A Younger Dryas plateau icefield in the Monadhliath, Scotland, and implications for regional palaeoclimate

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Abstract

A record of Younger Dryas glaciation in Scotland is well established. However, the role of the Monadhliath, a significant plateau area extending over 840 km^2 in central Scotland, has never been investigated systematically. We present the first systematic glacial geomorphological mapping across the whole region, which has led to the identification of hitherto-unrecorded glacial and associated landforms. The spatial distribution of these landforms indicates that the last phase of glaciation in the area was that of a local plateau icefield. In addition, a clear morphostratigraphical signature provides a strong indication that the icefield dates to the Younger Dryas (12.9-11.7 ka), which is supported by numerical ages in the southeast of the study area. Based on the geomorphological evidence and 2D glacier surface profile modelling, a 280 km^2 icefield is reconstructed. A novel approach is introduced to quantify plateau icefield thickness for equilibrium line altitude (ELA) and palaeoprecipitation calculations, resulting in greater overall data confidence compared to traditional reconstruction methods. The ELA for the whole icefield is calculated to be 714 ± 25 m, whilst the ELAs of individual outlet glaciers range from 560 m in the west to 816 m in the east, demonstrating a significant W-E precipitation gradient across the region during the Younger Dryas. These ELAs compare well with those calculated for Younger Dryas ice masses reconstructed in neighbouring regions and are in good agreement with overall precipitation patterns suggested for Scotland during this time. Whilst the total amount of precipitation calculated from these ELAs is highly dependent on the method used, irrespective of this, the study suggests a more arid Younger Dryas climate in the region compared to the present day.

Keywords

plateau icefield, Younger Dryas, Scotland, glacier reconstruction, palaeoclimate

1. Introduction

The Younger Dryas Stadial, equivalent to Greenland Stadial-1 (GS-1; 12.9-11.7 ka) (Rasmussen et al., 2006; Lowe et al., 2008) and correlated with the Loch Lomond Stadial (LLS) in Scotland, was a period of rapid climate change at decadal and centennial time scales (Anderson, 1997; Tarasov and Peltier, 2005; Lukas, 2011). Understanding atmospheric and oceanic drivers for rapid climate change is important in the prediction and evaluation of future climate change scenarios, especially in the amphi-North Atlantic region where detailed records document the nature of this change (e.g. Bakke et al., 2009). Scotland is a key area in this respect in that it allows linking of the terrestrial signature of Younger Dryas climatic change, manifest in numerous well-preserved glacial sediment-landform assemblages, to records from elsewhere. The evidence preserved in Scotland can be regarded as unique in a

European context, largely because of absent to very limited post-depositional modification of glacial and associated sediment-landform assemblages (e.g. Benn and Lukas, 2006; Golledge, 2010). However, despite a long history of research in the Lateglacial record in Scotland, there are still numerous areas in Scotland that have not been investigated in detail and of which very little is known about the palaeoglaciology, making regional comparisons of glacier dynamics and palaeoclimate difficult and our understanding of events during the Last Glacial-Interglacial Transition (LGIT) very incomplete.

The Monadhliath (Fig. 1) are one such area, having received a very limited proportion of research attention over the last 100 years. Previous research is generally confined to the southern and eastern parts of the region only, and limited to work by British Geological Survey (BGS) officers at the beginning of the 20th Century (Barrow et al., 1913; Hinxman and Anderson, 1915), J.A.T. Young in the 1970s (Young, 1977, 1978), and more recently by Auton (1998), Phillips and Auton (2000), Gheorghiu et al. (2012) and Trelea-Newton and Golledge (2012). Extensive research has however been undertaken in the adjoining region to the southwest around Glen Roy where the extent of glaciation is well established (e.g. Sissons, 1978, 1979a; Sissons and Cornish, 1983; Peacock, 1986; Lowe and Cairns, 1991; Fabel et al., 2010; Palmer et al., 2010, 2012), although the extent and timing of glaciation in Glen Turret is still subject to debate (cf. Benn and Evans, 2008; Peacock, 2009).

Recent work by Boston (2012a, b) has resulted in the first systematic mapping of the Monadhliath and the adjoining area northeast of Glen Roy (Fig. 1). This regional assessment of glaciogenic landforms and sediments provides a new insight into ice mass fluctuations and dynamics in this critical central part of the Scottish Highlands. It adds to other recent work in the region, which has hinted at the presence of a Younger Dryas plateau icefield based on numerical modelling (Golledge et al., 2008) and limited local geomorphological evidence from selected valleys (e.g. Gheorghiu et al., 2012; Trelea-Newton and Golledge, 2012), despite a traditional belief that major ice masses did not build up in the area during the Younger Dryas (Sissons, 1979b). The lack of a previous regional assessment of the field evidence, likely also hampered by the persistent paradigm that localised cirques, and not adjacent plateau surfaces, were the sole sources of any former ice masses, has meant that previous reconstructions of any ice mass have been highly tenuous; therefore, the extent and dynamics of any such ice mass in the area, if any, are currently unknown. The aims of this paper are therefore to 1) present geomorphological evidence for the presence of local plateau ice during the LGIT, 2) establish a relative chronology for glacial events in the region based on an examination of the morphostratigraphical evidence, 3) reconstruct the extent of ice masses relating to the last phase of glaciation, 4) provide estimates of former precipitation in the Central Scottish Highlands during this time based on the ice mass proportions and 5) discuss wider regional implications for palaeoclimate during the Younger Dryas.

2. Study Area

The Monadhliath are primarily underlain by various Late Precambrian psammites and semipelites belonging to the Grampian Group, of which the two oldest subgroups, the Glenshirra and Corrieyairack successions, dominate (Stephenson and Gould, 1995). Pelitic to semipelitic schists from the younger Appin Group are also exposed in Glen Roy (Phillips and Key, 1992), and the older Pre-Dalradian Central Highland Migmatite Complex crops out to the north of Newtonmore and Kincraig, and is termed the Glen Banchor succession (Robertson and Smith, 1999; Smith et al., 1999; Strachan et al., 2002). Igneous intrusions of granite, microdiorite and granodiorite of predominately Silurian age also occur in the region, forming the Corrieyairack, Allt Crom and Findhorn plutons (Stephenson and Gould, 1995).

Numerous high-level psammitic and granitic erratics occur, mostly of local origin, although some granitic erratics are tentatively suggested to be of Rannoch Moor provenance (Jarman, 2013).

The study area comprises an upland area of plateau covering approximately 840 km^2 of the central Highlands of Scotland, bounded to the north by the Great Glen and to the south by Strathspey (Fig. 1). The plateau consists of rounded summits and has been dissected into twenty-five main catchments. The overall plateau slopes northwards and its altitude ranges from c. 900 m in the south, with individual summits as high as 945 m (Carn Dearg, NH 636 024), to c. 600 m in the north. As a result, the main watershed runs from west to east across the southern edge of the plateau, and this asymmetry is manifest in short, steep catchments on the south side of the plateau, with the majority of the plateau being drained by rivers flowing northwards or eastwards within much larger catchments. Valleys descending from the plateau to the south tend to have steep backwalls separating the valley floor from the plateau above, whilst catchments in the north gently rise onto the plateau, often with no backwall.

3. Methods

Geomorphological mapping was undertaken through remote-sensing, using the NEXTMap Britain (NEXTMap) dataset (Intermap Technologies, 2007) and panchromatic aerial photographs, and a spatially extensive 12-week field campaign spread over two field seasons. Using three different approaches to mapping allowed the advantages of each to be combined to produce a detailed and accurate geomorphological map, with robust genetic interpretations. Further details on this approach can be found in Boston (2012a). Sedimentological analysis, following procedures outlined by Evans and Benn (2004), was also used alongside the geomorphological mapping to provide additional information on the processes leading to landform genesis. All methods relating to establishing relative ages, glacier reconstruction, equilibrium-line altitude estimation and calculations of palaeoprecipitation are covered in detail in the respective sections.

4. Results

4.1. Geomorphological evidence

Geomorphological mapping focussed on recording landforms that were pertinent to glacier reconstruction such as moraines, meltwater channels, drift limits and periglacial features. These landforms have been described in detail and presented in a geomorphological map by Boston (2012a, b). Due to the large area covered by the study, we present a summary of the glacial geomorphology of the region here, alongside map extracts (Figs. 2 and 3), a common approach when covering larger areas that were previously terra incognita (cf. Lukas and Bradwell, 2010).

A similar assemblage of landforms is found within many of the valleys that radiate from the Monadhliath plateau. Moraines and ice-marginal meltwater channels are the most dominant features and the upper parts of many valleys contain a series of small (approx. 5 m high, < 30 m wide), densely-spaced, sharp-crested moraines, separated by intervening ice-marginal meltwater channels ('Type 1' moraines (Boston, 2012a)). The proportion of moraines compared to meltwater channels varies between valleys (see Section 5.1), but in general all of the larger valleys contain clear sets of moraines. Particularly good examples of this landform assemblage occur within Coire Larach, Glen Chonnal and Corrie Yairack in the west (Figs. 1 to 3), Coire Easgainn and the Findhorn valley to the north, and within Gleann Ballach and Gleann Fionndrigh in the southeast (Figs. 1, 2). In many places, the moraines are heavily dissected into mounds and ridges, but their crestlines can be linked to reconstruct former ice fronts following established approaches (cf. Benn and Lukas, 2006, and references therein). Many of the lateral and latero-frontal moraines extend back into the highest parts of the catchments and partly onto the plateau, especially on the northern side where backwalls are absent (Fig. 2). Moraines are also found on the plateau, particularly in the east, concentrated within topographic lows (Boston, 2012a) (Fig. 2A).

Periglacial features are present within several valleys and on plateau summits. These include solifluction lobes, blockfields and talus slopes. In the upper parts of several valleys, talus slopes end abruptly, mid-way down the valley side, often culminating at the upper limit of moraines. In several of the southeastern valleys, this lower talus limit can be followed laterally up-valley until the talus ends and the limit continues as a sharp downslope boundary of a solifluction sheet (Fig. 2D). Blockfields are a prominent feature on many of the higher plateau summits (typically above 800-850 m), particularly in the southeast corner of the study area.

The lower parts of many valleys are characterised by a different landform signature. In these areas moraines are typically larger, both in terms of their heights (often > 10 m) and widths (up to 70 m), and possess more rounded crestlines (‘Type 2’ moraines (Boston, 2012a); Figs. 2 and 3). Others form subdued mounds on the valley floor. These moraines are also spaced further apart, often sporadically, and do not always have distinct ice-marginal meltwater channels between them. Prominent river terraces (> 3 m in height) are also present within these parts of the valleys, sometimes associated with alluvial fans, and thick talus accumulations cover many valley sides (Fig. 2A).

The spatial distribution of features within many of the valleys provides unequivocal evidence that the ice within them was sourced from the plateau. This is manifest in the location of many lateral moraines, some of which extend partly onto the plateau in catchments that lack a backwall, whilst a number of moraines occur on the plateau itself. Ice-marginal meltwater channels are also found at the plateau edge in several locations, again indicating ice flow radiating from the plateau. Additionally, distinct downslope limits of solifluction lobes in the southeast corner of the Monadhliath lead onto the plateau and provide a clear indication of the former ice surface (cf. Sissons et al., 1973; Sissons, 1977; Lukas and Bradwell, 2010).

4.2. Chronology

Numerical ages have been published for three valleys in the southeastern part of the study area by Gheorghiu et al. (2012) and indicate a Younger Dryas age for the most recent glacial event. However, in the absence of numerical age estimates in the remainder of the Monadhliath, we present a relative chronology for palaeoglacial events in the region. Our relative chronology is developed on the basis of morphostratigraphy, which uses “the spatial relationship between individual landforms to assign them to events or periods” (Lukas, 2006, p.721). Since this approach forms the basis of our reconstruction, the key components are explained in some detail below. We then discuss this chronology in the light of previously published absolute dates.

4.2.1 Identifying a relative chronology

Although a morphostratigraphical approach has been used in a large proportion of former ice mass reconstructions in Scotland for several decades, it has not always been formally acknowledged (e.g. Sissons, 1974; Ballantyne, 2002a, 2007a, b; Benn and Ballantyne, 2005; Finlayson, 2006; Lukas, 2006; Lukas and Bradwell, 2010; Finlayson et al., 2011). This approach has been particularly effective at spatially correlating mapped ice limits to chronologically well-constrained limits in areas where there is limited numerical dating. This is a practical solution within a context where it is usually unrealistic to be able to ascertain numerical dates for every glacier limit (e.g. Lukas and Bradwell, 2010; Finlayson et al., 2011).

Recently, Lukas (2006) formalised this approach within the context of LGIT glaciation by identifying a set of geomorphological criteria that can be used to test whether a particular landsystem is of Younger Dryas age. This is based upon the long-standing realisation that strongly contrasting sediment-landform associations have been found inside and outside prominent glacier limits across most areas of Upland Britain and elsewhere (Lukas, 2006, and references therein).

Criteria include differences in the type and frequency of moraines, number of river terraces inside and outside clear moraine sequences, the distribution/orientation of glaciofluvial landforms such as eskers and kames, the number, distribution and arrangement of glacially-transported boulders or boulder clusters, the thickness of sediments on sediment-covered slopes, the maturity of talus slopes, the distribution of periglacial landforms and the location and termination of palaeoshorelines in areas near the coast (Lukas, 2006).

In any catchment, contrasts in the type and frequency of moraines (e.g. Types 1 and 2 here) are taken as the first indication of a clear glacial limit at the outermost moraine of the upper type (in this case the outermost Type 1 moraine) (cf. Lukas, 2006). If such a limit *coincides* with a contrast in, for example, the number of river terraces, this is taken as further confirmation that the limit in question marks a distinct palaeoglacial event. A contrast in river terraces has frequently been reported to manifest itself in several river terraces outside this outermost moraine, with an abrupt decline in the number, or complete disappearance, of river terraces inside, i.e. upstream of, this limit (Lukas, 2006, and references therein). The explanation for a difference in the number of river terraces and how this relates to glaciation has been given elsewhere in more detail (Lukas, 2006), but has been recognised as a key morphostratigraphical tool for > 100 years (Penck and Brückner, 1901/1909; Bridgland and Westaway, 2008). Sometimes, a notable difference in the thickness of sediments and/or talus on slopes inside and outside this limit can also be found, and this has been demonstrated to relate to the time available for subaerial (frost) weathering and periglacial activity on slopes (e.g. Ballantyne, 1991). Furthermore, such limits can often be traced obliquely across slopes from clear outermost moraines in the valley bottom, strongly indicating that they represent the remnants of a former ice lobe; consequently, such vertical limits can be utilised in glacier reconstruction by aiding interpolation towards the former accumulation area (e.g. Lukas, 2010, p.187). In this example, three lines of evidence (moraines, river terraces and differences in sediment thickness on slopes) converge to enable the differentiation of one distinct glacier limit.

Lukas (2006) reviewed a comprehensive body of literature dating back to the 1970s that supports this approach, and, to date, this formalised approach has been utilised as a first-order approximation to *infer* a Younger Dryas age in two independent areas (e.g. Lukas, 2005, 2006; Finlayson and Bradwell, 2008) and has been used to guide later numerical dating programmes to test the hypotheses put forward by Lukas (2006). In both cases, numerical dating has confirmed the hypotheses of a Younger Dryas age of lake sediments accumulated just outside the Younger Dryas limits (Lukas and Bradwell, 2010; Lukas et al., 2010) and key boulders on the outermost moraines concerned (Lukas and Bradwell, 2010; Finlayson et al., 2011). It has therefore been powerfully demonstrated that this approach is robust *if used in its intended way*. For example, a key component of these contrasting sediment-landform associations is that *their recognition formed the basis for establishing absolute age control* at relevant sites, so, in other words, only after detailed mapping and identification of such contrasts would a dating programme commence. Where this key geological principle of *'relative prior to absolute dating'* has been ignored, as we will argue and demonstrate below in Section 4.2.2, numerical ages conflict with the geomorphological interpretation, because the samples have been taken out of context, i.e. without a clear understanding of the landsystem that was sampled.

Following the criteria outlined above and by Lukas (2006), two different glacial landform assemblages are recognised within the upper and lower parts of the valleys in the Monadhliath, and indicate deposition during two phases of local glaciation. The landform signature in the upper parts of a large number of valleys is very similar to areas elsewhere in Scotland that are inside dated Younger Dryas limits, and therefore we argue that the most recent phase of glaciation in the Monadhliath also occurred at this time. Clearly identifiable limits can be found in most major valleys in the study area, and as discussed above, limits are most confidently assigned when several lines of evidence converge at the same location (Table 1). In minor valleys where abundant direct evidence constraining ice limits was not available, ice limits are based on spatial relationships with neighbouring valleys, and indirect evidence mapped such as the configuration of meltwater channels.

Assignment of landforms in the upper parts of valleys to the Younger Dryas implies that glacial features in the lower parts, notably the Type 2 moraines, relate to an older phase of local plateau icefield glaciation and/or deposition by regional ice in the Spey Valley. We suggest that this is most likely to have occurred at the end of the Dimlington Stadial (correlated with Greenland Stadial 2 (GS-2); Lowe et al., 2008) following thinning and retreat of regional ice cover, as prior to this the region was submerged beneath the Last British-Irish Ice Sheet (cf. Clark et al., 2012).

This relative chronology is supported by morphostratigraphic correlation of the glacier limits in the southwest of the study area to the Younger Dryas ice-dammed lake systems of Glen Roy. This association has previously been suggested by Johnson-Ferguson (2004) and Benn and Evans (2008) through the presence of moraines on top of subaqueous grounding-line fans at the head of Glen Turret. These fans emanate from the Teanga Bige and Teanga Mòire catchments and are associated with deposition within the 350 m lake (Boston, 2012a, b) (Fig. 3). Here, we argue that these moraines represent the maximum limit of Younger Dryas ice in Glen Turret, based on the same morphostratigraphic criteria applied to the rest of the study area and supported by the altitudes of Younger Dryas limits (360-465 m) in all neighbouring valleys (see Boston et al., 2013, for further discussion). As found in other parts of the Monadhliath, there is strong evidence in the southwest for a more extensive phase of local plateau glaciation prior to the Younger Dryas. This is particularly apparent in valleys such as Glen Buck, Glen Chonnal, Glen Shesgnan and Corrie Yairack where there are numerous large, sporadically-spaced moraines, often surrounded by a prominent river terrace. We therefore argue that the Glen Turret Fan was also deposited during this earlier phase of plateau glaciation. This suggests that a similar ice-damming event to that during the Younger Dryas may have occurred during this time. This is not unreasonable given that this phase of glaciation most likely occurred during an 'unzipping' of local and regional ice at the end of the Dimlington Stadial, which would provide prime conditions for ice-damming as has been proposed elsewhere along the Spey Valley (e.g. Brazier et al., 1998; Phillips and Auton, 2000; Everest and Kubik, 2006; Auton, 2013).

4.2.2 Previous absolute dating work

Boulders sampled from within three valleys in the southeast of the Monadhliath have been analysed for ^{10}Be surface exposure dating (SED) (Gheorghiu et al., 2012), yielding individual age estimates between 9.7 ± 0.9 ka and 12.5 ± 1.2 ka for the most recent phase of deglaciation, using a global production rate in the CRONUS Earth calculator. The validity of this production rate has recently been questioned (Balco et al., 2009; Putnam et al., 2010; Fenton et al., 2011), and these dates have subsequently been recalculated to between 11.3 ± 0.7 ka and 14.9 ± 0.9 ka, using a local production rate of 3.92 ± 0.18 a/g/yr (Gheorghiu and Fabel, 2013).

The results of this work broadly support the relative chronology of events identified above, in that the numerical ages support the assertion that the most recent phase of glaciation in the Monadhliath occurred during the Younger Dryas. However, there are, in some cases large, discrepancies between the maximum limits of the Younger Dryas glaciers reconstructed by Gheorghiu et al. (2012) and those presented here (Fig. 4). We argue that this results from a disregard of the geomorphological and morphostratigraphic relationships by Gheorghiu et al. (2012), which adversely affected their sampling strategy and interpretation of their results, as elaborated on below.

One of our key criticisms of the work is that the glacier limits reconstructed by Gheorghiu et al. (2012) appear to be based entirely upon where their samples were taken rather than on the clear geomorphological contrasts described above. For example, Gheorghiu et al. (2012, p.

142) state that “the YD limit of the glacier in Gleann Lochain is indicated by the deposition of Moraine B at 11.8 ± 1.1 ka BP” (recalculated to 13.6 ± 0.8 ka; Gheorghiu and Fabel, 2013) (Moraine B located at position B in Fig. 4). However, by only sampling this moraine, they have failed to test whether this moraine, rather than other moraines down valley, represents the Younger Dryas limit. We argue that Moraine A (position A in Fig. 4) is part of the same landform assemblage and that its orientation indicates deposition by ice sourced from Gleann Lochain, debouching as a piedmont lobe into Glen Banchor, rather than by regional ice as suggested by Gheorghiu et al. (2012). In addition, the morphostratigraphy indicates that moraines A and B belong to a different palaeoglacial event to those moraines further up valley; moraines A and B belong to an assemblage of large (up to 10 m) sporadically-placed moraines with rounded crestlines which are surrounded by a well-defined river terrace (Type 2), in contrast to moraines in the upper part of the valley that are closely-spaced, with a dense network of intervening meltwater channels and no well-developed river terraces (Type 1). We therefore argue that Moraine B is more likely to pre-date the Younger Dryas and suggest that the date is too young, as a result of exhumation or toppling during moraine stabilisation (e.g. Lüthgens and Böse, 2012). Given that the revised date for this moraine of 13.6 ± 0.8 ka largely predates the beginning of the Younger Dryas, the most reasonable and likely interpretation is that the moraine was deposited during deglaciation towards the end of the Dimlington Stadial.

In the neighbouring valley of Gleann Ballach, two boulders sampled on top of a large ridge (H; Fig. 4) give a mean age of 11.2 ± 1.1 ka (recalculated to 13.2 ± 0.8 ka; Gheorghiu and Fabel, 2013). Whilst Gheorghiu et al. (2012) acknowledge a glaciolacustrine origin for the ridge, caused by an ice-damming event by Late Devensian regional ice in Glen Banchor, their continued use of exposure ages to define glacier limits has led to placement of the Younger

Dryas limit at the northern side of this ridge. The authors describe the glacier as having "abutted the north side of the pre-existing lake deposit" (p.142), failing to recognise the abundant evidence for glaciotectonic disturbance and possible tectonic thickening of the sequence, predominantly in a northeasterly direction from ice located in Glen Banchor to the south (Auton, 2013).

Further evidence of this regional ice advancing northwards into Gleann Ballach is found immediately to the north of the glaciolacustrine ridge, in the form of a series of subdued lateral moraines (I; Fig. 4). These moraines terminate as they are cross-cut by younger moraines from local ice moving southwards down Gleann Ballach. The outermost local lateral moraine therefore depicts the maximum position that a subsequent local glacier advance reached, indicating that it did not reach as far south as the glaciolacustrine ridge. This is supported by the presence of a significant terrace, which continues upstream of the ridge, until it reaches the outermost local moraine. This morphostratigraphical evidence indicates a Younger Dryas age for the local moraines in agreement with the remainder of the surface exposure ages obtained by Gheorghiu et al. (2012). However, based on this morphostratigraphical and sedimentological evidence, we argue that the surface exposure ages for boulders on the glaciolacustrine ridge should be interpreted as being too young, and the ridge should not have been used to mark the maximum limit of a Younger Dryas glacier. A likely cause for the young age, although dismissed by Gheorghiu et al. (2012), is that fine-grained (i.e. glaciolacustrine) sediments can easily be entrained by strong katabatic wind in glacier forelands, leading to surface lowering (e.g. Reuther et al., 2005; Schaller et al., 2009; Lukas et al., 2012) and boulder exhumation.

Lastly, a major discrepancy occurs between the surface exposure ages and the morphostratigraphical signature in the upper part of Gleann Chaorainn (Fig. 4). Here, both Boston (2012b) and Trelea-Newton and Golledge (2012) interpret the closely-spaced moraines and the absence of a prominent terrace or extensive periglacial features (cf. Lukas, 2006) in the upper part of the valley to be indicative of deposition during the Younger Dryas. However, the three ages obtained from boulders on moraines (J; Fig. 4) (interpreted by Gheorghiu et al. (2012) as a fluvially-dissected surface of till) are widely scattered: 16.2 ± 1.5 ka, 14.1 ± 1.3 ka and 11.6 ± 1.0 ka (recalculated to 19.3 ± 1.1 ka, 16.8 ± 0.9, 13.7 ± 0.8; Gheorghiu and Fabel, 2013). Gheorghiu et al. (2012) dismiss the youngest date as the result of either exhumation or overturning and therefore do not reconstruct a Younger Dryas glacier in this valley. Thus, there is inconsistency both between the three dates and the way in which Gheorghiu et al. (2012) interpret these ages compared to those in the other two valleys, with no clear reasoning provided. Based on morphostratigraphical relationships, we argue that these dates are too old and may have been affected by nuclide inheritance. This is certainly possible, given that the orientation of the moraines and meltwater channels indicates a short, thin outlet glacier, which may not have had the erosive capacity to 'reset the nuclide clock' (cf. Heyman et al., 2011).

Based on the similar landform assemblages and clear morphostratigraphic signature for two palaeoglacial events across the three valleys discussed above and Gleann Fionndrigh from which no numerical ages were obtained, we have reconstructed glaciers in the heads of all four valleys during the Younger Dryas. This fits with the geomorphological evidence across the Monadhliath for a Younger Dryas plateau icefield. In comparison, as a consequence of their reliance on surface exposure ages to define Younger Dryas glacier limits, Gheorghiu et al. (2012) present a very inconsistent reconstruction of Younger Dryas ice in this area. The

Younger Dryas maximum in Gleann Lochain is depicted at approximately 470 m OD near the mouth of the valley, whilst ice is restricted to 800 m OD on the plateau to the north of Gleann Chaorainn, resulting in a sharp rise in their calculated ELAs (Fig. 4).

In summary, our main criticism of the work by Gheorghiu et al. (2012) is the lack of understanding of the landsystem and morphostratigraphy in the study area, which should always form the precursor to a landform dating programme (Lowe and Walker, 2014). Only from such a solid foundation would a systematic and well-targeted process of sampling for chronological reconstruction have been possible. Gheorghiu et al. (2012) appear to have followed an emerging methodology in which the sampling and analytical procedure of SED is valued more than the geomorphological context that it is being employed to decipher. Where this has happened elsewhere (e.g. Southern Scandinavian Ice Sheet margin: Rinterknecht et al., 2005, 2006; Houmark-Nielsen et al., 2012; Lüthgens et al., 2011; New Zealand: Schaefer et al., 2009; Winkler and Matthews, 2010; Kirkbride and Winkler, 2012), it has invariably led to premature chronological reconstructions and resulted in unnecessary confusion rather than progress. We welcome rigorous testing of the relative chronology presented here, based on a detailed sampling strategy designed to actually test the maximum Younger Dryas glacier limits that are strongly indicated by the morphostratigraphical signature.

4.3. Younger Dryas plateau icefield reconstruction

The identification of Younger Dryas glacier limits, combined with geomorphological evidence that ice was sourced from the plateau, allowed the former plateau icefield to be reconstructed within ESRI ArcGIS. Where distinct lateral moraines, upper limits of sediment (drift limits), and sharp lower boundaries of talus slopes and solifluction lobes occur, the upper limit of the former ice surface could be confidently reconstructed and extrapolated to help constrain the ice surface on the plateau and in neighbouring valleys.

In several areas geomorphological evidence was lacking, so that constraining ice thickness across much of the plateau was near impossible to estimate based on the geomorphological evidence alone. This was in part due to uncertainty as to what the lower boundaries of the blockfields and solifluction lobes on the plateau summits represent. Blockfields are relict features that are proposed to have formed as far back as the Neogene (Nesje, 1989; Rea et al., 1996; Whalley et al., 1997, 2004; Sumner and Meiklejohn, 2004; Fjellanger et al., 2006; Paasche et al., 2006), although recent work indicates they may have predominantly formed through physical weathering, such as frost wedging, under periglacial conditions during the Quaternary (Ballantyne, 1998, 2010a; Goodfellow et al., 2009; Goodfellow, 2012; Hopkinson and Ballantyne, 2014). Irrespective of this debate, the extensive blockfields on many of the summits in the Monadhliath would not have formed on ice-free summits during the Younger Dryas alone (sensu. Fabel et al., 2012). Their presence would therefore suggest that these locations have not been subject to extensive periods of glacial erosion either during the Younger Dryas or earlier phases of glaciation. This indicates that the blockfields either remained as nunataks above any warm-based ice (e.g. Ballantyne, 1997; Ballantyne et al., 1997), or were covered and protected by cold-based ice (e.g. Whalley et al., 1981; Gellatly et al., 1988; Kleman, 1994; Rea et al., 1996; Fjellanger et al., 2006; Ballantyne, 2010b; Fabel et al., 2012). Similarly, although the solifluction lobes could have formed during the Younger Dryas, the absence of a sharp lower boundary to any of the lobes on the plateau summits makes it difficult to use them to delineate the upper ice surface, in contrast to the solifluction lobes at the plateau-valley transitions in the southeast of the study area. This lack of a sharp boundary indicates that the solifluction lobes could also either be relict features that were covered by cold-based ice during the Younger Dryas, or that they are still active, as conditions for solifluction have prevailed during the Holocene in areas above 550 m in Scotland (Ballantyne, 2008).

In order to resolve the issues concerning plateau ice thickness and to examine the assumption that all valley glaciers were connected to ice on the plateau above them, two 2D glacier surface profile models were applied in combination with the available geomorphological evidence. This approach was used to guide ice thickness in the reconstruction and ensure that the minimum and maximum boundaries were glaciologically feasible. Like the majority of glacier reconstructions, particularly for Younger Dryas glaciers in Scotland, the method was still predominantly subjective due to the emphasis on the interpretation of the geomorphological evidence, but examination of the model outputs allowed a more objective element to be included, particularly in the areas where geomorphological evidence was lacking.

The first model by Benn and Hulton (2010) is a ‘perfectly plastic’ flowline model adapted from Nye (1951, 1952) and van der Veen (1999), and various versions of this model have been used in previous glacier reconstructions (e.g. Schilling and Hollin, 1981; Locke, 1995; Rea and Evans, 2007; Carr and Coleman, 2007; Vieira, 2008). The authors specifically highlight its use for defining otherwise unconstrained ice surfaces on plateaux. Inputs to the model are bed topography and ice surface elevations, where known along the glacier centre line. A shape factor (f) is also required to estimate the effect of lateral drag from the valley sides. Basal shear stress is then altered to initially constrain the modelled profile to any known ice surface elevations before being used to extrapolate the profile to areas where geomorphological evidence is absent.

A key problem with this model is the wide range of shear stresses that can be used to generate the ice surface profile. In modern valley glaciers, basal shear stresses are generally estimated to range between 50 and 150 kPa (Benn and Evans, 2010). However, there is a large difference in the heights of the modelled surface profiles that are produced using the lower and upper extremes of this range and the valley topography often causes a greater constraint on the scope of possible ice surface elevations; in many valleys geomorphological evidence dictates that the ice surface must have extended onto the plateau and/or geomorphological evidence in neighbouring valleys dictates maximum ice thickness. Therefore, shear stresses of a similar magnitude to those in the geomorphologically-constrained part of the reconstructed glacier were used to extrapolate the former glacier surface into unconstrained areas. These were typically around 50 kPa. The shear stresses were increased if the topography steepened and lowered if the gradient decreased, including an adjustment to zero near the ice divide (Benn and Hulton, 2010). The Benn and Hulton (2010) model was therefore used with most confidence to reconstruct the minimum and maximum surface elevations of the upper parts of outlet glaciers that were well-constrained for the majority of their profile (Fig. 5).

The second model used was devised by Ng et al. (2010) to test whether a col was formerly submerged by ice and, if so, the minimum height of the ice above the col. This is achieved through examination of variations in the curvature of modern glacier surface profiles using eqn. (1) (Nye, 1951, 1952).

$$h(x) = C\sqrt{x} \tag{1}$$

and the constant C as:

$$C = \left(\frac{2\tau_0}{\rho g}\right)^{\frac{1}{2}} \tag{2}$$

where x is the horizontal distance from the ice margin in an up-glacier direction, h is the height above the glacier margin, τ_0 is the basal shear stress, ρ is the ice density and g is gravitational acceleration.

Ng et al. (2010) favour this equation, which defines the parabolic surface profile of a glacier resting on a flat bed rather than a slope, to overcome the problem of extrapolating the bed topography from beneath modern glaciers. The constant C describes how 'stiff' the flow of ice is and this value increases with ice stiffness. Ng et al. (2010) calculate two values for C from the surface profiles of 200 ice masses (Fig. 6). C^* describes the whole of the glacier profile and $\tilde{C}$ 'best fits' eqn. (2) to the glacier profile (see Ng et al., 2010 for details). Based on the range of C^* and $\tilde{C}$ values within the modern dataset, Ng et al. (2010) define their minimum values based upon glacier length, since the influence of bed topography on the surface profile increases as glacier length decreases. Ng et al. (2010) specify that the minimum value for C^* within valley glaciers is 5.2 $m^{1/2}$ and demonstrate that this can be used to identify the minimum ice surface height at the ice divide (Fig. 6).

In the present study area the lengths of the former outlet glaciers are relatively small and therefore lie at the lower end of the range of glacier lengths in the Ng et al. (2010) dataset, where C values are severely affected by the bed topography. As a result, for the majority of outlets, the minimum value of 5.2 $m^{1/2}$ for C^* produced ice profiles that did not reach the height of the col despite geomorphological evidence clearly indicating the presence of ice over some cols. In other areas, where either the valley gradient was low, or the majority of the outlet catchment was on the plateau, the minimum C^* value of 5.2 $m^{1/2}$ produced an unrealistically high minimum surface profile.

The modern analogue dataset used by Ng et al. (2010) was acquired from one of the authors (I.D. Barr, pers. comm., 2011) and is plotted below (Fig. 7) to encompass glacier length, the *C** value and, importantly, glacier relief (H). Appropriate minimum, maximum and 'typical' *C** values were then identified from Figure 7 for individual former outlet glaciers in the study area. To do this, for each outlet glacier to be reconstructed, a range of ± 50 m either side of its altitudinal relief and ± 30% of its length were used to define an envelope of *C** values from which a minimum, maximum and 'typical' *C** value was taken. This method, although admittedly crude, provided a way of selecting minimum and maximum *C** values for each outlet glacier that provided a more realistic constraint on the range of possible ice thicknesses, based on the glacier's length and altitudinal range, than the original suggestion by Ng et al. (2010). However, for some outlets, this still resulted in a wide range of potential ice surface elevations, potentially due to the fact that the location of the modern glaciers used in Figure 7 was not considered and therefore differences in the thermal regime between the modern and reconstructed glaciers were not taken into account here.

The use of the two modelling approaches, *combined with constraints from the geomorphological evidence*, allowed minimum and maximum heights for the plateau ice surface to be estimated. An example of this process is shown in Figure 8. Since the geomorphological evidence indicates a plateau icefield, ice surface altitudes of individual outlet glaciers in neighbouring valleys and across the plateau (i.e. north-south) had to be treated together as the altitude across ice-divides must be the same (Fig. 8B). This meant that the modelled surface profiles could not always be followed exactly (although were followed as closely as possible), even where geomorphological evidence was lacking (Fig. 8C). In a complex plateau ice cap setting this seems to be the only feasible, transparent and reproducible approach.

Thus, in an attempt to acknowledge and quantify varying levels of uncertainty across the reconstruction, the approach described above was used to reconstruct a minimum, maximum and 'average' thickness icefield (Fig. 9). These minimum and maximum reconstructions were then used to provide a quantitative estimate of uncertainty in subsequent ELA and palaeoprecipitation calculations (see Section 4.4). Ice contours were drawn at 50 m intervals to depict ice thickness. The contours were hand drawn and their position estimated using the underlying topography and guided by the surface profiles produced by the Benn and Hulton (2010) model. Following the shape of contours on modern glaciers, the contours were drawn to curve downglacier in the estimated ablation zone and upglacier in the estimated accumulation zone, whilst more or less straight across in the zone around the estimated ELA. The estimated ablation and accumulation zones will have some bearing on the ELAs calculated in Section 4.4, since the curvature of the contours controls the size of each altitudinal envelope used to calculate the ELAs. However, this is a standard procedure for drawing glacier contours in reconstructions (e.g. Gray, 1982; Benn and Ballantyne, 2005; Carr and Coleman, 2007; Ballantyne, 2002a, 2007a, b; Lukas and Bradwell, 2010) and its influence on the calculated ELAs is likely to be minimal. In areas where the minimum, maximum and average reconstructions varied, the contours were also drawn so that the maximum reconstruction contours depicted thicker ice than the average and minimum reconstructions.

The results of this unprecedented integrated approach to palaeoglacier reconstruction indicate that two coalescent plateau icefields, covering a total area of 280km^2, developed over the central part of the Monadhliath and the adjoining western upland area during the Younger Dryas, with all former valley glaciers fed by and connected to ice sourced on the plateau. The most obvious area of uncertainty is in the southeast, concerning whether blockfield-covered summits were indeed nunataks, and, if so, to what extent (Fig. 9). The geomorphological evidence (lateral moraines and sharp lower boundaries of talus slopes and solifluction lobes) (Fig. 2D) prescribes that the glaciers in Gleann Lochain, Gleann Ballach and Gleann Fionndrigh were connected to ice on the plateau. However, the spatial extent of this connection is unknown and could be as limited as that depicted in the minimum reconstruction (Fig. 9B), with implications for the longevity of the connection of these outlet glaciers to plateau ice during retreat.

4.4. Palaeoclimate during the Younger Dryas

The equilibrium line altitude (ELA) is the altitude on a glacier's surface at which net annual accumulation is equal to ablation, thus providing an indication of where glacier mass balance is equal to zero. The ELA is therefore sensitive to variations in precipitation (correlated with accumulation) and melt-season air temperature (correlated with ablation), and hence changes in the ELA provide an important indicator of fluctuations in local to regional climate (Benn and Evans, 2010). The ELA is therefore often used in palaeoclimate reconstructions based on the former dimensions of glaciers at their maximum extent (e.g. Ballantyne, 2002a, 2007a, b; Benn and Ballantyne, 2005; Bakke et al., 2005a, b; Rea and Evans, 2007; Lukas and Bradwell, 2010; Finlayson et al., 2011; Bendle and Glasser, 2012). Calculated palaeo-ELAs are usually referred to as steady-state ELAs (Benn et al., 2005) since an assumption is made that the reconstructed glacier was in equilibrium with climate at the time it was at its maximum position. However, this is notional since it is unlikely that all glaciers reconstructed within a particular area were in equilibrium at their maximum positions due to short-lived advances (Roe, 2011). It is also unlikely that individual outlets reached their maxima simultaneously (Section 5.2; Lukas and Benn, 2006; Lukas and Bradwell, 2010).

Whilst a number of methods have been suggested to calculate the ELAs of former glaciers (see Meierding (1982) for a comprehensive review), the three used predominantly in previous reconstructions of Younger Dryas glaciers in Scotland are the Area Altitude Balance Ratio (AABR; Furbish and Andrews, 1984; Benn and Gemmell, 1997; Osmaston, 2005), the Accumulation Area Ratio (AAR; Porter, 1975; Torsnes et al., 1993; Benn and Lehmkuhl, 2000) and the Area Weighted Mean Altitude (AWMA; Sissons, 1974). The AABR is currently considered the most reliable method for calculating palaeo-ELAs since it takes into account glacier hypsometry and incorporates different accumulation and ablation gradients through the use of a balance ratio (Benn and Ballantyne, 2005; Osmaston, 2005; Rea, 2009). However, a key assumption of this method is that both the accumulation and ablation gradients are linear, a factor that remains unknown for any former glacier reconstructions. The majority of accumulation and ablation gradients of glaciers within a dataset of sixty-six modern glaciers examined by Rea (2009) fulfilled this assumption, however, indicating that the assumption is likely to hold true at least for the majority of reconstructed glaciers. The AABR method can only be considered representative, however, if an appropriate balance ratio is used (Rea, 2009). Previous palaeoglaciological research in Scotland has used AABRs of 1.67, 1.8 and 2.0 following Benn and Gemmell (1997). More recent work by Rea (2009) calculated an average AABR of 1.9 ± 0.81 (± is one standard deviation) for twenty-three mid-latitude maritime (MLM) glaciers examined in a larger study.

The Rea (2009) dataset provides the most extensive range of balance ratios from a range of glacier types, but the majority of these glaciers were in retreat, rather than in steady-state equilibrium as assumed for the maximum extent of reconstructed glaciers (Rea, 2009), which may introduce additional error into the palaeo-ELA calculations. In addition, the wide ranges on the derived AABRs for specific regions identified by Rea (2009) are significant, with the standard deviation of ± 0.81 for MLM glaciers, demonstrating a wide scope of balance ratios of glaciers at similar latitudes. Nonetheless, the Rea (2009) dataset currently provides the largest number of balance ratios for MLM glaciers, thereby providing the most representative range of AABRs for calculating AABR ELAs in the study area. In addition, inclusion of one standard deviation (± 0.81) into the ELA calculations encompasses the previously used balance ratios of 1.67, 1.8 and 2.0. The differences between the ELAs calculated using these balance ratios is actually very small (a few metres; Table 2), particularly in comparison to other uncertainties related to palaeoclimate reconstruction, which are outlined throughout. We therefore suggest that an AABR of 1.9 ± 0.81 covers a credible range of values and is appropriate for use in future studies of Younger Dryas glaciers in Scotland. Considering the Monadhliath Icefield specifically, however, it must be noted that whilst the MLM dataset contains AABR values for five Norwegian plateau icecap outlet glaciers, AABRs for a whole icefield or icecap are not included. Additionally, these outlet glaciers have AABRs between 1.19 and 1.64, substantially lower than the average of 1.9, although within one standard deviation, indicating that ELAs calculated here for individual glaciers using the AABR of 1.9 ± 0.81 should be considered as minima. This is because lower AABR values will result in higher ELA estimations.

An ELA is estimated using this approach for the average-thickness Monadhliath Icefield using 50 m contour interval areas. In the same way, ELAs are also calculated for those individual outlet glaciers which had a range of relief spanning at least four 50 m contour intervals. Ice divides for these glaciers are assumed to be at the current watershed. ELAs are also calculated for the minimum and maximum thickness icefield reconstructions in order to quantify uncertainty associated with the reconstruction, alongside the uncertainties associated with the 1.9 ± 0.81 AABR. These ELAs and associated ranges are presented in Table 2, which for comparison includes ELAs calculated from the previously used AABRs of 1.67, 1.8 and 2.0, AARs of 0.5 and 0.6, and the AWMA methods for the region (e.g. Sissons and Sutherland, 1976; Sissons, 1979b; Ballantyne and Wain-Hobson, 1980; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Finlayson et al., 2011). The ELAs calculated for each glacier using the different methods fall within a range of 10% of one another showing a reasonable level of agreement between each method. Within potential errors, these ELAs are therefore largely indistinguishable.

Using the 1.9 ± 0.81 AABR value, the ELA for the average Monadhliath Icefield is 714 ± 25 m. Also using the average reconstruction, there is a clear rise in ELAs of individual glaciers from west to east across the region, with a range of 560 – 646 m in the west, 649 – 754 m in the central sector and 738 – 816 m in the east (Fig. 10). The highest ELAs are found on the southern side of the plateau and the lowest occur on the northern side, although the difference between the north and south-facing glacier ELAs is less pronounced than differences between those in the east and west. This is consistent with other studies that have identified a strong west-east precipitation gradient (e.g. Benn and Ballantyne, 2005; Lukas and Benn, 2006; Golledge, 2010; Lukas and Bradwell, 2010) indicating that conditions for glacier growth were more favourable on the west of the Monadhliath, due to a dominant eastward movement of moist airmasses from the Atlantic Ocean (cf. Golledge et al., 2008; Palmer et al., 2012). Since the AABR method is considered to be the most reliable, ELAs calculated using an AABR of 1.9 ± 0.81 are used to estimate former precipitation in the area.

In order to uphold a steady-state ELA, a balance is required between annual precipitation and summer air temperatures to maintain mass flux through the ELA and keep the glacier in equilibrium (Benn and Evans, 2010). As precipitation levels decrease, regional ELAs tend to rise because lower temperatures are required to balance accumulation and ablation totals. Examination of the association between mean summer temperature and total annual precipitation at the ELA of 70 glaciers from mid- and high- latitudes allowed Ohmura et al. (1992) to develop an equation to describe this relationship:

$$P_a = 645 + 296T_3 + 9T_3^2 \qquad (3)$$

Where P_a is the annual precipitation (mm a^{-1}) and T_3 is the 3-month mean summer temperature (ºC) at the equilibrium line.

This relationship allows either temperature or precipitation to be calculated if an independent value is known for the other. It has been used extensively to derive palaeoprecipitation at the ELA of former glaciers in combination with an independent temperature proxy (Kerschner et al., 2000; Carr, 2001; Bendle and Glasser, 2012), usually derived as mean July temperatures inferred from subfossil chironomids assemblages, for Scottish Younger Dryas glaciers (e.g. Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Finlayson et al., 2011). Its suitability for palaeoclimatic reconstruction has been debated, however, due to differences in incoming radiation and factors such as aspect, wind direction and topography that may affect the precipitation-temperature relationship at a local scale (Dahl and Nesje, 1996; Dahl et al., 1997; Kaser and Osmaston, 2002; Benn et al., 2005; Evans, 2006; Braithwaite, 2008; Golledge et al., 2010). Others argue that this 'smoothing-out' of local variations is advantageous, since it leaves the dataset more reliable compared to attempting to account for local variability in areas where no modern glacier data exists (Benn and Ballantyne, 2005;

Lukas and Bradwell, 2010). Despite potential shortcomings, and in the absence of a more suitable equation, the Ohmura et al. (1992) relationship has formed the basis for the majority of recent palaeoclimatic reconstructions based on palaeo-ELAs in Scotland.

There is, however, a discrepancy in the amount of precipitation suggested by palaeoglaciological reconstructions that have used this relationship (e.g. Benn and Ballantyne, 2005; Lukas and Bradwell, 2010) and general circulation models (GCMs) (e.g. Björck et al., 2002; Jost et al., 2005), where the GCMs suggest a much drier climate. Golledge et al. (2010) suggest this inconsistency could be explained by an increase in continentality, caused by extensive sea-ice development in winter (Isarin et al., 1998; Isarin and Rensen, 1999), which may cause the Younger Dryas climate in Scotland and resultant glacier dimensions to fall outside of or near the margins of validity of the Ohmura relationship. This is supported by biological proxies and periglacial evidence that indicates that the annual temperature range in Scotland during the Younger Dryas was about 30°C (Atkinson et al., 1987; Ballantyne and Harris, 1994). As an acknowledgment of this complexity and arguments made above, Golledge et al. (2010) advocate the use of a new function specifically designed for Younger Dryas palaeoprecipitation calculations in Scotland. Golledge et al. (2010) argue that this new precipitation-temperature function (eqn. 4), which is based upon a model of the Scottish Younger Dryas ice cap (Golledge et al., 2008), and incorporates Younger Dryas climate estimates from the Greenland ice core record and regional biological proxy data (annual temperature range, 30°C), will provide a more realistic estimation of Younger Dryas palaeoprecipitation in Scotland and recommend its adoption into such studies.

$$P = S(14.2T_3^2 + 248.2T_3 + 213.5) \qquad (4)$$

Where P is effective precipitation, S is the seasonality constant; S = 1 for neutral type, S = 1.4 for summer-dominated and S = 0.8 for winter-dominated precipitation seasonality, T_3 is the mean 3-month summer temperature (°C) at the ELA.

The seasonality constant (S) allows the equation to be altered to account for neutral-, winter- and summer-type precipitation. Golledge et al. (2010) argue that the time of year at which the majority of precipitation falls will alter the amount that is required to maintain a glacier in steady state for a particular ELA. For example, in summer a large proportion of precipitation will either be lost to ablation or will fall as rain, enhancing ablation by advective heat transfer, and therefore larger annual quantities are required to maintain the same ELA. Therefore Golledge et al. (2010) recommend use of the term *effective* precipitation.

Palaeoprecipitation for the Monadhliath is calculated using both the Ohmura et al. (1992) and the Golledge et al. (2010) equations, enabling comparison with previous studies and comparison between the two methods. Two values for summer temperature at sea level of 8.5 ± 0.3° C (mean July temperature) and 6.38° C are used, in accordance with other recent work (e.g. Benn and Ballantyne, 2005; Finlayson, 2006; Lukas and Bradwell, 2010; Finlayson et al., 2011). The 8.5 ± 0.3° C value is based on chironomid data from Whitrig Bog (125 m OD) and Abernethy Forest (220 m OD), which lie approximately 260 km southeast and 30-90 km east of the Monadhliath respectively (Brooks and Birks, 2000, 2001; Brooks et al., 2012). Further details can be found in Benn and Ballantyne (2005).

The second value of 6.38° C is derived from modelling experiments by Golledge et al. (2008). It is advocated by Golledge (2008) as an alternative value that takes into account the effects of localised cooling of air temperatures by glaciers (Khodakov, 1975; Braithwaite, 1980; Singh et al., 2000; Hughes and Braithwaite, 2008), unlike chironomid-derived temperatures that are derived from ice-free areas, and has been used to calculate palaeoprecipitation by Finlayson et al. (2011) for the Younger Dryas icecap on Beinn Dearg in northern Scotland.

Use of either of these temperatures, derived from the coldest part of the Younger Dryas, is potentially at odds with a complex build-up history of Scottish Younger Dryas ice masses, which is likely to have started in more temperate conditions prior to the Younger Dryas (cf. Lukas and Bradwell, 2010). This means that precipitation calculated using this temperature value could potentially underestimate precipitation totals during glacier build-up, adding further uncertainty to these values.

In order to transform the mean July summer temperature of 8.5 ± 0.3° C into a mean summer temperature, as required by both equations, eqn. (5) was used, as advocated by Benn and Ballantyne (2005) from analysis of meteorological data from Scotland and Scandinavia. This equation assumes that the current Scandinavian and Scottish summer climates are good analogues for summer climate in Scotland during the Younger Dryas. However, in the absence of any other data, the equation is used here and considered to be a reasonable approximation.

$$T_3 = 0.97T_J \qquad (5)$$

where T_3 is the mean summer temperature and T_J is the mean July temperature (°C) at the ELA.

Environmental lapse rates are highly variable and a wide range has been documented in modern glacial and periglacial environments (e.g. Jonsell et al., 2013; Pike et al., 2013), thus adding significant uncertainty to this stage of the palaeoprecipitation calculation. To keep our calculations consistent with previous palaeoglacier reconstructions in Scotland, we use environmental lapse rates of 0.006-0.007 °C m^{-1} to derive T_3 at the ELA for both temperature values.

An average lapse rate of 0.0065 °C m^{-1} is used for the average-thickness Monadhliath Icefield ELAs, and palaeoprecipitation values for the Monadhliath Icefield and its outlet glaciers are displayed in Table 3 following the methods described above. The lower (0.006 °C m^{-1}) and upper (0.007 °C m^{-1}) lapse rates, are incorporated into the uncertainty at a later stage using the minimum and maximum-thickness icefield reconstructions, and a standard error of ± 200 mm also added to incorporate variations in the relationship between air temperature and ablation (Ohmura et al., 1992). The lower boundary of uncertainty (minimum precipitation) is calculated using a 1.09 AABR ELA (lower error bracket of the AABR 1.9 ± 0.81) for the maximum icefield reconstruction, since this produces the highest ELA. The lower margin (8.2 °C) of the mean July sea-level temperature of 8.5 ± 0.3 °C is then used alongside the lowest adiabatic lapse rate (0.006 °C m^{-1}). Conversely, the upper boundary of uncertainty (maximum precipitation) is calculated using a 2.71 AABR ELA from the upper boundary of the 1.9 ± 0.81 balance ratio from the minimum icefield reconstruction, using a mean July sea-level temperature of 8.8 °C and an adiabatic lapse rate of 0.007 °C m^{-1}. Of note is the use of the maximum thickness icefield reconstruction to calculate the minimum precipitation values and vice versa due to the effect that a thicker or thinner plateau icefield has on raising or lowering the ELA, and that using a lower lapse rate with a the lower July sea-level temperature produces a higher July temperature at the ELA. However, these combinations were selected with the sole purpose of producing the largest range of potential precipitation values, and thus acknowledge all uncertainties regarding balance ratios, July temperatures and environmental lapse rates.

Following the method outlined above and incorporating these uncertainties, average palaeoprecipitation of the Monadhliath Icefield is calculated at 1829 ± 491 mm a^{-1} at the 714 m ELA using the Ohmura relationship and a mean July temperature at sea level of 8.5 ± 0.3° C. Using this temperature, values for summer-dominated, neutral and winter-dominated precipitation with an annual temperature range of 30° C are 1809 ± 592 mm a^{-1}, 1292 ± 480 mm a^{-1} and 1034 ± 424 mm a^{-1} at the ELA respectively. Precipitation values for the icefield calculated using the summer temperature of 6.38° C are significantly lower (Table 3). Palaeoprecipitation is also calculated for major outlet glaciers using the mean July temperature of 8.5 ± 0.3° C and shows a decrease in precipitation from west to east, reflecting the rise in ELAs.

In order to compare the precipitation values with other studies the equivalent sea-level precipitation totals are calculated. Since precipitation increases non-linearly with altitude, Ballantyne (2002a) devised a relationship to calculate the corresponding precipitation at different altitudes using:

$$P_{Z1} = P_{Z2}/(1+P^*)^{0.01(Z2-Z1)} \quad (6)$$

Where P_{Z1} and P_{Z2} are the amounts of precipitation (mm a^{-1}) at sea level and the ELA respectively. P^* is the proportional increase in precipitation per 100 m increase in elevation. Based on a dataset for Ben Nevis, which is approximately 30 km southwest of the study area, Ballantyne (2002) shows that P^* = 0.0578 and this value is used here.

The sea-level equivalent precipitation for the Monadhliath Icefield is found to be 1224 ± 409 mm a^{-1} using the Ohmura et al. (1992) equation, and 1211 ± 480 mm a^{-1}, 865 ± 400 mm a^{-1}, and 692 ± 360 mm a^{-1}, for the summer-, neutral- and winter-dominated precipitation types respectively using the Golledge et al. (2010) function (Table 3).

4.5. The cooling effect of glaciers on local air temperatures

Unlike the value of 6.38 °C derived from modelling experiments (Golledge, 2008), the chironomid-derived July palaeotemperature of 8.5 ± 0.3 °C does not account for the effect of air temperature cooling by glaciers. Consequently, the 2°C difference in these values results in a stark contrast between the palaeoprecipitation values calculated for each temperature (Table 3). This is investigated further for a selected number of outlet glaciers and indicates that use of 6.38 °C results in a reduction in precipitation values of up to 50% compared to those calculated using the 8.5 ± 0.3 °C July sea-level temperature inferred from chironomid assemblages, dependent on ELA (Table 4).

According to Khodakov (1975) glaciers may reduce the surrounding air temperature by between 1.6 and 1.9°C for a glacier of 10-20 km in length (eqn. (7)).

$$\log\Delta T = 0.28\log L - 0.07 \qquad (7)$$

Where ΔT is the change in temperature (°C) and L is the length the glacier (km).

Using eqn. (7) to reduce the temperature at the ELA for the 8.5 ± 0.3 °C July sea-level temperature, precipitation values are recalculated for a selected number of outlet glaciers (Table 4). The effect of glacier cooling on air temperature cannot be considered for the icefield as a whole or more complex outlet glaciers, however, since no single value can be obtained for glacier length.

Since the majority of outlet glaciers in the Monadhliath are less than 5 km in length, the reduction in temperature calculated using eqn. (7) is lower than the 2°C used by Finlayson et al. (2011) and results in precipitation reductions of between 17% and 33% for the Ohmura et al. (1992) relationship, and between 23% and 46% for the Golledge et al. (2010) equation (Table 4), dependent on glacier length and ELA. It is clear from these calculations that, whilst the cooling effect that glaciers have on local air temperatures is significant, there is considerable uncertainty surrounding the amount that temperature at the ELA of individual glaciers will be lowered, which in turn can notably alter the amount of precipitation calculated. We suggest that the precipitation totals presented in Table 3 for the 8.5 ± 0.3 °C July sea-level temperature should be considered first to allow comparison with previous studies, but that these should be regarded as maximum values to account for any effects of glacier cooling.

5. Discussion

5.1. Icefield dimensions and thermal regime

The Younger Dryas plateau icefield reconstructed in this study covers an area of c. 280 km^2 and is of similar proportions to that generated through proxy-climate-based numerical modelling by Golledge et al. (2008) (Fig. 11). The model is of insufficient resolution to reproduce ice flowing into any of the major outlet valleys, remaining on the plateau only, but it provides a remarkably close match in terms of eastwards extent of plateau ice to that based on the empirical field evidence presented here. In the west, the model is unable to reproduce the established limits of the West Highlands Ice Cap, since it cannot replicate ice-dammed lake formation in Glen Spean, Glen Roy and Glen Gloy and therefore is less comparable with this Monadhliath Icefield reconstruction. Golledge et al. (2008) suggested that the

Monadhliath Icefield remained separate from the main West Highlands Ice Cap because of these lakes, and this previously-suggested ice configuration conforms well with the empirical field evidence presented here.

In terms of glacier thermal regime, the geomorphological evidence suggests that the Monadhliath Icefield was polythermal. This is linked to the earlier discussion on blockfields (Section 4.3), where coverage of blockfields by ice, as in this reconstruction, suggests that these areas were cold-based (cf. Fabel et al., 2012), whilst moraines in the outlet valleys indicate warm-based ice. This assertion of cold-based plateau ice is also evidenced by a large quantity of ice-marginal meltwater channels that occur on the plateau, including those at the plateau edge above Gleann Ballach, which are indicative of a cold-based landsystem (cf. Dyke, 1993; Rea and Evans, 2003). Cold-based Younger Dryas plateau ice also allows preservation of other, older landforms, and it is possible that some of the large isolated moraines on the plateau may relate to an earlier phase of glaciation.

In other parts of the plateau, smaller, closely-spaced moraines occur (Boston, 2012a, b) and are suggested to be of Younger Dryas age due to their morphology, close spacing and position in the source area of reconstructed Younger Dryas outlet glaciers. These moraines appear to indicate that, during the final stages of Younger Dryas deglaciation, plateau ice actively retreated into topographic lows on the plateau. This occurred mainly between Glen Markie and the Eskin and Abhainn Crò Chlach catchments in the eastern sector of the Monadhliath, but also further west above Coire Laogh, Glen Brein, Coire Easgainn and Corrie Yairack.

From examination of the geomorphology within the outlet valleys, it is apparent that the density of moraines varies significantly between them. Of the 53 outlet valleys identified, only seventeen have well-preserved, closely-spaced recessional moraines. Glaciers in the remaining 36 are reconstructed based on the presence of sporadic moraines, ice-marginal meltwater channels and evidence from neighbouring valleys that suggests that ice must have been present. Of these valleys, 13 are dominated by lateral meltwater channels, with some moraines, whilst the remainder contain only lateral meltwater channels. While not every valley in other similarly-sized areas in Scotland contains moraines, the number of valleys without clear constructional glaciogenic landforms in the Monadhliath is certainly distinctive (cf. Lukas and Bradwell, 2010; Finlayson et al., 2011). We admit that due to the lack of geomorphological evidence in these valleys, we are less confident of the presence or extent of outlet glaciers here, however, based on reconstructed ice thicknesses on the plateau and in neighbouring areas, it seems unlikely that ice was altogether absent from these areas.

Reasons for the limited number of moraines are numerous and include: 1) low debris turnover due to a lack of debris entrainment, possibly coupled with a thin snout, which could lead to a cold-based thermal regime (Ó Cofaigh et al., 2003); 2) a lack of sediment readily available for re-entrainment (cf. Ballantyne, 2002b, c), although, in most of the outlet valleys, stream exposures reveal thick diamictic deposits in the valley bottoms, indicating a plentiful supply of debris; 3) some outlet glaciers may have experienced uninterrupted rather oscillatory retreat, which may be linked to 1) above; 4) moraines could have been deposited but not preserved, either due to a) a well-coupled glacial and fluvial system that allowed proglacial streams to immediately remove any sediment that was deposited at the ice margin or b) the presence and subsequent meltout of buried ice (Lukas, 2007; Evans, 2009; Brook and Paine, 2012), and 5) burial of small, moraines by a) peat or b) subsequent hillslope processes (e.g. Müller et al., 1983) (particularly relevant for lateral moraines), 6) differences in bed topography (e.g. slope steepness) (Barr and Lovell, 2014).

Of these factors low debris turnover (1) and/or low moraine preservation due to the presence of buried ice (4b), continuous retreat (3) and burial by hillslope processes and peat formation (5) are the most likely reasons for the limited number of observable moraines in these outlet valleys. Both factors 1 and 4b are indicative of cold-based to polythermal conditions at the glacier bed (Ó Cofaigh et al., 2003). This is plausible given the low gradient of these valleys, which often descend gently from the plateau, with no backwall, meaning that little strain heating would have occurred as the glacier flowed from the plateau, in comparison to glaciers that flowed into the larger valleys (e.g. Evans, 2010). The majority of these glaciers are also reconstructed as thinner than the major warm-based glaciers, which would again support the notion that these glaciers did not reach pressure melting point (Rea and Evans, 2003) and that permafrost could penetrate underneath thinner snouts (cf. Björnsson et al., 1996). On this basis, we therefore suggest that the resulting geomorphology within the outlet valleys most likely lies on a process-form continuum between cold-based and warm-based thermal regimes (cf. Evans, 2010), resulting in a landscape dominated by meltwater channels and moraines to a varying extent. This mosaic of thermal regimes has previously been recognised beneath both plateau icefields and ice sheets by numerous authors including Sugden (1968), Dyke (1993), Rea et al. (1998), Evans et al. (2002), Hall and Glasser (2003), Kleman et al. (2008) and Evans (2010).

5.2. Timing of maximum glaciation

The timing of the maximum extent of Younger Dryas glaciation in Scotland has been subject to significant discussion recently, and opinions diverge from a late stadial maximum (Palmer et al., 2010, 2012; MacLeod et al., 2011) to the more traditional mid-stadial maximum (cf. Benn et al., 1992; Benn and Ballantyne, 2005; Golledge et al., 2008; Lukas and Bradwell, 2010; Ballantyne, 2012), whilst recent controversial work might indicate early deglaciation of Rannoch Moor, beneath the central zone of the West Highlands Icefield, at 12.2 ka BP (Bromley et al., 2014). In this context we merely note that the role of Rannoch Moor as a centre of ice dispersal is not entirely clear and that there may be issues with the lack of context provided by these authors. The stratigraphical relationship of the western part of the Monadhliath Icefield with the Glen Roy ice-dammed lakes suggests that outlet glaciers in this area reached their maxima following lake drainage, which Fabel et al. (2010) and Palmer et al. (2010, 2012) suggest occurred towards the end of the stadial. This could have been caused by a change from a calving to terrestrial ice margin rather than a climatic signal, however (e.g. Reitner, 2007). Conversely, a late-stadial advance in the east would be at odds with the SED ages obtained by Gheorghiu et al. (2012) which on the whole suggest the outlet glaciers reached their maxima in the early to mid-stadial, after recalculation using a local production rate (Gheorghiu and Fabel, 2013) (12.0 ± 0.9 and 13.0 ± 0.7 ka BP in Gleann Lochain, and 12.6 ± 0.7 and 13.3 ±0.7 ka BP in Gleann Ballach). However, differences between dates on the same moraine and the uncertainties of 700 to 900 years associated with the SED method make this difficult to assess.

There is a lack of significant end moraines in many valleys, where the largest interpreted Younger Dryas moraines are often located behind the outermost moraines (e.g. Coire Easgainn, the Findhorn Valley, Gleann Ballach, Corrie Yairack, Coire Larach). Some of the mapped river terraces also start just within the outermost moraines (e.g. Glen Shesgnan, the Dulnain Valley). These morphostratigraphic relationships have been observed elsewhere in the Scottish Highlands (e.g. Sissons, 1974) and have been suggested to indicate a two-phased Younger Dryas advance (see also Peacock et al., 1989; Merritt et al., 2003), whereby the outlet glaciers are suggested to have only briefly remained at their maximum positions, prior to slight retreat and stabilisation for longer further upvalley. Such a hypothesis of two-phase Younger Dryas glaciation may be tested in future as higher-resolution numerical dating methods are refined or become available.

5.3. Regional Palaeoclimatic Inferences

The decrease in precipitation from west to east across the Monadhliath indicates a steep precipitation gradient, with reductions in precipitation of up to 43% in the east compared to the west, based on sea-level equivalents. Table 5 compares the palaeoprecipitation at sea level calculated for the Monadhliath Icefield with sea-level equivalent palaeoprecipitation values for published ELAs of other Younger Dryas ice masses in Scotland, which are recalculated using identical methods to those described in Section 4.4 to enable comparison. The marked contrast between precipitation totals in the west and east of the Monadhliath closely fits to those calculated for neighbouring areas; values calculated for the western sector are very similar to those calculated for Creag Meagaidh and Drumochter, showing strong consistency across studies in the western part of the field area (Benn and Ballantyne, 2005; Finlayson, 2006), whilst precipitation in the eastern sector of Monadhliath is similar to that of the Gaick Plateau and southeast Grampians (Sissons, 1974; Sissons and Sutherland, 1976). In this respect, precipitation calculated for the Cairngorms seems anomalously low (Sissons, 1979c) and a re-evaluation of some of this older work using newer methods for mapping, glacier reconstruction and ELA calculation would be helpful for better assessing regional precipitation patterns, as has been demonstrated elsewhere (i.e. Lukas and Bradwell, 2010).

5.4. The effect of seasonality on precipitation at the ELA

The assertion made by Golledge et al. (2010) that the Younger Dryas climate in Scotland was more arid than currently recognised using the Ohmura et al. (1992) relationship is highly dependent on the degree of seasonal precipitation bias, and the reduction of air temperature caused by each glacier as discussed above. As shown in Tables 3 and 5 a summer-type precipitation for the function produces a similar level of precipitation as the Ohmura relationship, whilst neutral- or winter-type precipitation would require significantly less precipitation to maintain the same ELA. Recent research indicates that the development of winter sea ice in the North Atlantic during at least the first half of the Younger Dryas (Bakke et al., 2009) diverted storm tracks south of the sea-ice margin and towards continental Europe, generating a more stable, arid environment in Scotland and Norway (Isarin et al., 1998; Isarin and Rensen, 1999; Bakke et al., 2009; Golledge et al., 2010; Palmer et al., 2012). By contrast, increasing temperatures in the latter part of the stadial are thought to have caused the break-up of this sea ice, allowing storm tracks to move northwards across Scotland and northern Norway, the frequency of which was controlled by the fluctuating sea-ice margin (Bakke et al., 2009). These changes in sea-ice extent suggest that during the early part of the stadial winter precipitation was likely to be suppressed due to extensive sea ice and also possibly reduced during the summer months. In the latter part of the stadial, however, the increase in the number of storm tracks passing over Scotland would have led to increased precipitation in potentially both the summer and winter months (Palmer et al., 2012). This is contrary to the pattern suggested previously by Benn et al. (1992).

The timing of maximum glaciation (Section 5.2) is therefore critical in evaluating the seasonality bias that may have occurred. A mid-stadial maximum would indicate that the majority of precipitation arriving on Scottish glaciers occurred during the summer months and therefore a summer-precipitation bias, with totals similar to those derived using the Ohmura relationship, may be most appropriate. On the other hand, a late-stadial maximum resulting from increased precipitation during both the winter and summer months, suggests a neutral precipitation bias, although a summer bias during glacier build-up would still need to be accounted for.

It is evident that there is a large level of uncertainty surrounding identification of an appropriate degree of seasonality bias to include in the Golledge et al. (2010) equation. The function is clearly useful in drawing attention to and beginning to understand the complexities surrounding how palaeoprecipitation might be calculated using glaciers as a proxy. However, further constraint on 1) the timing of maximum glaciation, 2) the degree of seasonal precipitation bias, and 3) the amount of localised air temperature cooling at the glacier surface is required before confident estimates of palaeoprecipitation in Scotland can be made. As a first approximation, however, we suggest that by assuming an initial mid-stadial maximum (summer-type precipitation), and taking into consideration the effects of glacier air temperature cooling not accounted for in Tables 3 and 5, values derived from a neutral-type precipitation may be most appropriate for the Monadhliath Icefield. We reaffirm, however, that despite uncertainties regarding absolute precipitation values, we have confidence in the trend suggested by the data, which indicates an eastwards decline in precipitation, in line with previous work (e.g. Sisson, 1979b).

5.5. Comparison to modern precipitation totals

Within the context of the previous discussion, we finally attempt to make a comparison to modern precipitation within the Monadhliath. We reiterate that with the uncertainties associated with seasonal precipitation bias and the effect of glacier air temperature cooling, the use of the neutral-type precipitation values in Tables 3 and 5 is a first approximation to enable some comparison to be made, and all three precipitation types are shown in Table 6. Average precipitation for ten years (2000 to 2009 inclusive) was acquired for three sites in the

Monadhliath (Table 6). The modern data also shows a significant decrease in precipitation across the area, varying from 2112 mma^{-1} at Braeroy Lodge (220 m OD; NN 336 914) in the west, to 1317 mm a^{-1} and 1325 mm a^{-1} further east at the Spey Dam (270 m OD; NN 582 973) and Coignafearn Lodge (390 m OD; NH 710 179) respectively. These data therefore denote a steeper present-day precipitation gradient than indicated by the reconstructed Younger Dryas precipitation estimations presented (Tables 3 and 6). As a result, the greatest difference occurs in the west, where modern precipitation is 800 mm a^{-1} higher than the neutral-type precipitation calculated for the Younger Dryas. Overall, the comparison indicates that if a neutral-type precipitation is assumed, then precipitation across the whole of the Monadhliath was significantly reduced compared to present. Indeed, even if a summer-type precipitation bias or the Ohmura relationship is used, the results still indicate reduced precipitation in this central part of Scotland during the Younger Dryas compared to the present day.

6. Conclusions

The main findings of this research are summarised as follows:

- The first systematic geomorphological mapping of the Monadhliath reveals evidence for two phases of plateau icefield glaciation following deglaciation of the last British-Irish Ice Sheet.
- We argue, using morphostratigraphical principles, that the most recent phase of plateau icefield glaciation occurred during the Younger Dryas (Loch Lomond Stadial, GS-1, 12.9-11.7 ka). This is supported by surface exposure dates in the southeast of the region (Gheorghiu et al., 2012), which indicate the presence of ice in two outlet valleys during the Younger Dryas.
- The Younger Dryas icefield is reconstructed using a combination of geomorphological evidence and two 2D glacier surface profile models (Benn and Hulton, 2010; Ng et al.,

2010). This indicates that at this time two coalescent icefields were present over the central Monadhliath plateau and adjoining western upland to the north of Glen Roy, covering an area of approximately 280 km^2.

- The greatest uncertainty in the reconstruction is associated with ice thickness on the plateau due to a lack of geomorphological evidence. Therefore minimum and maximum thickness icefields are also produced to attempt to quantify this uncertainty.
- ELAs are calculated for the icefield and all major outlet valleys, using the AABR, AAR and AMWA methods. Using an AABR of 1.9 ± 0.81, the ELA of the Monadhliath Icefield is calculated to be 714 ± 25 m, which takes into account uncertainty quantified using minimum and maximum thickness reconstructions. ELAs calculated for individual outlet glaciers show a clear rise in ELAs from west to east across the region, with a range of 560-646 m in the west, 649-754 m in the central sector and 738-816 m in the east.
- Assuming a mean July sea-level temperature of 8.5 ± 0.3 °C, palaeoprecipitation at sea level is estimated to have been 1224 ± 409 mma^{-1} using the Ohmura et al. (1992) equation and 1211 ± 480 mma^{-1}, 865 ± 400 mma^{-1}, 692 ± 360 mma^{-1}, using the Golledge et al. (2010) equation with summer-, neutral- and winter-type precipitation seasonality respectively. Whilst further work is required to establish the degree of seasonality that may have occurred during the Younger Dryas, we suggest here that the value derived from the neutral-type precipitation may be most appropriate in order to take into account the cooling effect of glaciers on air temperature. Compared with modern precipitation data, these values indicate lower precipitation during the Younger Dryas than at present in the Monadhliath.
- Comparison with other studies in Scotland shows that the figures calculated here fit well with the regional pattern of Younger Dryas ELAs, indicating an eastwards

decline in precipitation across Scotland. ELAs in the western sector of the Monadhliath (average 610 m) are comparable to those calculated for the nearby Drumochter Hills (626 m; Benn and Ballantyne, 2005) and Creag Meagaidh (625 m; Finlayson, 2006), whilst those for the eastern sector (average 777 m) are comparable to the Gaick (787 m; Sissons, 1974) and the southeast Grampians (790 m; Sissons and Sutherland, 1976).

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Figures

Figure 1. Topographic map to show the location of the Monadhliath in Scotland, and all valleys referred to in Table 1 and the text. UK Outline from Ordnance Survey © Crown copyright 2010. NEXTMap DSM hillshade model from Intermap Technologies (2007).

Figure 2. Extracts from the glacial geomorphological map presented by Boston (2012a). A) The Findhorn Valley in the northeast of the study area (NH 661 135); B) Corrie Yairack to

the southwest (NN 443 970); C) Corrie Easgainn, a tributary to Glen Killin in the central north (NH 523 078); D) Gleann Lochain (NH 632 006), Gleann Ballach (NH 652 013) and Gleann Fionndrigh (NH 665 017), tributaries of Glen Banchor in the southeast.

Figure 3. Glacial geomorphological map of Glen Turret and the upland area to the northeast. Adapted from Boston (2012a).

Figure 4. Glacial geomorphological map, adapted from Boston (2012a), of the four Glen Banchor tributary valleys that were examined by Gheorghiu et al. (2012). This map includes a comparison of the Younger Dryas limits identified in this paper (red) and by Gheorghiu et al. (2012) (yellow). Letters correspond to those used by Gheorghiu et al. (2012) and additional locations discussed in the text.

Figure 5. Example of the use of the Benn and Hulton (2010) model to extrapolate the surface profile of a glacier (Easgainn) from an area (usually the lower part) that is well constrained by geomorphological evidence to an area (often plateau) for which there is no evidence to constrain ice thickness. Both maximum and minimum thickness models are initially guided by the geomorphological evidence, in this case lateral moraines, which required a maximum yield stress of 250 kPa to replicate the steep surface profile of this particular glacier as it flowed from the plateau. Most modelled glaciers in this study required shear stresses below 200 kPa to match the geomorphological evidence, however. The shear stress was then reduced towards zero near the ice divide on the plateau, as recommended by Benn and Hulton (2010).

Figure 6. Mathematical equations used to define glacier surface profiles for a) the parabolic

surface of a glacier as defined by Nye (1951, 1952) (eqn. 1). b) Diagram to illustrate the use of $C*$ and $\tilde{C}$ to best fit the parabola to the surface profiles sampled by Ng *et al*. (2010). From Ng et al. (2010, p. 3242).

Figure 7. Scatter plot showing glacier relief (*H*) against *C**, where the size of the data points corresponds to the length of the glacier. Data from Ng et al. (2010).

Figure 8. Illustration of the combined approach used to identify maximum and minimum boundaries of ice thickness on the plateau for the Monadhliath Icefield: A) numbered glacier centre lines used in this example, which correspond to the table in C; B) conceptual diagram of the process by which modelled ice thicknesses were combined with geomorphological evidence and evidence in neighbouring valleys. In the example, the south facing glacier must be connected to the plateau due to modelled and geomorphological evidence in the neighbouring north-facing valley for a reasonable thickness of ice at the ice divide. Likewise the northern glacier cannot be as thick as modelled, due to constraints on ice thickness of the southern glacier; C) table presenting maximum, minimum and average values for ice thickness of major outlet glaciers in the central sector of the Monadhliath, based on glacier surface profile modelling (Benn and Hulton, 2010; Ng et al., 2010) and any geomorphological/topographic constraints as described in the notes section below. The units for C* are $m^{1/2}$ and No MA = no modern analogue.

Figure 9. Average (A), minimum (B) and maximum (C) reconstructions of the Younger Dryas Monadhliath Icefield. Numbered glacier outlets correspond to those in Table 2. NEXTMap DSM hillshade model from Intermap Technologies (2007).

Figure 10. Major outlet glaciers of the Monadhliath Icefield classed according to their ELA in groups of 50 m intervals, showing an increase from the western to central to eastern sectors. NEXTMap DSM hillshade model from Intermap Technologies (2007).

Figure 11. Comparison of the reconstructed Monadhliath Icefield (orange) in comparison to modelled Younger Dryas ice extent in the Monadhliath from an 'optimum fit' 500 m resolution three-dimensional thermomechanical ice-sheet model that was constrained by field evidence in other areas of Scotland. Adapted from Golledge et al. (2008).

Table 1. Criteria used to define the maximum limits of Younger Dryas outlet glaciers in the Monadhliath. The location of each valley is shown in Figure 1.

Table 2. Equilibrium line altitudes for the Monadhliath Icefield and major outlet glaciers calculated using the AABR, AAR and AMWA approaches. Uncertainty is calculated using the maximum and minimum icefield reconstructions, and for the ELAs calculated using the AABR of 1.9, the range of ±0.81 (Rea, 2009) is also included in the uncertainty. Glaciers are organised by area of the Monadhliath and from west to east within these groups. The dominant direction of flow for each glacier is given in brackets and the number corresponds to their location on Fig. 9.

Table 3. Palaeoprecipitation values and uncertainty for the Monadhliath Icefield and major outlet glaciers at their respective ELAs and at sea level, calculated using the AABR = 1.9 ± 0.81 ELA. The results of both the Ohmura et al. (1992) and Gollege et al. (2010) precipitation-temperature relationships are displayed, where S-type = summer type

precipitation, N-type = neutral type precipitation and W-type = winter type precipitation. Palaeoprecipitation values for the Monadhliath Icefield are presented for both a mean July temperature at sea level of 8.5 ± 0.3° C (Benn & Ballantyne, 2005) and a summer sea-level temperature of 6.4° C (Golledge, 2008). A mean July temperature at sea level of 8.5 ± 0.3° C only was used to calculate the palaeoprecipitation values for the major outlet glaciers. These glaciers are organised by area of the Monadhliath and from west to east within these groups.

Table 4. The percentage change from the precipitation values presented in Table 3 after including the effect of cooling by glaciers on the temperature at the ELA.

Table 5. Annual precipitation values for Younger Dryas sites in Scotland, arranged from west to east, based on published ELAs, but calculated using the methods described here. The table includes the Monadhliath Icefield and average values for the western, central and eastern sectors.

Table 6. A) Modern precipitation data averaged over a 10 year period between 2000 and 2009 for three Met Office weather stations in the western (Braeroy), central (Spey Dam) and eastern (Coignafearn) sectors of the Monadhliath. Uncertainty is one standard deviation. B) Average Younger Dryas precipitation, calculated using the Golledge et al. (2010) precipitation-temperature function, for the western, central and eastern sectors of the Monadhliath Icefield, with uncertainty calculated from the maximum and minimum icefield reconstructions, as described in the text. Modern precipitation data provided by the British Atmospheric Data Centre (NERC) and verified using the FetchClimate (2012) web application.