glaciers (e.g. MacLeod *et al.*, 2011; Palmer *et al.*, 2010) 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.

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 while others remain largely unmapped. Future work must seek to address such shortfalls to allow accurate reconstruction of the extent of LLS glaciation in all areas. Particular areas to target are the south-west 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. While some studies have produced high-resolution data (e.g. Rose, 1980, 1981; Ballantyne, 2002; 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 an improved scope for detailed advanced analysis. The specific areas to target include the Gaick Plateau and the south-east 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. The steep precipitation gradients imposed on the model to achieve this 'best fit' scenario accord well with palaeoclimatic inferences from glacier reconstructions. 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 cooling. This has led to discrepancies between modelled and empirical LLS glacier limits and uncertainty as to whether LLS ages have been assigned incorrectly to some outlets, for example the Callander lobe (Golledge *et al.*, 2008). While 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.

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 used before assigning an age to features, there continues to be a paucity of absolute dates, inhibiting our understanding of the timing and synchronicity of glaciation during the LLS. While the number of absolute ages continues to grow, implementation of a comprehensive dating programme will be necessary 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.

A speculative reconstruction of the lateral extent of the LLS glaciers at their maximum positions has been compiled from empirically based reconstructions from the published literature. This is intended as a tool to guide future work to refine this reconstruction. The shapefile provides a useful tool for other researchers to investigate controls on LLS glacier formation at a larger scale than previously possible.

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Abbreviations. AMOC, Atlantic Meridional Overturning Circulation; DTM, digital terrain model; ELA, equilibrium line altitude; LGM, Last Glacial Maximum; LLS, Loch Lomond Stadial; YD, Younger Dryas.

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