A spatially-restricted Younger Dryas plateau icefield in the Gaick, Scotland: Reconstruction and palaeoclimatic implications

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Abstract

Considerable research has been conducted in Scotland to reconstruct Younger Dryas glaciers and palaeoclimatic conditions, but our understanding remains incomplete. In this contribution, we examine the Gaick, a dissected plateau that extends over ~520 km^2 in the Central Grampians, Scotland. The extent and style of Younger Dryas glaciation in the Gaick has been repeatedly contested, although a model of extensive plateau icefield glaciation has become generally accepted. This is despite well-documented issues with key elements of the plateau icefield reconstruction. We synthesise the results of recent geomorphological mapping in the Gaick and recognise a distinct morphostratigraphic signature in the upper parts of the western catchments. This differs markedly from sediment-landform associations in other parts of the area, and we argue this provides a strong indication of spatially-restricted Younger Dryas (~12.9–11.7 ka) glaciation in the Gaick. Our interpretation is independently supported by glacierisation threshold analysis, which implies that the eastern Gaick was unable to nourish Younger Dryas ice. We therefore contest the accepted paradigm of extensive Younger Dryas glaciation in this area. Based on the geomorphological evidence and glacier surface profile modelling, we reconstruct a ~42 km^2 plateau icefield that yields an equilibrium line altitude of 751 ± 46 m. Using this value, a sea-level precipitation value of 826 ± 331 mm a^{-1} is inferred for the Younger Dryas, which suggests considerably drier conditions than at present. Using recalculated glacier-derived precipitation estimates from Scotland, we present regional climate analysis that corroborates arguments for a strong west-east precipitation gradient across Scotland.

Keywords: plateau icefield; Younger Dryas; glacier reconstruction; palaeoclimate; Scotland

1. Introduction

The Younger Dryas Stadial, which broadly correlates to Greenland Stadial-1 (GS-1; ~12.9–11.7 ka; Lowe et al., 2008), was a period of rapid and high magnitude climatic change (e.g. Bakke et al., 2009) that potentially offers a valuable analogue for future climate change (cf. Lukas, 2011). Thus, elucidating climatic events during this period, and their impact on Earth systems (e.g. the cryosphere), is of high importance. Indeed, empirical studies of Earth system responses to climate change during the Younger Dryas are required to provide crucial palaeoenvironmental data for testing models used to forecast future climate change (e.g. Braconnot et al., 2012; Vaughan et al., 2013).

In Scotland, the Loch Lomond Stadial is broadly equated with the Younger Dryas, and it is manifest in the geological record as an abundance of well-preserved glacial sediment-landform assemblages (see Benn and Lukas, 2006; Golledge, 2010; Bickerdike et al., 2016, 2018a, b). This important glacial geological archive is unique in the context of the amphi-North Atlantic region in that it (a) has experienced very limited post-depositional modification and (b) relates to discrete terrestrial glaciers (i.e. cirque glaciers, valley glaciers and icefields) that can be reconstructed in three dimensions. Importantly, such glacier reconstructions can be used to derive palaeoclimatic variables for glaciated regions, which allows the terrestrial signature of Younger Dryas climatic change in Scotland to be linked to records from elsewhere.

Considerable research has been undertaken in western and northwestern Scotland to reconstruct glaciers relating to the Younger Dryas (e.g. Ballantyne, 2007a, b; Lukas and Bradwell, 2010; Finlayson et al., 2011). These glacier reconstructions have, in turn, been used to derive palaeoclimatic variables, namely equilibrium line altitudes (ELAs) and palaeoprecipitation estimates. By contrast, the central to eastern Grampians have received comparatively little recent attention (see Bickerdike et al., 2018a, for review). This disparity is problematic in assessing regional palaeoclimatic trends (e.g. precipitation gradients) across Scotland, as well as in efforts to reconcile empirical data from Scotland with atmospheric patterns in the broader amphi-North Atlantic region, as inferred from proxy evidence and modelling (e.g. Isarin et al., 1998; Brauer et al., 2008; Bakke et al., 2009; Lane et al., 2013; Schenk et al., 2018). Thus, investigations to establish the extent and style of Younger Dryas glaciation in the central to eastern Grampians are critical to providing empirically-based precipitation estimates that can be used for regional climatic assessments.

In this contribution, we examine the extent and style of Younger Dryas glaciation in the Gaick, located in the Central Grampians, Scotland (Fig. 1). While the majority of the eastern to central Grampians has been relatively neglected, the glacial-geological evidence in the Gaick has been repeatedly contested, and yet the situation in this important area remains unresolved. During the past

~100 years, several diverging and conflicting models have been postulated to explain the glacial sediment-landform associations in the Gaick (e.g. Barrow et al., 1913; Charlesworth, 1955; Sissons, 1974; Merritt, 2004; Merritt et al., 2004a). Most recently, it was argued that the sediment-landform associations were deposited entirely during retreat of the Last British-Irish Ice Sheet (Merritt, 2004; Merritt et al., 2004a, b), as opposed to during a phase of extensive Younger Dryas icefield glaciation (Sissons, 1974, 1980). This contentious model was refuted for the neighbouring Drumochter Hills (Fig. 1; Benn and Ballantyne, 2005), but debates in the Gaick have been left unsettled.

Recent work has resulted in a detailed geomorphological map of the glacial sediment-landform assemblages in the Gaick, together with targeted sedimentological analyses (Chandler, 2018; Chandler et al., 2018a). These data provide the necessary foundation to now re-evaluate the extent and style of any Younger Dryas glaciers in the Gaick. This contribution therefore synthesises the results of our investigations with the aim of (i) establishing a relative glacial chronology for the Gaick based on morphostratigraphic principles, (ii) reconstructing the extent, morphology and thickness of the Younger Dryas glaciers, (iii) providing estimates of precipitation in the Gaick during this period, and (iv) assessing the wider regional palaeoclimatic implications.

2. Study area

The Gaick is a dissected and undulating plateau, comprising ~520 km^2 in the Central Grampians, Scotland (Fig. 1). It is situated approximately between latitudes 56.834905° and 56.960439°N and longitudes 4.232468° and 3.655369°W (British National Grid; northings: NN 72–NN 88; eastings: NN 63–NN 69). The western Gaick comprises a coherent, gently-undulating plateau that covers ~40 km^2 mainly at altitudes of ~750–850 m OD, but with occasional peaks exceeding 900 m OD. To the east of the plateau is a glacial breach that attains depths of 300–450 m. Further eastwards, the topography is more undulating and is dissected by numerous catchments. As a result, the central Gaick lies at a relatively low altitude, mainly between ~600 and 700 m, but descending to as low as ~520 m in Glen Tarf to the east. The eastern Gaick contains isolated summits, mountain passes (or cols) and interconnected valleys, with the highest summits (>950 m OD) all occurring in the east.

A Neoproterozoic Precambrian succession of siliciclastic psammitic and semipelitic rocks (the ‘Grampian Group’) dominate the bedrock outcrop of the Gaick (see Stephenson and Gould, 1995; Leslie et al., 2006; Smith et al., 2011). The structural architecture is dominated by regional planar foliation and kilometre-scale recumbent folds (the ‘Gaick Fold Complex’), relating to the mid-Ordovician Grampian orogenesis (Leslie et al., 2006). Caledonian brittle deformation (late Silurian to Devonian) is also expressed in the area as major NE-trending faults, which guide several valleys in the Gaick (e.g. Coire Mhic-sith, Glen Tilt; Fig. 1).

In a regional palaeoglaciological context, the Gaick is situated close to (putative) areas of independent Younger Dryas glaciation to the east of the main ‘West Highlands Glacier Complex’ (Fig. 1). These localities include the Monadhliath and adjoining areas (Boston et al., 2015; Boston and Lukas, 2017), Creag Meagaidh (Finlayson, 2006; Jones et al., 2017), Drumochter (Benn and Ballantyne, 2005), the Cairngorms (Sissons, 1979a; Standell, 2014) and the Eastern Grampians (Sissons, 1972; Sissons and Grant, 1972). Given this, and the propensity of upland plateau to act as ‘seeding grounds’ for icefields and ice-caps (e.g. Manley, 1955; Rea and Evans, 2003; Hubbard et al., 2009; Evans et al., 2012), there is a clear need to evaluate the style of glaciation in the Gaick during the Younger Dryas.

3. Previous research

The accepted paradigm for Younger Dryas glaciation in the Gaick is one of extensive plateau icefield glaciation, as argued by Sissons (1974, 1980). Based on the distribution of ‘hummocky moraine’, meltwater channels and associated landforms, Sissons (1974) argued that the area was glaciated by a coherent plateau icefield during the Younger Dryas (Fig. 1a). That reconstruction also depicted convergence of ice flow from the Gaick and neighbouring Drumochter, which fed outlet glaciers in Glen Garry and Glen Truim. Moraines in Drumochter were subsequently linked to contrasting ‘inside and outside’ stratigraphic evidence from Loch Etteridge and the Pass of Drumochter (Fig. 1a), as well as a radiometric basal radiocarbon date from Loch Etteridge, in order to establish a Younger Dryas age (Sissons and Walker, 1974; Walker, 1975a, b). Through extrapolation to the assumed coeval and coalescent Gaick ice, a Younger Dryas age was therefore established for glaciation of the Gaick.

Although widely ‘accepted’ in the literature (cf. Bickerdike et al., 2018a, b), several issues have been identified with Sissons’ (1974) icefield reconstruction. Firstly, more recent studies have reported ice-pushed glaciolacustrine sediments in Coire Mhic-sith, which were formed in a lake dammed by an ice lobe to the southwest (see Lukas and Merritt, 2004; Benn and Ballantyne, 2005; Phillips et al., 2007). The presence of the deformed glaciolacustrine sediments in Coire Mhic-sith implies that ice did not flow southwestwards from the western Gaick plateau to feed ice masses in Glen Truim and Glen Garry (Fig. 1; Lukas, 2004; Lukas and Merritt, 2004; Benn and Ballantyne, 2005), contradicting this element of the earlier reconstruction. Secondly, no geomorphological evidence has been found in support of Gaick-sourced ice flowing southeastwards through Glen na Stalacir’ to connect with the lower part of the Garry ice lobe (Fig. 1; cf. Benn and Ballantyne, 2005), refuting a further component of the icefield reconstruction. Lastly, there are glaciological problems with the plateau icefield reconstruction, such as reconstructed ice divides leaving part of the central Gaick with no clear ice flow direction and depicted ice surface contours resulting in ice lobes having no definitive ice flow direction (see Merritt et al., 2004b, for discussion).

Owing to difficulties with the prevailing paradigm, an alternative model was conceived that argues the glacial landforms and sediments in the Gaick were formed during deglaciation of the last British-Irish Ice Sheet (Merritt, 2004; Merritt et al., 2004a, b). During deglaciation of the Gaick, it was envisaged that a coherent ice-sheet margin retreated actively towards its source from east to west. As deglaciation proceeded, topography was suggested to have had an increasingly important control, with the steep ice-sheet margin becoming fragmented by emerging complex topography. It was thought that these fragmented portions of the ice-sheet margin locally funnelled into valleys and corries. In order to reconcile the unequivocal evidence for upvalley (northerly) retreat in the southern valleys with their envisaged model of west-southwest retreat, it was suggested that the 'outlets' of the retreating ice-sheet margin initially retreated to the north and northwest (i.e. upvalley), forming 'hummocky moraine' and associated ice-marginal features indicative of active retreat (Merritt et al., 2004a). According to this model, ice subsequently retreated actively towards the source of the ice sheet (i.e. Rannoch Moor). Importantly, it was also argued that, if any ice masses were present during the Younger Dryas, they would have been small cirque/niche glaciers (Merritt, 2004).

This alternative explanation of the glacial landforms and sediments in the Gaick (and neighbouring Drumochter) has, however, received little support due to a rebuttal by Benn and Ballantyne (2004, 2005). These authors argued that the evidence in Drumochter was consistently at variance with southwestwards retreat of an ice-sheet margin and interpreted the glacial landforms in Drumochter as the products of a locally-nourished, asymmetric Younger Dryas mountain icefield (Fig. 1). Given the proximity of the Gaick and Drumochter, a corollary of this is that the ice-sheet deglaciation model (in its entirety) is unlikely to have any validity in the Gaick either. Thus, understanding of glacial events in the Gaick area remains open to debate. Resolving these outstanding issues is crucial to elucidating regional climate trends across Scotland, particularly as the Gaick could be the most easterly area of extensive Younger Dryas glaciation in Scotland.

4. Methods

Geomorphological mapping was undertaken through a combination of aerial photograph interpretation and field mapping, following established procedures (see Chandler et al., 2018b). The mapping focused on sediment-landform assemblages that could potentially be used for glacier reconstruction, such as ice-marginal moraines, glaciolacustrine deposits (e.g. grounding-line fans), meltwater channels, upslope limits of glaciogenic sediment cover ('drift limits'), terraces, and periglacial landforms (e.g. solifluction lobes, mountain-top detritus, talus cones). Further details on the geomorphological map production can be found in Chandler et al. (2018a). Sedimentological analyses, following established procedures (e.g. Evans and Benn, 2004; Lukas et al., 2013), were conducted alongside field mapping to elucidate processes of landform formation. Morphostratigraphic procedures used to construct a relative glacial chronology (see Lukas, 2006), along with glacier and palaeoclimatic reconstruction methods, are outlined in later relevant sections.

5. Results

5.1. Geomorphological evidence

The Gaick contains a wealth of glacial landforms and sediments that have been mapped and described in detail by Chandler (2018) and Chandler et al. (2018a). Owing to the size of the area, together with the density and complexity of the geomorphological evidence, we present here only a summary of the glacial geomorphology in the Gaick, alongside map extracts of selected areas (Fig. 2). This is an established approach that allows a considerable body of glacial-geological evidence to be synthesised effectively (cf. Lukas and Bradwell, 2010; Boston et al., 2015).

Clear and consistent variations in the sediment-landform signature can be found across the Gaick (from west to east) and within individual catchments (downvalley to upvalley). In the upper parts of the western valleys and cirques (e.g. Glen Edendon, Gaick Pass, Coire Chais), moraines and intervening meltwater channels are prevalent (Fig. 2). The moraines in these areas, which we classify as 'Type A moraines' (Table 1), are arranged as a series of inset transverse chains that trend obliquely downslope and can be extrapolated to form either (a) nested arcuate loops or (b) chevron patterns representing former glacier margins (cf. Lukas and Benn, 2006; Boston, 2012a, b). Based on their spatial arrangement, these ice-marginal moraines indicate upvalley retreat of Gaick-sourced ice masses. Typically, the lateral components of individual moraine arcs/chains exhibit considerably steeper gradients (~15–20 m per 100 m) on valley sides than those displayed by moraine chains found in other parts of the Gaick. The upper parts of the western Gaick valleys, where Type A moraines occur, usually contain a maximum of one river terrace level, and/or the upvalley limit of a major river terrace may terminate abruptly at the downvalley limit of Type A moraines. The valley slopes in these areas are plastered with glaciogenic sediment, while extensive talus/scree deposits are entirely absent.

The above sediment-landform associations contrast with those found in the lower parts of the western Gaick valleys (e.g. Glen Edendon, Gaick Pass) and in catchments elsewhere in the area (e.g. Glen Bruar, Glen Chaorainn, Glen Tarf). In those areas, 'Type B moraines' are the dominant moraine type, and Type A moraines are absent (Table 1; Fig. 2). As is the case for Type A, Type B moraines are ice-marginal and document upvalley retreat of Gaick-sourced ice. Although morphologically similar to Type A, Type B moraines are distinguishable based on three defining features: Firstly, Type B moraines are frequently restricted to a narrow band at the slope foot and do not trend obliquely up the valley slopes; additionally, lateral components of the moraine arcs/chains may exhibit much shallower gradients (often ~5 m per 100 m) than those categorised as Type A. Secondly, there may be a clear contrast in the slope features associated with the two moraine types, with the shallower Type B moraines associated with scree/talus-covered slopes (e.g. in Gaick Pass). In this case, the scree/talus normally descends to the valley floor and the upper surface of the moraines. Lastly, Type B moraines frequently occur in association with one or more prominent river terrace levels, contrasting with the sediment-landform associations in the upper parts of the western Gaick where terraces are often absent.

The lowermost parts of the valleys, along with areas around the periphery of the Gaick, are characterised by a further distinct landform signature. In these areas, a third moraine type ('Type C moraines'; Table 1) can be distinguished, which encompasses moraines that are inconsistent with deposition purely by outwardly-radiating glaciers (Fig. 2). That is, the moraine patterns indicate that these landforms were not solely formed by Gaick-sourced cirque, valley or outlet glaciers. Type C moraines manifest as (i) topographically-discordant moraines, (ii) cross-cut moraines, or (iii) moraines that occur around the flanks of the Gaick and document the retreat of external ice. Topographically-discordant moraines are those that descend in an *upvalley* direction (e.g. in Glen Edendon, Glas Choire, Glen Geldie). In the southern Gaick valleys, the topographically-discordant moraines are cross-cut/overprinted by moraines descending in the opposite (downvalley) direction.

There are also notable contrasts in the spatial distribution of what we term 'dissected glaciogenic material', which only occurs in the lower halves of the western valleys and in the eastern Gaick (Fig. 2). Dissected glaciogenic material represents a pre-existing, undulating glaciogenic sediment cover that has been dissected into mounds and ridges by fluvial activity, giving rise to a distinctive jigsaw-like erosional planform (cf. Chandler, 2018; Chandler et al., 2018a). This sediment-landform assemblage may occur in association and contiguously with Type B and C moraines, i.e. there is no sharp boundary between the moraines and dissected glaciogenic material or a distinct change in altitude between the landforms (e.g. Glen Tarf; Fig. 2). Conversely, dissected glaciogenic material and Type A moraines have a mutually-exclusive spatial relationship.

Clear contrasts are also apparent in the type and spatial extent of periglacial landforms on the higher ground in the Gaick. In the western Gaick, solifluction deposits and discontinuous mountain-top detritus are restricted to the periphery of the plateau and interfluves on the western flank. Instead, the plateau is characterised by meltwater channels and peat cover. Conversely, periglacial features are considerably more extensive in the eastern Gaick and include expansive areas of solifluction lobes and mountain-top detritus, along with numerous frost-shattered bedrock outcrops. Significantly, there are areas in the central, southern and northern sectors where meltwater channels occur in close association with, and are partially obscured by, periglacial phenomena. Thus, the relative relationships of the meltwater channels and periglacial deposits indicate that the meltwater channels pre-date the formation of the periglacial features in those areas. Moreover, the contrast in the prevalence of periglacial features between the western and eastern Gaick implies that those areas may have experienced different conditions (e.g. ice-covered vs. ice-free) during the most recent period of intense periglacial activity (i.e. the Younger Dryas; cf. Ballantyne and Harris, 1994; Ballantyne, 1998; Lukas, 2006).

5.2. Chronology

Common strategies for constraining the timing of glaciation in Scotland (and elsewhere) are to (a) use 'inside and outside' stratigraphic evidence and basal radiocarbon dates (e.g. Benn et al., 1992), or (b) directly date glacier limits (moraines) by applying cosmogenic nuclide methods to surface boulders (e.g. Lukas and Bradwell, 2010; Finlayson et al., 2011). Unfortunately, a paucity of suitable sampling sites and dateable material precluded the use of these geochronological methods in the present study. Specifically, surface boulders for cosmogenic nuclide dating are almost entirely absent from moraine surfaces (*contra* Sissons, 1974: p. 95) and, with respect to radiocarbon dating, there is a lack of strategically-located sampling sites for testing the age of inferred limits (i.e. both inside and outside a limit). In the absence of any published or new radiometric dates from the Gaick, we construct a relative glacial chronology using morphostratigraphy.

Essentially, a morphostratigraphic approach uses the relative spatial relationships between landforms (overprinting, juxtaposition), together with an assessment of systematic variations in the sediment-landform associations over an area (e.g. contrasts between the lower and upper parts of valleys), to establish a relative sequence for glacial events. Morphostratigraphy is now a well-established relative dating tool for differentiating between former distinct phases of glaciation in the Scottish Highlands, and the approach has been independently tested and verified through radiometric dating (e.g. Lukas and Bradwell, 2010; Finlayson et al., 2011). We thus refer readers to previous publications for more detailed description and criteria (see Lukas, 2006; Boston et al., 2015).

The relative chronology discussed below (Section 5.2.1) provides the fundamental basis for a rigorous and targeted radiometric dating programme, in which the sampling strategy is specifically designed to test the maximum glacier limits defined based on observed morphostratigraphic signatures. Moreover, the morphostratigraphic framework will facilitate extrapolation of radiometric dates across the Gaick from any glacier limits that become better constrained in the future, which is necessary (at least to some degree) in all localities across Scotland (cf. Lukas, 2006; Boston et al., 2015; Bickerdike et al., 2018a).

5.2.1. Morphostratigraphic framework

Applying the morphostratigraphic criteria defined by Lukas (2006) to the Gaick, we recognise three different glacial sediment-landform assemblages. The first sediment-landform assemblage is only found in the upper parts of the western Gaick catchments and is characterised by several defining, and converging, elements (see also Section 5.1):

(1) Type A moraines (see Table 1) are found exclusively in this sediment-landform assemblage (i.e. only in the western Gaick valleys), and they exhibit spatial patterns that are consistent with ice-marginal moraines and indicative of upvalley retreat of Gaick-sourced glaciers.
(2) Clear contrasts exist in river terrace patterns either side of the outer (downvalley) limits of the Type A moraines, namely a prominent river terrace may terminate downvalley (outside) of the outermost Type A moraines and/or terraces are absent upvalley (inside) of the moraine limits.
(3) Areas inside (upvalley) of the Type A moraine limits (i.e. the upper parts of the western valleys) are characterised by sediment-covered slopes, while extensive talus/scree deposits are absent. By contrast, slopes outside (downvalley) of the Type A moraine limit may display extensive talus or scree cover, which may descend to the slope foot and upper surface of Type B moraines.
(4) Dissected glaciogenic material is absent from the upper parts of the western valleys, and it has a mutually-exclusive relationship with Type A moraines. Conversely, both Type B and C moraines are interspersed with dissected glaciogenic material in some areas, i.e. moraines of those two types occur both downvalley and upvalley of dissected glaciogenic material (Fig. 2).
(5) The sediment-landform associations identified above are only evident in catchments around the western Gaick plateau (i.e. in the upper parts of the western valleys and cirques), which is largely devoid of relict periglacial landforms (e.g. mountain-top detritus, solifluction deposits).

The relative spatial relationships between sediment-landform associations (i.e. river terraces, sediment-covered slopes, talus/scree) inside and outside the limits of Type A moraines, as outlined above, are consistent with observations from inside (upvalley) and outside (downvalley) of Younger Dryas glacier limits elsewhere in Scotland (e.g. Benn and Ballantyne, 2005; Lukas, 2006; Lukas and Bradwell, 2010; Finlayson et al., 2011; Boston et al., 2015). The mutually exclusive relationship between Type A moraines and dissected glaciogenic material is compatible with these patterns, as dissected glaciogenic material has not previously been reported from Younger Dryas glacial landsystems and most likely indicates a different phase and style of glaciation (cf. Bickerdike et al., 2016, 2018a, b, Chandler, 2018, and references therein). Moreover, the mutual exclusion of the sediment-landform assemblage found in upper parts of the western valleys from higher ground with periglacial phenomena (i.e. in the east) may indicate that the western plateau was 'protected' by plateau ice during the last period of intense frost weathering in Scotland, i.e. the Younger Dryas (cf. Ballantyne and Harris, 1994; Lukas, 2006). Based on these multiple, converging lines of evidence, together with similarities in the spatial arrangements exhibited by Type A moraines and Younger Dryas ice-marginal moraines found elsewhere in Scotland (cf. Benn et al., 1992; Bennett and Boulton, 1993a, b; Lukas and Benn, 2006), we argue that the sediment-landform assemblage found in the upper parts of the western Gaick valleys relates to the Younger Dryas (~12.9–11.7 ka). Since this sediment-landform signature is restricted to the western Gaick (Table 2; Fig. 3), the morphostratigraphic relationships suggest that the Gaick experienced spatially-restricted glaciation during the Younger Dryas.

Our assignment of landforms only in the upper parts of the western valleys to the Younger Dryas therefore implies that the sediment-landform assemblages found elsewhere in the Gaick must relate to glacial events that pre-date the Younger Dryas. The two remaining sediment-landform assemblages are characterised by clearly distinguishable moraine patterns: the second sediment-landform assemblage, which occurs in lower parts of the western valleys and is dominant in catchments elsewhere within the Gaick (i.e. in the east), includes Type B moraines (see Table 1). Type B moraines are topographically concordant (i.e. descend downslope in a downvalley direction) and therefore relate to a phase of local glaciation. By contrast, the third sediment-landform assemblage is characterised by moraines (Type C) that are incompatible with radial flow of Gaick-sourced glaciers and thus relate to retreat of external ice from the area, most likely during deglaciation of the Last British-Irish Ice Sheet. Cross-cutting between Type C moraines in the south-facing valleys (see Section 5.1) provides evidence for the 'unzipping' of external ice and Gaick-sourced glaciers, with subsequent re-advance of the outlet/valley glaciers and overprinting of the moraines (Chandler, 2018). The relative relationships of moraines, with Type B moraines occurring *upvalley* of Type C moraines but *downvalley* of Type A (i.e. Younger Dryas) moraines, suggests that Type B moraines relate to a phase of local glaciation that occurred following thinning and retreat of the Last British Ice Sheet, but prior to the Younger Dryas. We suggest that this most likely occurred at the end of the Dimlington Stadial (correlated with Greenland Stadial 2 (GS-2); Lowe et al., 2008), as has similarly been suggested in the Monadhliath (cf. Boston et al., 2015), Cairngorms (cf. Everest and Kubik, 2006) and elsewhere in Scotland (see Ballantyne and Small, 2018, for review).

5.2.2. Regional chronological framework

Although no radiometric dates are currently available for the Gaick, our relative glacial chronology can be placed within a regional chronological framework. Recalibrated conventional and AMS basal radiocarbon dates from Loch Etteridge, located ~5.5 km to the northwest (Fig. 1), yielded dates of 16,989–14,398 cal a BP and 15,655–15,255 cal a BP, respectively (recalculated using OxCal; Sissons and Walker, 1974; Everest and Golledge, 2004). These radiocarbon dates indicate that Loch Etteridge was ice free throughout the Younger Dryas and last deglaciated before ~16.9–14.4 ka. Conversely, no Lateglacial organic and minerogenic deposits have been retrieved from the Pass of Drumochter, situated at the southwestern edge of the Gaick (Fig. 1). Basal sediments from that site instead indirectly correlate with Holocene organic sediments at Loch Etteridge that yielded a radiocarbon age of 9405 ± 260 ^{14}C a BP (11,399–9,912 cal a BP; recalculated using OxCal; Walker, 1975a). The contrasting stratigraphic evidence from these two sites, together with the available radiocarbon dates, thus implies that Drumochter was glaciated during the Younger Dryas (cf. Benn and Ballantyne, 2005). The morphostratigraphic relationships that we have identified in the western Gaick (see Section 5.2.1) are very similar to, and consistent with, those independently dated in Drumochter. Thus, the available evidence suggests that the most recent glacial phase in these adjacent areas was broadly penecontemporaneous, i.e. it occurred during the Younger Dryas.

To the northeast of the Gaick, sediment-landform signatures similar to those that we observed in the Gaick have been independently dated using cosmogenic nuclide exposure dating in Glen Geusachan, southern Cairngorms (Fig. 1). Everest and Kubik (2006) obtained pre-dominantly pre-Younger Dryas ages from moraines (≈ Type B) in the downvalley portions of Glen Geusachan: ages ranged from 17.4 ± 1.1 ka to 12.8 ± 1.1 ka (arithmetic mean ~14.6 ka) when recalculated using the 'local' production rate from northwest Scotland (NWHLPR11.6; Ballantyne and Stone, 2012), erosion rates of 1 mm ka^{-1} and Lm scaling (Lal, 1991; Stone, 2000). Conversely, Standell (2014) obtained ages of 11.9 ± 0.6 ka, 10.8 ± 0.6 ka and 12.1 ± 0.7 ka (arithmetic mean ~11.6 ka) from moraines (≈ Type A) located further upvalley, when calculated using the Loch Lomond local production rate (LLPR; Fabel et al., 2012), erosion rates of 1 mm ka^{-1} and Lm scaling. The ages of the glacial landforms in the Gaick, as inferred from independently-dated morphostratigraphic relationships elsewhere in Scotland (e.g. Lukas, 2006; Lukas and Bradwell, 2010; Finlayson et al., 2011), are thus supported by analogy to dated glacier limits in the wider region.

5.3. Plateau glacierisation threshold analysis

Our revised interpretation of the glacial geomorphological record in the Gaick challenges the accepted paradigm of extensive Younger Dryas glaciation in the Gaick (Sissons, 1974), and it also contrasts with the alternative argument of only small cirque/niche glaciers during the Younger Dryas (Merritt, 2004). In view of this, and in the absence of radiometric dates for the Gaick, we here examine the concept of a critical threshold necessary for plateau glacierisation in order to independently test the Younger Dryas glacier limits identified from clear morphostratigraphic relationships.

Observations from modern plateau icefields have shown that there is a close, non-linear relationship between summit breadth and altitude above the regional firn line, whereby smaller summits/plateaux must attain a higher altitude to sustain an icefield (see Manley, 1955, 1959; Rea et al., 1998; Rea and Evans, 2003). We build on this concept here and use it to evaluate whether the Gaick summits/plateaux were capable of sustaining plateau ice cover during the Younger Dryas (see also Supplementary Material). For this purpose, it is necessary to extrapolate a regional firn line across the Gaick. In our analysis, we assume that the firn line and ELA are equivalent (cf. Porter, 2001) and take published, empirical ELA estimates from areas around the Gaick to establish a regional ELA gradient that, in turn, can be used to derive hypothetical ELAs for the Gaick. Four scenarios were defined that recognise an increase in the ELA in a north–east direction with horizontal distance across the region from a ‘reference point’ (Drumochter; Fig. 1) to the southwest:

(1) The ELA gradient (**5.96 m per km**) established for the Monadhliath plateau icefield (Boston et al., 2015) is employed as a lower end member, which considers west-east ELA variations across a nearby plateau icefield. This gradient would mean a non-linear ELA gradient existed between Drumochter (to the southwest) and the Cairngorms (to the northeast).

(2) The second scenario is based on the ELA gradient (**6.53 m per km**) between Drumochter (ELA = 626 m; Benn and Ballantyne, 2005) and the Cairngorms (ELA = 920 m), where the Cairngorms value is based on the ‘cirque glacier’ configuration of Standell (2014).

(3) The third scenario uses the regional ELA value (942 m) calculated by Sissons (1979a) for cirque-style and valley-style glaciers in the Cairngorms to derive an ELA gradient between Drumochter and the Cairngorms (**7.02 m per km**).

(4) The upper end member (**7.47 m per km**) is based on a plateau-ice configuration for the Cairngorms (Standell, 2014). The additional ice-mass input from plateaux to surrounding outlet glaciers substantially increases the ELA value (962 m) in comparison to that for the ‘cirque glacier’ configuration (Standell, 2014), as a result of the increase in contributing areas at higher altitudes (Rea et al., 1999). This fourth scenario incorporates ‘candidate’ (potential) Younger Dryas glaciers identified by Standell (2014), which are sites where the “geomorphological evidence is uncertain” or “limited” (Standell, 2014: p. 282).

Using the above ELA gradients, hypothetical ELA values have been calculated for the Gaick and then employed to establish areas that lie above the calculated, hypothetical ELAs (Fig. 4), whereby the ELA value was calculated for 1 x 1 km grid squares and then the NEXTMap DSM was classified to show relative heights above the hypothetical ELA (see Supplementary Material).

The analysis shows that there is a clear trend of decreasing area above the hypothetical ELA eastwards across the Gaick, with the eastern sectors more strongly influenced by the prescribed ELA gradient (Figs. 4 and 5). Irrespective of the gradient employed, only the western Gaick contains large areas >100 m above the hypothetical ELA. Significantly, these areas correspond well with the Younger Dryas glacier limits inferred from clear morphostratigraphic relationships (Section 5.2). Elsewhere, isolated areas attain altitudes >100 m above the hypothetical ELA, but these are considerably less extensive and occur on relatively narrow, rounded summits (Fig. 4). Indeed, large areas of the central and eastern Gaick lie at altitudes below the projected ELAs, leaving no clear source area for valleys with abundant glacial landforms, such as Glen Bruar and Glen Tarf.

We have also used the hypothetical ELA data to establish where selected summits (n = 47) in the Gaick plot against the 'Manley curve' (Fig. 6; Manley, 1955). Summit breadths were calculated following the definition of Manley (1955), where summit breadth is the horizontal distance across the area lying within 30 m altitude of the actual summit in the direction of the prevailing winds. In this case, the dominant palaeo-wind direction in Scotland during the Younger Dryas is thought to have been from the southwest, as inferred from the spatial distribution of glacial and periglacial landforms (e.g. Sissons, 1979b, c, 1980; Ballantyne and Kirkbride, 1986; Ballantyne, 1989, 2002a) and from observations of *in situ* relict wind-polished boulders and bedrock (Christiansen, 2004).

Plotting of the Gaick summit data against the Manley curve shows that only summits in the west plot above the curve (Fig. 6). Several summits in the western sector (n = 5) plot well inside the envelope of uncertainty, along with a single summit from the northern Gaick. Summits in the central (n = 3), northern (n = 2) and southern (n = 2) sectors also plot inside the envelope of uncertainty, but closer to the boundary (Fig. 6). Indeed, the two northern summits fall outside the envelope of uncertainty in scenarios 3 and 4 (Figs. 6c and d). Overall, most summits plot below the critical curve and outside the envelope of uncertainty, particularly in the case of the southern Gaick.

Our plateau glacierisation threshold analysis provides independent support for only a spatially-restricted Younger Dryas icefield in the western Gaick. Furthermore, the occurrence of several western summits in the envelope of uncertainty (Fig. 7) implies that the Gaick experienced only marginal glacial conditions. Such conditions would have limited the proclivity of the summits to nourish plateau ice. However, there is a caveat to the above analysis: the definition of summit breadth may be too strict, particularly where summits/plateaux are broad and relatively flat (cf. Rea et al., 1998). In this scenario, a single high point 30 m above the surrounding (average) plateau level would discount the plateau from sustaining ice cover. Moreover, an ice mass on a broad plateau may be sufficiently thick to attain a height above the regional firn line, irrespective of whether the plateau itself lies above the firn line.

Despite the above uncertainties, the multiple lines of independent evidence from the glacial geological record and the glacierisation threshold analysis suggests that (a) only the western Gaick sustained ice during the Younger Dryas and (b) conditions elsewhere in the area were unfavourable for glacierisation.

5.4. Younger Dryas icefield reconstruction

The clarity and spatial coherence of the geomorphological evidence pertaining to the Younger Dryas, together with the clearly definable glacier limits, allow the reconstruction of a former plateau icefield. Following identification of the glacier limits using morphostratigraphy (Table 2), outlet glaciers were reconstructed in ArcMap™ using available geomorphological evidence to constrain their latero-frontal limits (e.g. ice-marginal moraines, meltwater channels, 'drift limits'). This process necessitated both interpolation between individual 'marker points' (e.g. individual moraines) and extrapolation of former ice surface gradients to the opposite valley sides, where geomorphological constraints were insufficient or absent. In the latter case, the ice surface gradient dictated by the geomorphological evidence on one side of the valley was mirrored on the opposite slope to produce a symmetrical glacier tongue. Although potentially problematic due to its subjectivity (cf. Ng et al., 2010), this extrapolative approach provided an efficient means to reconstruct former glacier margins where lateral limits were well constrained, as has been done elsewhere in Scotland (e.g. Ballantyne, 1989, 2002; 2007a, b; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Boston et al., 2015).

Although the Younger Dryas geomorphological signature is well preserved in the Gaick, constraints on former ice surface elevations were limited, and the thickness of the ice on the plateau was difficult to constrain. To resolve this, two 2D glacier surface profile models were used to (a) test the assumption that all the identified outlet glaciers were connected to ice on the western Gaick plateau (i.e. plateau vs. alpine style of glaciation) and (b) constrain ice thickness on the plateau surface. Specifically, we used the 'perfectly plastic' flowline model of Benn and Hulton (2010) and a model developed by Ng et al. (2010) that implements Nye's (1951, 1952) theoretical parabola as a surface profile descriptor (for technical details, see Supplementary Material). In total, we modelled glacier surface profiles along 23 flowlines, with hypothetical ice divides defined by the positions of present-day watersheds (Fig. 7).

The modelling results and geomorphological evidence were integrated using the procedure devised by Boston et al. (2015), which acknowledges and quantifies varying uncertainties associated with plateau ice thickness (Fig. 8). Model outputs were employed in combination with available geomorphological constraints to establish realistic minimum and maximum elevation boundaries for the former ice surface. In scenarios where glacier surface profiles were modelled either side of a col, the minimum ice surface elevation was taken as the highest minimum constraint, as only one modelled surface profile needs to cross above the col to imply that plateau ice feasibly submerged the col (cf. Ng et al., 2010; Boston et al., 2015). Conversely, the lowest maximum constraint was taken as the maximum glacier surface. Together, these values define an 'envelope of likelihood' based on evidence in neighbouring catchments, i.e. catchments situated either side of an ice divide (Fig. 8). Where geomorphological evidence and modelling was only available for one side of a col, a similar procedure was followed to define minimum (highest minimum constraint) and maximum (lowest maximum constraint) boundaries, but with lower confidence as the envelope is based on evidence from a single catchment.

At most of the ice divides, the envelopes of likelihood were defined based on surface profiles obtained from the Benn and Hulton (2010) model (Table 3). For seven cases, the ice surface elevations were corroborated by 'typical' values obtained using the Ng et al. (2010) modelling approach, whilst the maximum values from the Ng et al. (2010) equation are consistent with the envelopes for three other cases (flowlines 14, 15 and 23; Table 3). This adds confidence to the defined envelopes of likelihood. Circumstantial evidence from nearby ice divides provided additional confidence, with ice surface elevations for neighbouring divides corresponding well (e.g. ice divides c–f; Fig. 7 and Table 3). To further test the envelopes of likelihood defined using the main flowlines, glacier surface profiles were also modelled along the flowline from the southern Loch an Dùin limit (see Fig. 7). The results agreed with the defined envelopes, with ice surface elevations estimated by the Benn and Hulton (2010) model occurring within the range of values established for the main flowline.

The Gaick Icefield was manually reconstructed in ArcMap™ by integrating (i) the Younger Dryas limits identified using morphostratigraphy, (ii) extrapolated ice surface gradients from geomorphological constraints (e.g. lateral moraines, lateral meltwater channels) and (iii) the envelopes of likelihood for the ice surface elevations (Table 3). For areas with little or no geomorphological constraint on the ice surface (i.e. on the plateau), the envelopes of likelihood were utilised to establish minimum, mid-range and maximum ice surface elevations. That is, the lower boundary of the envelope of likelihood was taken as the minimum ice surface elevation, the median value as the mid-range ice surface elevation, and the upper boundary as the maximum ice surface elevation. The final reconstructed ice surfaces also took into consideration topographic constraints (e.g. relief either side of a col) and circumstantial evidence (e.g. nearby geomorphological evidence, ice surface elevations at neighbouring cols). Following Boston et al. (2015), all the available evidence and modelling data were used to produce minimum, mid-range and maximum reconstructions for the Gaick Icefield. This allows varying levels of uncertainty associated with the icefield reconstruction to be quantified in subsequent ELA and palaeoprecipitation calculations (see Section 5.5).

Ice surface contours were manually drawn in ArcMap™ at 50 m intervals to depict ice surface elevation, with their position estimated based on the underlying topography. In some cases, ice surface contours were partly guided by surface profiles produced by the Benn and Hulton (2010) model. However, the modelling was generally only used to establish ice surface elevations at cols (ice divides), owing to uncertainties relating to ice fluxes (convergent and divergent flow) along the glacier flowlines. In accordance with contour patterns observed on modern glaciers, the ice surface contours were drawn so that they displayed increasingly convex and concave forms in the downglacier and upglacier directions, respectively. This is a standard procedure for delineating glacier surface contours on reconstructions (e.g. Ballantyne, 1989, 2002a, 2007a, b; Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Boston et al., 2015). Although this manual approach is partly subjective, it was favoured over geostatistical interpolation methods (e.g. Barr and Clark, 2011; Carrasco et al., 2013; Pellitero et al., 2016) due to (i) uncertainties over the applicability of the Benn and Hulton (2010) model to the complex terrain in the Gaick (e.g. ice fluxes at topographic junctions) and (ii) the absence of geomorphologically-constrained surface profiles in all catchments (Fig. 7).

The integrated approach indicates that a spatially-restricted (~42 km^2) and highly asymmetric plateau icefield developed over the Gaick during the Younger Dryas (Figs. 9 and 10). On the (north)western flank of the icefield, ice is restricted to the plateau rim and small outlet glaciers, whilst large outlets flowed into Glen Edendon and the Gaick Pass on the eastern side of the Gaick Icefield. The most complex sector of the icefield is at the Cama' Choire – Glen Edendon junction (NN 705 797), where three outlets converged to form the large Edendon glacier. Further downvalley, the Edendon glacier became coalescent with the Glas Glacier (NN 722 782) and Meadair glacier (NN 708 781).

Geomorphological and sedimentary evidence from the valley of Loch an Dùin indicates that ice diverged from the Edendon and Gaick glaciers and flowed into either end of the valley, blocking drainage (Chandler, 2018; Chandler et al., 2018a). Sedimentary evidence for glaciolacustrine deposition takes the form of a glaciotectonised grounding-line fan and a gravitationally-modified Gilbert-type delta at the northern end of the lake (Chandler, 2018). The evidence from this valley can be correlated morphostratigraphically with the Younger Dryas limit of the Gaick Glacier (Figs. 9 and 10).

Overall, the different scenarios have a limited influence on the position of the ice surface contours (Fig. 10), as indicated by the small envelopes of likelihood (typically <30 m; see Table 3). The tight constraint on the range of ice surface elevations is likely to at least partly be a function of flowline convergence, with most flowlines joining the main Edendon and Gaick Pass outlet flowlines (Fig. 7). Consequently, large parts of the flowlines were replicated several times during the modelling process, reducing the potential range of ice surface elevations on the plateau.

5.5. Younger Dryas palaeoclimatic reconstruction

We now use the three-dimensional reconstructions of the Younger Dryas Gaick Icefield (Figs. 9 and 10) to establish the palaeoclimatic boundary conditions. The procedures employed to derive palaeoclimatic variables from Younger Dryas glaciers in Scotland are now well-established in the literature; thus, for brevity, the underlying principles and assumptions of ELA and palaeoclimate calculations are not discussed in detail here. Thorough explanations of all the procedures we employed can be found in Boston et al. (2015) and references therein.

5.5.1. Equilibrium line altitude reconstruction

The ELA provides a sensitive link between glaciers and climate that can be used to derive palaeoclimatic variables from reconstructed ice masses (e.g. Bakke et al., 2005a, b; Benn and Ballantyne, 2005; Lukas, 2007; Mills et al., 2012; Barr and Clark, 2011; Boston et al., 2015; Pellitero et al., 2015). Various methods can be used to calculate ELAs for palaeo-ice masses, which are based on the underlying assumption that, for steady-state conditions, there are known relationships between the ELA and glacier geometry (see Meierding, 1982; Benn et al., 2005; Pellitero et al., 2015). In this study, we used the accumulation area altitude ratio (AAR) and area-altitude balance ratio (AABR) methods, approaches that have been applied in various topographic and glaciological settings (e.g. Lukas, 2007; Stansell et al., 2007; Mills et al., 2009, 2012; Bendle and Glasser, 2012; Carrasco et al., 2013; Boston et al., 2015). ELAs for the minimum, mid-range and maximum Gaick Icefield reconstructions were calculated using the automated ArcGIS toolbox of Pellitero et al. (2015).

The ELA calculations employed a range of AARs and AABRs (Table 4), following established procedures (see Boston et al., 2015, and references therein). The AABR method is widely assessed as the most reliable approach in approximating the 'true' climatic ELA at present, since it incorporates both glacier hypsometry and variable mass balance gradients in the computation of palaeo-ELAs (cf. Osmaston, 2005; Benn and Ballantyne, 2005; Benn et al., 2005; Rea, 2009). For modern mid-latitude maritime glaciers, a balance ratio (BR) of 1.91 ± 0.81 (± is one standard deviation) is considered representative (cf. Rea, 2009; Boston et al., 2015, for further details). Thus, a BR of 1.91 ±

0.81 is argued to provide a credible range of values for Younger Dryas ice masses in Scotland, and ELA3 (Table 4) is considered the most appropriate ELA value for palaeoclimatic reconstruction.

Using a BR of 1.91 ± 0.81, and taking into account uncertainties associated with the thickness of the reconstructed icefield, yields an ELA estimate of 751 ± 46 m for the Gaick Icefield (ELA3; Table 4). ELAs calculated for the individual glacier basins are relatively consistent across the icefield, with all glacier basins apart from Glas Choire occurring within ~15% of one another (725–820 m) and within the associated uncertainties. Moreover, the different ELA calculation methods yield values within ~10% of each other, demonstrating a reasonable level of agreement and indicating that, within potential errors, the ELAs are largely indistinguishable. This general consistency between the different methods has been shown in previous ELA reconstructions for Younger Dryas glaciers across Scotland (e.g. Benn and Ballantyne, 2005; Lukas and Bradwell, 2010; Boston et al., 2015).

Although the ELAs are generally consistent for the main catchments, the low ELA (574 ± 26 m) calculated for the Glas Glacier merits further discussion, since it is ~22–25% lower than the ELA for the whole icefield (751 ± 46 m). This difference implies that the Glas Glacier may have only been viable due to site-specific (topoclimatic) conditions. Topoclimatic factors (e.g. snowblow, avalanching, topographic shading) can play an important role in augmenting glacier mass balance and lowering ELAs (e.g. Sissons and Sutherland, 1976; Mitchell, 1996; Nesje and Dahl, 2000; Coleman et al., 2009; Mills et al., 2009, 2012; Chandler and Lukas, 2017). Previous Scottish studies have shown that an eastward decline in ELAs is a characteristic feature of many Younger Dryas icefields, including the neighbouring Drumochter Icefield. This pattern has been argued to reflect eastward snow redistribution by strong westerly winds (e.g. Ballantyne, 1989, 2002a; Benn and Ballantyne, 2005; Finlayson et al., 2011; Pearce, 2014), but there is no plateau area to the west of Glas Choire. Thus, redistribution of snow by westerlies is unlikely to explain the low ELA value of the Glas Glacier.

A second dominant palaeo-wind direction, from the northwest to north, has also been identified by measuring the orientation of windpolished forms on boulder surfaces in the Scottish Highlands, including in the adjacent Drumochter area (Christiansen, 2004). Given the presence of a plateau surface to the northwest/north of Glas Choire (see Figs. 7 and 9), snow may have been transferred to the former glacier surface through snowblow from the plateau, thereby augmenting glacier mass balance and lowering the ELA (cf. Coleman et al., 2009; Mills et al., 2009, 2012; Chandler and Lukas, 2017). In addition, the Glas Glacier catchment contains a substantial basin – one of the few well-developed cirques in the area – that could have allowed the preferential accumulation of snow from snow-bearing winds from the northwest to north. Thus, it is possible that these topoclimatic factors counteracted the effects of enhanced direct insolation on south-facing glaciers and provided propitious conditions for nourishing a valley glacier in Glas Choire; it seems highly unlikely that the Glas Glacier could have existed without the influence of snowblow, given the substantial difference in the ELA values.

An alternative explanation for the lower ELA value is that the glacial landforms used to delimit the Glas Glacier may instead pre-date the Younger Dryas. Despite the strong and consistent morphostratigraphic relationships evident in the Gaick, the possibility that all glacial landforms in Glas Choire pre-date the Younger Dryas cannot be ruled out altogether, given that the morphostratigraphic framework in the Gaick has yet to be further tested by radiometric dating. That said, the morphostratigraphic signature provides high confidence that the landforms pertaining to this reconstructed glacier and those further downvalley (i.e. outside the inferred former glacier limit) relate to distinct glacial phases.

5.5.2. Palaeoprecipitation reconstruction

Since palaeo-ELA calculations assume that the reconstructed glacier was in equilibrium with climate, palaeoclimatic conditions for a former glacial period can be established using empirical relationships between summer temperature and annual precipitation at the ELA (e.g. Kerschner et al., 2000; Benn and Ballantyne, 2005; Stansell et al., 2007; Rea and Evans, 2007; Mills et al., 2012; Barr and Clark, 2011; Boston et al., 2015). In most cases, the Ohmura et al. (1992) empirical relationship has been used for glacier-based climatic reconstructions. This relationship is based on a dataset of modern mid- and high-latitude glaciers and is defined as

$$P = 9T_a^2 + 296T_a + 645 \qquad (1)$$

where P is the sum of winter accumulation and summer precipitation (mm a^{-1}) (water equivalent) and T_a is the mean ablation season temperature (°C) at the ELA. This relationship allows either temperature or precipitation to be calculated where an independent value is known, and it has been used extensively to derive palaeoprecipitation estimates using an independent temperature proxy, typically mean July temperatures inferred from subfossil assemblages of chironomidae (e.g. Benn and Ballantyne, 2005; Lukas, 2007; Lukas and Bradwell, 2010; Boston et al., 2015).

The suitability of the simple Ohmura et al. (1992) relationship for calculating palaeoprecipitation has, however, been the subject of much debate, owing to suggestions that the precipitation-temperature relationship is influenced by more complex relationships between air temperature, energy balance and topographic factors at local to regional scales (cf. Braithwaite, 2008; Hughes and Braithwaite, 2008; Lukas and Bradwell, 2010; Golledge et al., 2010, and references therein). Conversely, it has been argued that amalgamating the global data and ‘smoothing’ local variations is advantageous for calculating palaeoclimate data from glacier reconstructions, since it is more reliable than attempting to account for local variability in locations where no modern glacier data are available (e.g. Benn and Ballantyne, 2005). Aside from these issues, the Ohmura et al. (1992) global dataset incorporates ice masses with various glaciodynamic regimes, including surging and calving glaciers, which would have significant implications for glacier geometry and thus ELAs. In turn, this would affect glacier-climate relationships at the ELA, i.e. the ELA may be 'out of phase' with climate. Despite these potential shortcomings, the Ohmura et al. (1992) relationship has been used for many glacier-based palaeoclimatic reconstructions in Scotland owing to the absence of a more suitable equation.

Central to the debate regarding the use of the Ohmura et al. (1992) relationship for palaeoclimatic reconstructions in Scotland are discrepancies between palaeoprecipitation values derived using Eq. 1 and general circulation models, with the latter suggesting a much drier climate during the Younger Dryas (cf. Golledge et al., 2010; Boston et al., 2015, for further details). To address this issue, a new precipitation-temperature (*P/T*) function specific to the Younger Dryas in Scotland was advocated by Golledge et al. (2010):

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

where P is effective precipitation, T_a is the mean temperature (°C) of the ablation season and S is the seasonality factor; S = 1 for neutral-type; S = 1.4 for summer-dominated or S = 0.8 for winter-dominated precipitation. This *P/T* function permits an annual temperature range of 30°C, which is consistent with available biological proxy data (e.g. Atkinson et al., 1987; Isarin et al., 1998; Witte et al., 1998; Lie and Paasche, 2006).

We used both *P/T* equations (Eqs. 1 and 2) to calculate palaeoprecipitation for the Gaick in order to facilitate comparison with previous studies and between the two approaches. Quantitative estimates of palaeotemperatures are required input for both equations and a summer temperature at sea level of 8.5 ± 0.3°C (mean July temperature) was used, in accordance with recent studies (see Boston et al., 2015, and references therein). This bracketing estimate was devised by Benn and Ballantyne (2005) based on chironomid data from Whitrig Bog (~170 km to the southeast; Brooks and Birks, 2000, 2001) and Abernethy Forest (~45 km to the northeast; Brooks et al., 2012). A second temperature value of 6.38 °C was also employed, which was derived from modelling experiments by Golledge et al. (2008). This value was suggested as an alternative that accounts for the effects of localised glacier cooling of air temperatures, unlike the chironomid-inferred temperature (C-IT) value that is derived from ice-free areas (cf. Boston et al., 2015, and references therein).

To generate a mean summer (ablation season) temperature from the C-IT value of 8.5 ± 0.3°C, as required by both *P/T* functions, a relationship proposed by Benn and Ballantyne (2005) was used:

$$T_a = 0.97T\text{j} \qquad (3)$$

where T_a is the mean summer temperature and T_j is the mean July temperature (°C) at the ELA. The relationship is based on analysis of meteorological data from Scotland and Scandinavia, and it is assumed that the current Scandinavian and Scottish summers are suitable analogues for Younger Dryas climate in Scotland. In the absence of other suitable data, the equation is deemed to be a reasonable approximation and followed in Scottish palaeoglaciological studies (e.g. Lukas and Bradwell, 2010; Boston et al., 2015). Using the C-IT value of 8.5 ± 0.3°C and Eq. 3, and assuming typical environmental lapse rates of 0.006–0.007 °C m^{-1}, the mean summer temperature at the ELA of the Gaick Icefield is 3.5 ± 1.0 °C. An alternative temperature function has been advocated by Golledge et al. (2010) that incorporates the effect of local glacier cooling, but this has not been used as (i) it would preclude direct comparison with previously-published work from Scotland and (ii) it is not possible to account for the effect of glacier cooling on air temperature for the Gaick Icefield as a whole (cf. Boston et al., 2015; Chandler and Lukas, 2017).

Following procedures outlined by Boston et al. (2015), and applying the palaeotemperature value of 3.5 ± 1.0 °C to the *P/T* functions, yielded palaeoprecipitation estimates of 1795 ± 539 mm a^{-1} using the Ohmura et al. (1992) equation, and 1763 ± 657, 1260 ± 527 and 1008 ± 461 mm a^{-1} for the summer-, neutral- and winter-types, respectively, using the Golledge et al. (2010) equation (Table 5). These precipitation values represent a range of climatic regimes, highlighting the effect seasonality could have upon glacier mass balance (cf. Golledge et al., 2010; Finlayson et al., 2011).

To allow comparison of the palaeoprecipitation values derived from the Gaick Icefield with those from other ice masses in Scotland, it is necessary to calculate sea-level equivalent precipitation totals. Since precipitation is known to increase non-linearly with altitude (e.g. Ballantyne, 1983), the corresponding precipitation at different altitudes can be calculated by

$$P_{Z1} = P_{Z2} / (1 + P^*)^{0.01(Z2 - Z1)} \qquad (4)$$

where P_{Z1} and P_{Z2} represent the precipitation at altitudes Z1 and Z2, and P^* is the proportional increase in precipitation per 100 m increase in elevation (Benn and Ballantyne, 2005). Employing data from Ben Nevis, Ballantyne (2002) demonstrated that $P^* = 0.0578$, and this is used here. The relationship yields sea-level equivalent precipitation estimates of 1177 ± 333 mm a^{-1} using the Ohmura et al. (1992) equation, and 1156 ± 410, 826 ± 331 and 661 ± 292 mm a^{-1} using the Golledge et al. (2010) function with summer-, neutral- and winter-type precipitation seasonality, respectively (Table 5).

6. Discussion

6.1. Icefield dimensions and their significance

The Younger Dryas plateau icefield reconstructed in this study only covers an area of ~42 km^2 and thus differs markedly from the prevailing paradigm of extensive glaciation in the Gaick during this period, as is depicted on numerous compilation maps of Younger Dryas ice extent in Britain (e.g. Golledge, 2010; McDougall, 2013; Lukas et al., 2017; Bickerdike et al., 2018a, b). We contest the reconstruction of Sissons (1974) based on two independent, but corroborating, lines of evidence. Firstly, contrasts in the morphostratigraphic relationships evident across the area (west vs. east) and/or within valleys (upvalley vs. downvalley) are indicative of distinct glacial phases. The sediment-landform signature in the upper parts of the western valleys is very similar to areas inside Younger Dryas limits elsewhere in Scotland, whereas the sediment-landform assemblages found in other parts of the Gaick are akin to associations found outside Younger Dryas limits (cf. Lukas, 2006; Boston et al., 2015, and references therein). Secondly, our glacierisation threshold analysis offers convincing, and independent, support for restricted Younger Dryas glaciation. It indicates that the eastern Gaick was most likely incapable of supporting the Younger Dryas icefield envisaged by Sissons (1974) due to these areas lying well above the projected regional firnline/ELA. Based on our investigations, the earlier reconstruction appears to overestimate ice coverage during the Younger Dryas by ~85%.

Our findings also refute suggestions that (a) the glacial sediment-landform assemblages in the Gaick relate to the southwestwards retreat of the last British-Irish Ice Sheet and (b) the area nourished (at most) small cirque or niche glaciers during the Younger Dryas (cf. Merritt, 2004; Merritt et al., 2004a, b). With respect to (a), the pattern of ice-marginal landforms in the upper parts of the western Gaick catchments clearly indicates upvalley retreat of locally-sourced glaciers towards the plateau, and we found no geomorphological evidence for the subsequent retreat of ice southwestwards across the plateau. In terms of (b), the sediment-landform signature found in the upper portions of the western Gaick valleys is consistent with that observed in radiometrically-dated Younger Dryas landsystems elsewhere in Scotland (e.g. Lukas, 2006; Lukas and Bradwell, 2010; Finlayson et al., 2011). Our interpretation agrees with the situation in the adjacent Drumochter Hills, where the ice-sheet retreat hypothesis has similarly been found to be incompatible with the available evidence (cf. Benn and Ballantyne, 2005).

In a broader context, the reconstruction of a spatially-restricted plateau icefield in the western Gaick highlights the need for a re-evaluation of Younger Dryas glaciation further to the east (Fig. 1). Substantial valley glaciers and an icefield have been reconstructed in the Eastern Grampians (cf. Sissons, 1972; Sissons and Grant, 1972; Sissons and Sutherland, 1976). However, in light of our revised model of spatially-restricted Younger Dryas glaciation in the Gaick, and the presence of a wealth of pre-Younger Dryas glacial landforms, there is the possibility that some or most of the glacial landforms in the Eastern Grampians may pre-date the Younger Dryas.

Aside from implications for Younger Dryas glacier reconstructions elsewhere, our reconstruction of a spatially-restricted Gaick Icefield demonstrates the need for renewed numerical modelling of glacier-climate interactions during this time, as the 'optimum fit' model of Younger Dryas glaciation in Scotland (Golledge et al., 2008) considerably overestimates ice coverage in the Gaick. Importantly, this may indicate a steeper eastwards reduction in precipitation across Scotland (80% in the 'optimum fit' model) than previously envisaged or, alternatively, the gradient may have been non-linear. Local topoclimatic conditions (e.g. the presence of plateaux) may have introduced further complexity to the location of potential source areas. Moreover, scavenging effects of the West Highlands Glacier Complex and Drumochter Icefield may also have been influential, i.e. precipitation may have been dominantly captured by these ice masses, leading to more arid conditions on the leeward (Gaick) side. Given that there have also been recent re-evaluations of Younger Dryas glaciation in several other areas of Scotland (e.g. Finlayson et al., 2011; Pearce, 2014; Boston et al., 2015), revised numerical simulations of Younger Dryas glacier extent are now warranted.

6.2. Regional palaeoclimatic inferences

Taken together with new or recently-revised ELA estimates for other localities in Scotland, our palaeoclimatic data for the Gaick allow an examination of regional palaeoclimate during the Younger Dryas. The palaeoclimatic variables for the Gaick Icefield are compatible with a general eastward rise in ELA and a concomitant decline in precipitation across Scotland (Table 6): the Gaick Icefield ELA value (751 ± 46 m) is higher than that for the Drumochter Icefield (626 m; Benn and Ballantyne, 2005) but lower than the regional Cairngorms ELA value (918 m; Standell, 2014). Of note is the steep northeastwards increase in ELAs and thus reduction in effective precipitation (18–22%) between the immediately adjacent Drumochter and Gaick icefields, which corresponds well with the hypothetical ELA gradients (Section 5.3). It has been inferred that Scotland was predominantly affected by strong southwesterly and westerly winds during the Younger Dryas (e.g. Sissons, 1979a, b; Ballantyne, 1989, 2002; Christiansen, 2004; Benn and Ballantyne, 2005), and the steep gradient between Drumochter and the Gaick seems to support this notion. Under a scenario of prevailing winds from the (south)west, storm tracks would have passed directly over the Western Highlands and Drumochter

Hills before reaching the Gaick. Thus, precipitation may have been preferentially captured by the windward mountains and ice masses therein, resulting in more arid conditions in the Gaick.

In a broader, Scotland-wide context, it has previously been suggested that steep eastward and northward precipitation gradients were present during the Younger Dryas (e.g. Sissons, 1979b; Ballantyne, 2002; Golledge et al., 2008). The availability of new empirical palaeoclimatic data from the Gaick, together with recalculated precipitation estimates from other regions across Scotland (Table 6), now permits a re-evaluation of these inferred patterns. We use the neutral-type precipitation values for our analysis, in view of uncertainties surrounding the Ohmura et al. (1992) equation (see Section 5.5.2) and seasonality bias during the Younger Dryas (see discussion below).

Plotting of the neutral-type precipitation values from the Golledge et al. (2010) equation against latitude reveals that, where the entire dataset is used, there is no statistically significant relationship[1] between precipitation and latitude (r^2 = 0.1798, p = 0.090; Fig. 11a). Visually, however, the covariance plot suggests that there is a cluster of points that may systematically vary with latitude. This cluster principally includes islands along the west coast of Scotland (sites 1–6, see Fig. 11a and Table 6). Linear regression analysis of this subset and the Orkney Islands reveals that, for the Scottish islands, there is a statistically significant relationship between precipitation and latitude (r^2 = 0.7537, p = 0.011), in accordance with previous analyses (Ballantyne, 2007a, b). However, including data from the mainland sites along the western seaboard (Sutherland: Lukas and Bradwell, 2010; Ben More Coigach: Chandler and Lukas, 2017) does not result in a statistically significant fit. Thus, it appears that latitudinal variations (i.e. a northward increase) in precipitation only prevails for the Scottish islands, perhaps reflecting their maritime settings.

With respect to precipitation and longitude, there is qualitatively a strong negative correlation between these two variables (Figs. 11b). However, the r^2 value (0.3696, p = 0.010) is relatively low, and there is a notable anomalous value (Fig. 11b, site 16). Topographic controls on snow supply were argued to have been critical in determining the location of glaciers in that area (the Orkney Islands) by Ballantyne et al. (2007). Similarly, Chandler and Lukas (2017) established that snowblow played an important role in augmenting mass balance of cirque glaciers at Ben More Coigach (Site 7). Given that the ELAs in these two areas are likely a function of local topoclimatic factors, rather than regional climate trends, the relationship between precipitation and longitude was re-examined with the aforementioned sites excluded (Fig. 11c). This analysis reveals a very highly statistically significant relationship between precipitation and longitude (r^2 = 0.7733, p < 0.001). Thus, the empirical precipitation data implies a dominant eastward decline in precipitation during the Younger Dryas, in

[1] Following common convention, $p < 0.05$ is taken to indicate a statistically significant fit between the regression line and the data.

accordance with earlier inferences (e.g. Sissons, 1979b; Benn and Ballantyne, 2005). Based on the regression equation in Fig. 11c, the eastward decline in the neutral-type precipitation (Golledge et al., 2010) can be crudely estimated as ~195 mm per 100 km.

From the above analyses, it appears that precipitation in Scotland primarily varied with longitude during the Younger Dryas, but with a less dominant latitudinal variation. Plotting the palaeoprecipitation data relative to their geographic coordinates shows the dominant trend of eastward-declining precipitation, but with a general increase in precipitation northwards along the western coast of Scotland (Figs. 11d and e). Moreover, multivariate regression analysis indicates a very highly statistically significant relationship between precipitation and geographic location ($r^2 = 0.7944$, $p < 0.001$). Based on this strong geographic control, we have used geostatistical interpolation (splines) to generate a crude isopleth map for annual precipitation patterns during the Younger Dryas from the empirical data points (Fig. 12). This shows high precipitation totals (1800–2400 mm a^{-1}) along the west coast, with considerably more arid conditions in the Central and Eastern Grampians (400–1000 mm a^{-1}).

The significance of the observed palaeoprecipitation pattern across Scotland is twofold. Firstly, the high glacier-derived precipitation totals along the western seaboard of Scotland accord with numerical modelling that suggests strong zonal circulation and enhanced cyclonic activity during the Younger Dryas, particularly during winter (Isarin et al., 1998; Renssen et al., 2001). Indeed, the simulations predicted that the mid-latitude storm track strengthened and shifted to western Europe, with maximum activity at 57°N, i.e. approximating to the latitudinal range of Scotland. Secondly, the overriding eastwards precipitation gradient (see Figs. 11 and 12) implies that Scotland was primarily influenced by strong westerlies during the Younger Dryas, which focused precipitation along the west coast of Scotland and led to a steep W–E precipitation gradient (cf. Sissons, 1979b; Ballantyne, 2007a, b). This agrees with proxy evidence for strong westerlies in the wider North Atlantic region (e.g. Bakke et al., 2009; Lane et al., 2013).

The west–east trend is particularly significant and has important implications for elucidating seasonality bias at the Younger Dryas maximum. Proxy and modelling evidence suggests that the development of winter sea ice in the North Atlantic during the first part of the Younger Dryas diverted storm tracks south of the sea-ice margin and towards continental Europe (cf. Isarin et al., 1998; Isarin and Renssen, 1999; Brauer et al., 2008; Bakke et al., 2009; Golledge et al., 2010; Palmer et al., 2012). Conversely, increased temperatures during the latter part of the stadial caused the break-up of sea ice and thus allowed moisture-bearing westerlies to move northwards across Scotland and southern Norway (Bakke et al., 2009). These changes in sea-ice extent imply winter precipitation was suppressed during the early part of the stadial, owing to the extensive sea ice cover. However, the increased frequency of storm tracks passing over Scotland in the latter part of the Younger Dryas would have resulted in enhanced precipitation, possibly during both the summer and winter (Palmer et al., 2012). Considering that the glacier-derived precipitation indicates the dominant influence of moisture-bearing westerlies, this *may* indicate that Younger Dryas glaciers in Scotland reached their maxima during the latter part of the stadial when moisture-bearing winds penetrated northwards more frequently, which is consistent with varve chronologies from Scotland (MacLeod et al., 2011; Palmer et al., 2012). The alternative scenario of a Younger Dryas maximum in the early part of the stadial (cf. Bromley et al., 2014, 2016, 2018) would imply that glacier mass balance and thus geometry was dominantly influenced by summer precipitation, i.e. a summer precipitation bias (cf. Golledge et al., 2010).

There are, however, two caveats to the foregoing discussion of precipitation patterns across Scotland during the Younger Dryas: Firstly, the analysis does not account for localised cooling at the glacier surface, which can be significant for large glaciers (cf. Khodakov, 1975; Boston et al., 2015); and, secondly, it is assumed that all ice masses across Scotland reached their maxima synchronously, but the maxima most likely occurred diachronously. Thus, these two factors introduce uncertainty into the precipitation patterns. With respect to the *specific* precipitation totals, further complexity is introduced by (i) the time period from which the temperature value used for the calculations is taken (i.e. early, mid, or late stadial; cf. Lukas and Bradwell, 2010) and (ii) the influence of seasonality (see above). These issues are interlinked with ongoing debate surrounding the timing of the Younger Dryas maximum, which remains unresolved (cf. Golledge et al., 2008; Palmer et al., 2010, 2012; MacLeod et al., 2011; Ballantyne, 2012; Bromley et al., 2014, 2016, 2018; Small and Fabel, 2016a, b).

Although the above analysis is presented based on the use of neutral-type precipitation values, uncertainties regarding the appropriateness of the specific precipitation values shown in Table 6 (i.e. the influence of seasonality bias) are not a significant factor in elucidating the *general* pattern: the steep W–E gradient would be apparent irrespective of the precipitation value used, since the glacier-derived palaeoprecipitation estimates are fundamentally a function of the ELAs. Indeed, regression analyses with the different precipitation types yield negligible differences (r^2 ± 0.01, or lower; Table 7). Nevertheless, further constraints on (i) the effect of glacier cooling and (ii) the timing of maximum glaciation are required before palaeoprecipitation can confidently be estimated.

6.3. Comparison with modern precipitation

Within the context of the caveats outlined above, this section provides a first approximation of differences in the Younger Dryas and present-day precipitation regimes in the Gaick. For the purposes of this analysis, and to facilitate comparison, the neutral sea level-equivalent value is used (826 ± 331 mm a^{-1}). This precipitation value is considered an acceptable first estimate given uncertainties regarding an early Younger Dryas maximum (= enhanced effective precipitation; i.e. summer-type required) and localised glacier surface cooling (= reduced effective precipitation).

Modern precipitation data were obtained for two sites in the vicinity of the Gaick Icefield limits, namely Allt Dubhaig (NN 642 726; 416 m OD) and the Tromie Dam (NN 762 881; 436 m OD). These modern data were used to generate average precipitation totals for a twenty-year period where continuous data existed (1992–2011 inclusive) and adjusted to sea level using the procedure outlined in Section 5.5.2. The twenty-year average annual precipitation totals were 1313 ± 240 mm a^{-1} and 1143 ± 144 mm a^{-1} for Allt Dubhaig and the Tromie Dam, respectively (British Atmospheric Data Centre). Taking the average of these modern precipitation totals (1228 ± 192 mm a^{-1}), the data suggest that modern precipitation totals are 402 mm a^{-1} (49%) higher than the Younger Dryas precipitation value. Thus, the data imply that the Gaick was considerably more arid than the present-day, in accordance with suggestions by Golledge et al. (2010). Similarly, more arid conditions are implied for the adjacent Drumochter, based on comparison of the Allt Dubhaig total and the neutral-type precipitation estimate from the Drumochter Icefield (1054 ± 359 mm a^{-1}; Table 6). However, the difference is noticeably reduced, with modern precipitation being only 259 mm a^{-1} (~25%) higher than the Younger Dryas value. Thus, the precipitation data imply that the Drumochter Icefield exerted an influence on the Gaick Icefield through precipitation scavenging. Nevertheless, it is reaffirmed here that the differences in the precipitation totals should only be considered as a first approximation given uncertainties relating to (i) the influence of local glacier surface cooling, (ii) seasonality bias, and (iii) the timing of maximum glaciation.

7. Conclusions

The main findings of this research are as follows:

- We synthesised the results of geomorphological mapping in the Gaick and recognised the presence of three different sediment-landform assemblages in the area, which are suggested to relate to three separate phases of glaciation.
- Using morphostratigraphic principles, we argued that a distinct sediment-landform signature only found in the upper parts of the western Gaick catchments relates to a phase of spatially-restricted plateau icefield glaciation that occurred during the Younger Dryas (Loch Lomond Stadial; GS-1; ~12.9–11.7 ka). This finding refutes the accepted paradigm of extensive Younger Dryas glaciation across the whole Gaick, as proposed by Sissons (1974).

- On the basis of the morphostratigraphic relationships evident in the Gaick, and the assignment of only landforms in the upper parts of the western valleys to the Younger Dryas, it is inferred that most of the glacial landforms in the area pre-date the Younger Dryas. We argue that these landforms relate to (a) deglaciation of the last British-Irish Ice Sheet and (b) a subsequent phase of local glaciation at the end of Greenland Stadial 2.
- We tested our interpretation of the sediment-landform associations using a novel approach to examine 'plateau glacierisation thresholds' (see Supplementary Material), which we developed based on seminal work by Manley (1955). Using a hypothetical gradient derived from empirically-based ELA estimates from elsewhere, this GIS analysis suggested that only the western Gaick would have been able to support ice masses during the Younger Dryas, which is an independent corroboration of the geomorphological evidence.
- Minimum, mid-range and maximum three-dimensional reconstructions of the Younger Dryas icefield were produced using a combination of geomorphological evidence and two 2D glacier surface profile models, following the procedure devised by Boston et al. (2015) for palaeo-plateau icefield settings. This integrated approach indicated that the Gaick Icefield covered an area of ~42 km^2 at that time.
- ELAs were reconstructed using the AAR and AABR methods, and they were computed using an ArcGIS ELA toolbox (Pellitero et al., 2015). Using an AABR of 1.91 ± 0.81 yielded an ELA value of 751 ± 46 m for the Gaick Icefield, which takes into account uncertainties associated with the thickness of ice on the plateau.
- Assuming a mean July sea-level temperature of 8.5 ± 0.3 °C, palaeoprecipitation at sea level is estimated to have been 1177 ± 333 mm a^{-1} using the Ohmura et al. (1992) equation, and 1156 ± 410, 826 ± 331 and 661 ± 292 mm a^{-1} using the Golledge et al. (2010) function with summer-, neutral- and winter-type precipitation seasonality, respectively. There remains debate regarding seasonality bias and the most appropriate precipitation function, but we consider the neutral-type precipitation value (826 ± 331 mm a^{-1}) to be the most appropriate estimate. Comparison with modern precipitation data indicates considerably lower precipitation during the Younger Dryas than at present.
- (Geo)statistical analyses of all published glacier-derived precipitation totals for the Younger Dryas in Scotland confirmed the presence of a prevailing west-east precipitation gradient at this time, whilst a south-north rise in precipitation was found for the Scottish Islands.
- Our study provides important empirical glaciological and climatic data that could be used as input for modelling Younger Dryas glacier-climate interactions, as well as for testing models used for predictive modelling of future climate change.

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Supplementary Material

S1. Method for calculating and displaying heights above the hypothetical equilibrium-line altitude.

S2. Glacier surface profile modelling procedure and input data.

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Figure captions

Fig. 1. Map showing the current understanding of Younger Dryas glacier extent in areas surrounding the Gaick, based on data in Sissons and Sutherland (1976), Benn and Ballantyne (2005), Finlayson (2006), Golledge (2010), Standell (2014) and Boston et al. (2015). Also shown (by the red dashed line) is the Gaick icefield reconstruction from Sissons (1974). The extent of ice-dammed lakes in the Lochaber area is from Turner et al. (2014). Note, the West Highlands Glacier Complex is generalised (no nunataks are shown), as no empirical reconstruction has been produced for the entire ice mass. Sites providing chronological constraint that are discussed in this paper: 1 = Loch Etteridge (Walker, 1975a); 2 = Pass of Drumochter (Walker, 1975a); and 3 = Glen Geusachan (Everest and Kubik, 2006; Standell, 2014). (b) Principal topography of the Gaick, Central Grampians, Scotland. The locations of geomorphological map extracts shown in Fig. 2 are indicated by the white frames. Scale and orientation are provided by British National Grid (10 km intervals). Elevation data is from Ordnance Survey contours. © Crown Copyright Ordnance Survey, an EDINA Digimap/JISC supplied service. Underlying hillshade relief model in (b) was derived from the NEXTMap Britain™ dataset (Intermap Technologies Inc.).

Fig. 2. Glacial geomorphological map extracts for (a) Glen Edendon (NN 714 722) and adjoining areas, (b) southern Gaick Pass (NN 731 820), (c) Glen Tarf (NN 944 798) and adjoining areas, (d) Glas Choire (NN 737 765) and adjacent areas, (e) Glen Bruar (NN 822 779), and (f) Glen Geldie (NN 938 869) and neighbouring areas. The map extracts are taken from the glacial geomorphological map presented by Chandler et al. (2018a). For locations, please see Fig. 1.

Fig. 3. Extract from the glacial geomorphological map presented by Chandler et al. (2018a) showing the Younger Dryas glacier limits (red lines) identified in the Gaick.

Fig. 4. Predicted areas above the hypothetical ELA for a range of scenarios and the relative heights of these areas above the ELA, produced by manipulating the NEXTMap Britain™ DSM. Underlying hillshade relief model derived from the NEXTMap Britain™ dataset (Intermap Technologies Inc.).

Fig. 5. Topographic profiles across the Gaick illustrating the northeastwards increase in the projected ELA and concomitant decrease in areas lying above the ELA. For description of the four scenarios, please see the text.

Fig. 6. Plots of the Gaick summits against the inverse relationship between summit breadth and height above equilibrium line altitude (ELA) required for snow accumulation proposed by Manley (1955). The 'uncertainty envelope' (grey fill) beneath the curve highlights where modern plateau icefields have plotted underneath the 'Manley curve' (see Rea et al., 1998). The inset map shows the locations of the sectors used for the analysis, which were broadly defined based on the present-day watersheds (see also Fig. 1, for location map). Note, any summits situated at the southern limit of the northern sector were taken as being in the northern sector. For description of the four scenarios, please see the text.

Fig. 7. The main flowlines and hypothetical ice divides used for surface profile modelling in this study. The ice divides are based on the positions of the present-day watersheds. Numbered flowlines correspond to the numbers and letters in Table 3. The underlying hillshade model was derived from the NEXTMap Britain™ dataset (Intermap Technologies Inc.).

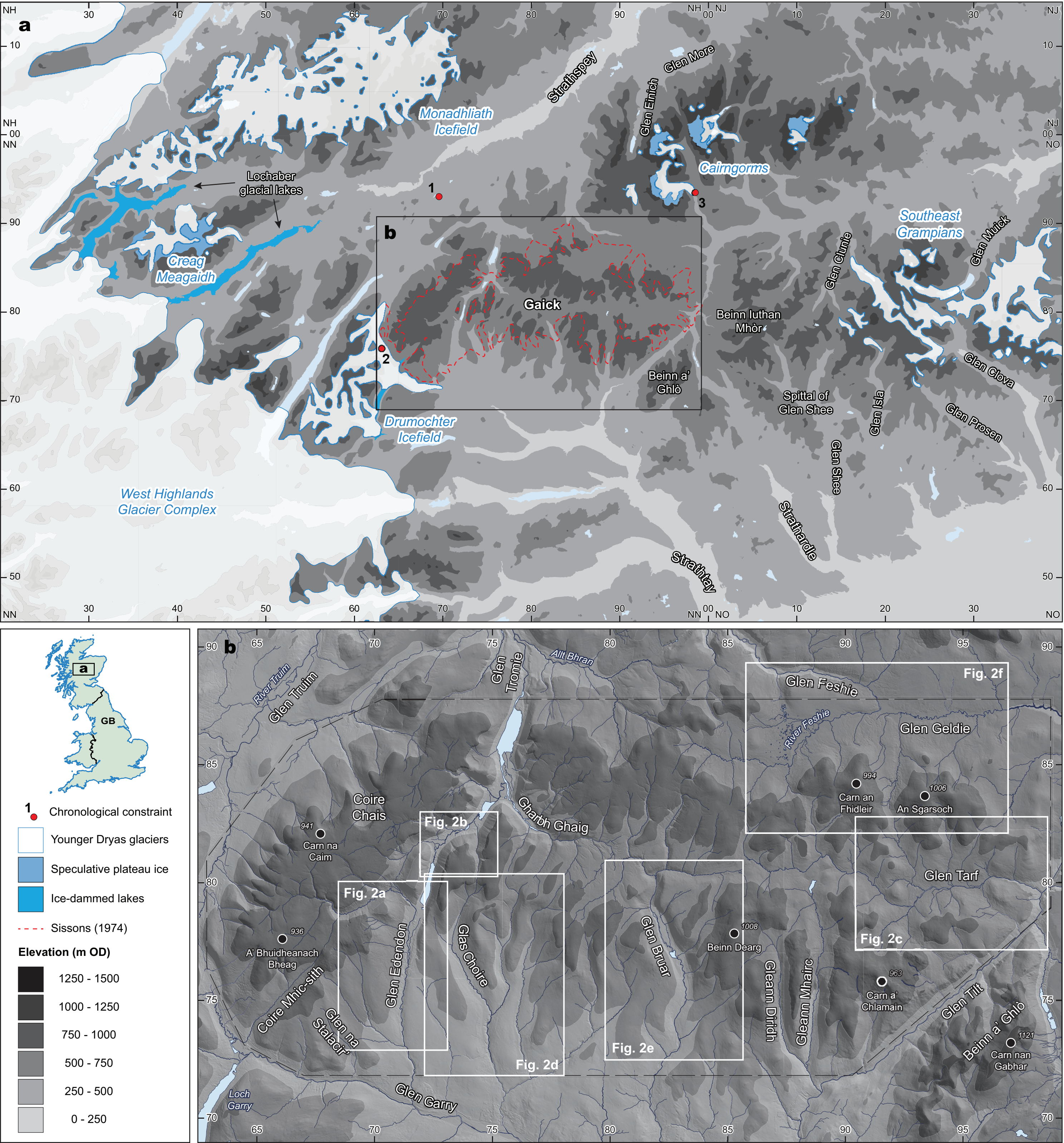
Fig. 8. Conceptual diagram illustrating how the modelled ice surface elevations from either side of an ice divide (i.e. neighbouring catchments) were assimilated to define an 'envelope of likelihood' for the ice surface elevation. This follows the approach devised by Boston et al. (2015) for palaeo-plateau icefield settings. Note, the Ng et al. (2010) modelling approach was only employed to estimate ice surface elevation at the col.

Fig. 9. The reconstructed 'mid-range' Younger Dryas Gaick Icefield. Part of the Drumochter icefield (modified from Benn and Ballantyne, 2005) has been included for context. There remains some uncertainty over the Glas Glacier (and its connection with the Edendon lobe) due to the low ELA value calculated for this glacier (see text). Elevation data was derived from the NEXTMap Britain™ DSM and overlain on a hillshade model, also derived from NEXTMap (Intermap Technologies Inc.).

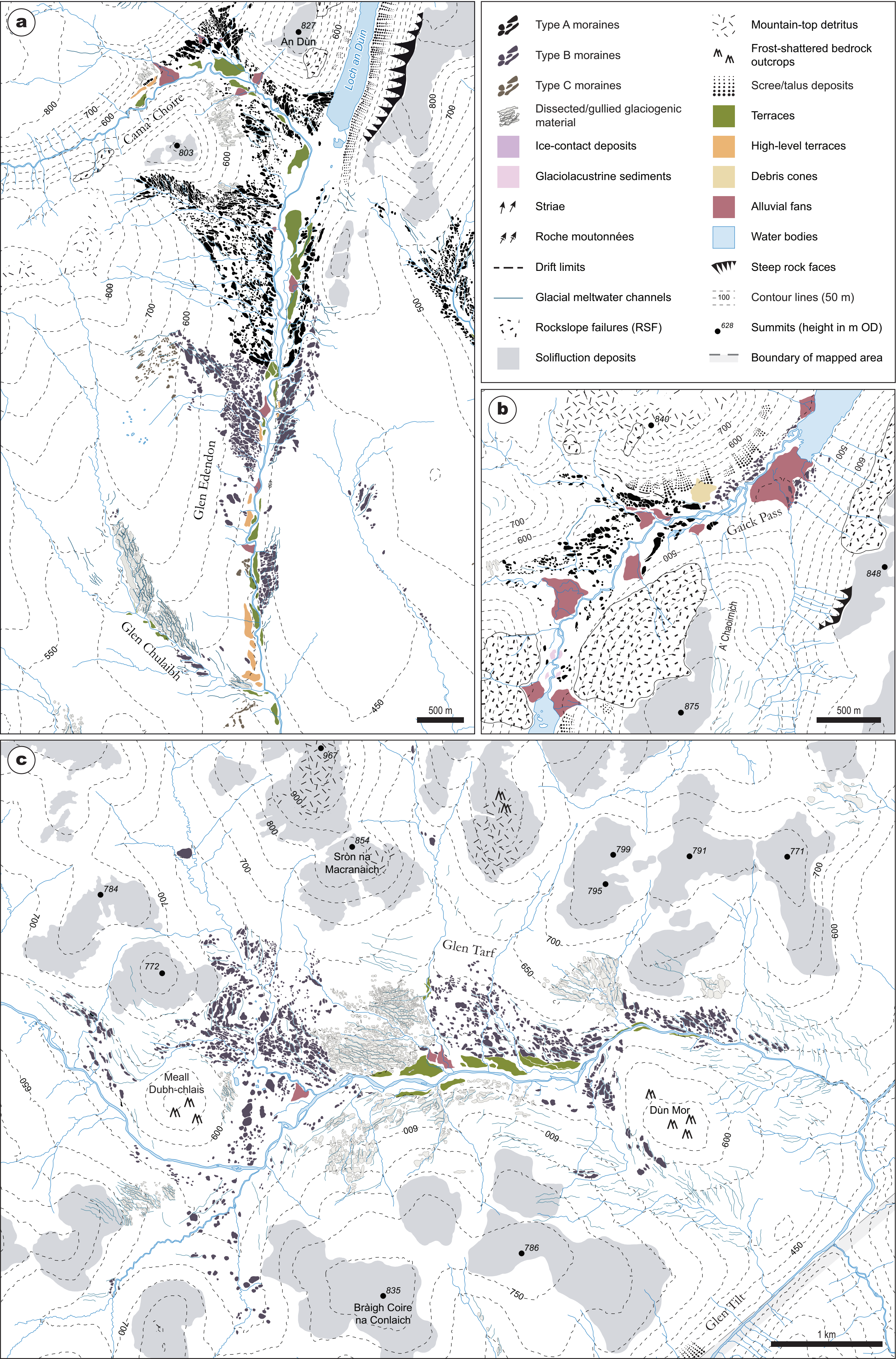
Fig. 10. Minimum (a), mid-range (b) and maximum (c) reconstructions of the Gaick Younger Dryas plateau icefield, with a comparison of the ice surface contours from all the reconstructions overlain on the maximum reconstruction in (d). Elevation data was derived from the NEXTMap Britain™ DSM and overlain on a hillshade model, also derived from NEXTMap (Intermap Technologies Inc.).

Fig. 11. Covariance plots (a–c) and bubble plots (d–e) comparing glacier-derived palaeoprecipitation estimates for the Younger Dryas in Scotland and geographic location (latitude, longitude). Covariance plots: (a) all available palaeoprecipitation estimates and latitude, with the regression line for the Scottish Islands (SIs) also shown by the green dashed line; (b) all available palaeoprecipitation estimates and longitude, with an anomalous value from Orkney (Ballantyne et al., 2007) circled in red; and (c) palaeoprecipitation and longitude, where data for both Ben More Coigach (7; Chandler and Lukas, 2017) and Orkney (16; Ballantyne et al., 2007) have been excluded. Corrie glaciers in these two localities are likely to have been augmented by topoclimatic factors. Bubble plots in (d) and (e) show the quantity of precipitation (indicated by the width of the circles) relative to geographic location. Data from Ben More Coigach and Orkney are excluded from the plot in (e), as above. The inset map (top right) shows the location of each data point, as recorded in Table 6. For explanation of the statistical data, see the text.

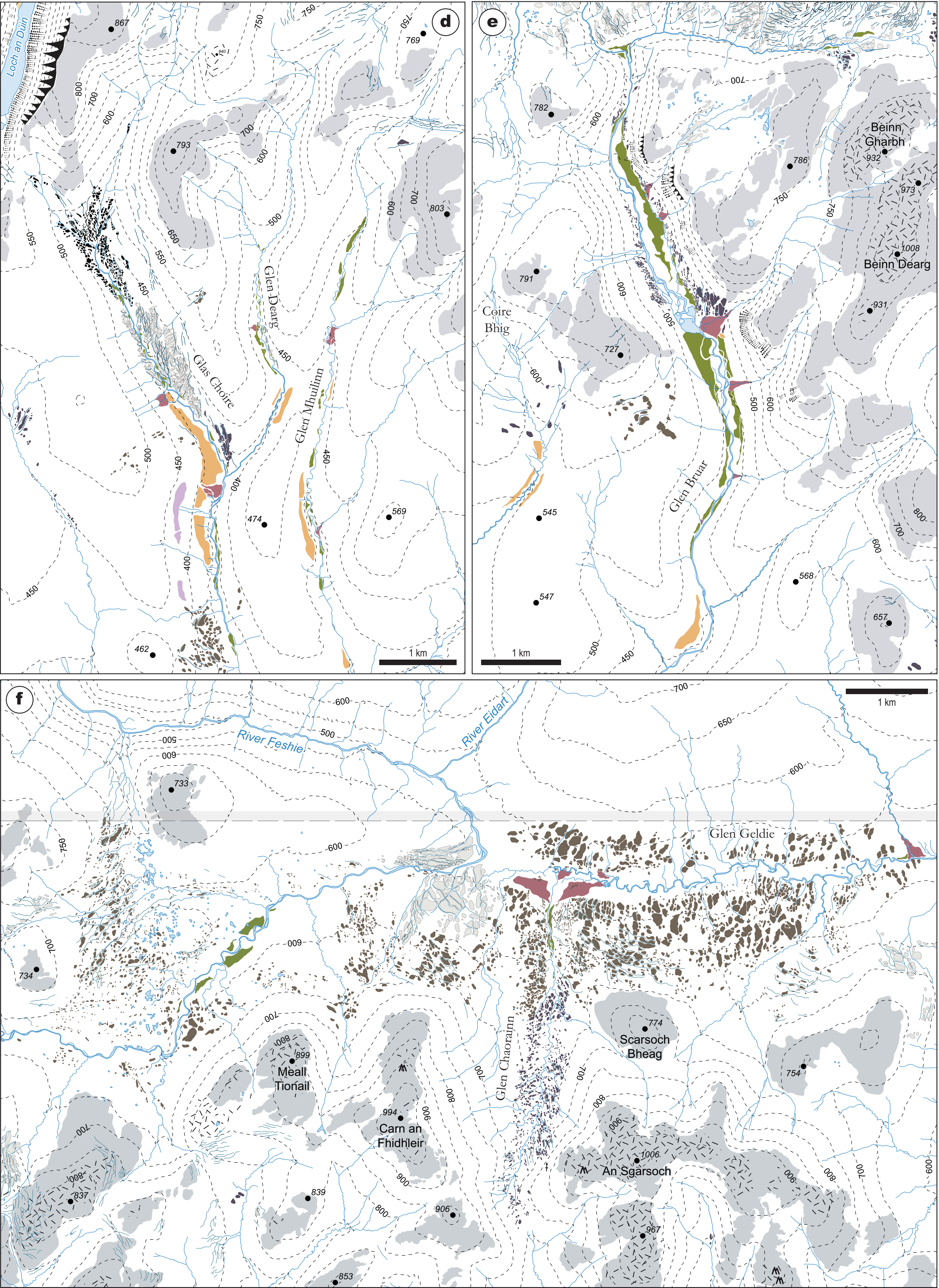
Fig. 12. Isopleth map showing the precipitation pattern across Scotland during the Younger Dryas, based on geostatistical interpolation of available glacier-derived palaeoprecipitation estimates and the neutral-type precipitation regime of Golledge et al. (2010). The splines method was used for geostatistical interpolation due to the irregularly-spaced nature of the data (cf. Akkala et al., 2010). Note, estimates for Ben More Coigach and Orkney have been excluded from the interpolation as they do not conform with the regional precipitation pattern, i.e. glacier mass balance in these localities was likely augmented by topoclimatic factors. The maximum limits of currently recognised Younger Dryas ice masses are shown by white solid lines. Background hillshade model is derived from SRTM data.

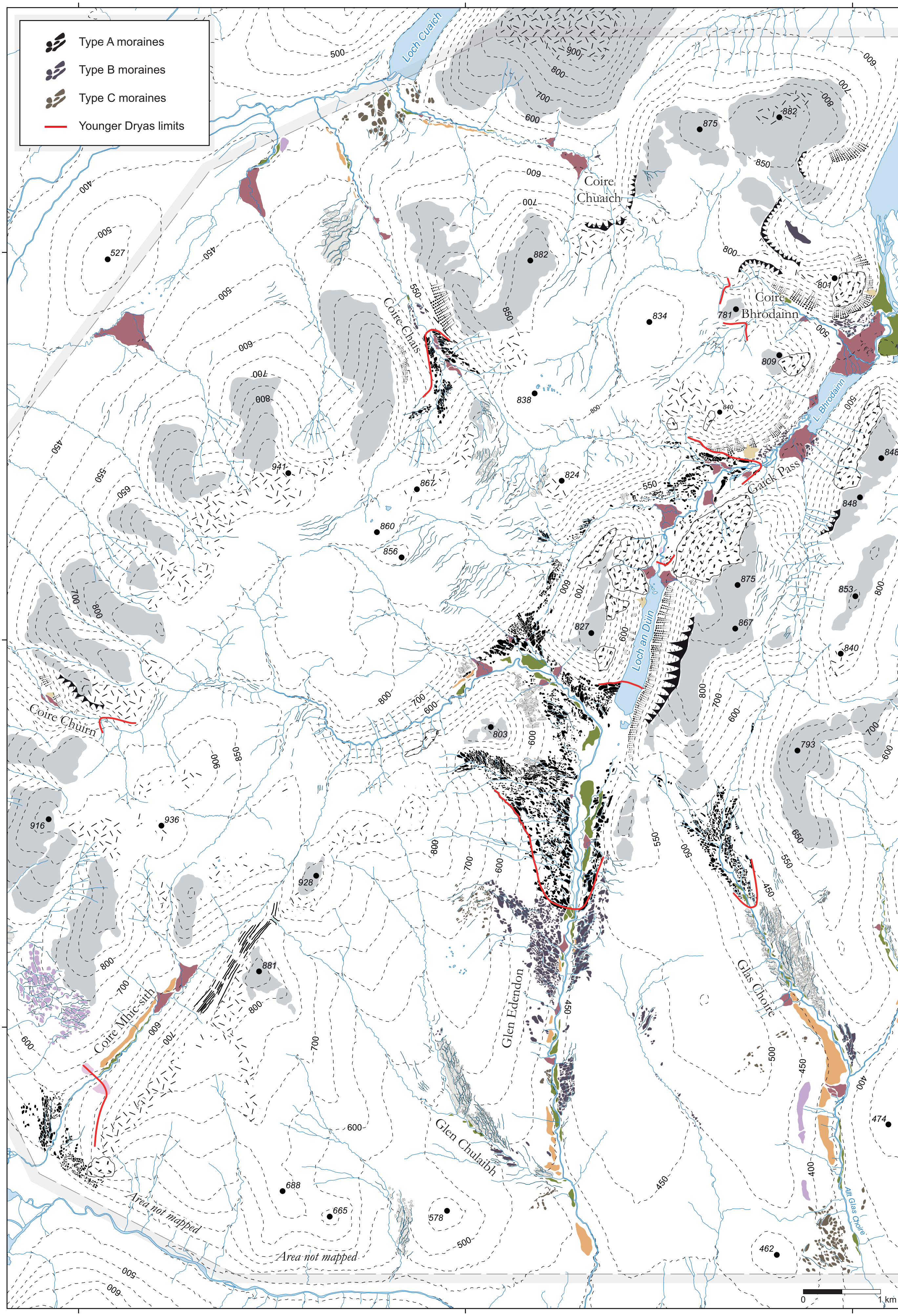


a
Monadhliath Icefield
Strathspey
Glen More
Glen Einich
Cairngorms
Lochaber glacial lakes
Creag Meagaidh
Southeast Grampians
Glen Muick
Glen Clunie
Gaick
Beinn Iuthan Mhòr
Glen Clova
Beinn a' Ghlò
Spittal of Glen Shee
Glen Isla
Glen Prosen
Drumochter Icefield
Glen Shee
West Highlands Glacier Complex
Strathardle
Strathtay
b
River Truim
Glen Truim
Glen Tromie
Allt Bhran
Glen Feshie
River Feshie
Glen Geldie
Carn an Fhidleir
An Sgarsoch
Coire Chais
Carn na Caim
Gharbh Ghaig
Glen Tarf
Glen Edendon
Glas Choire
Glen Bruar
Beinn Dearg
Gleann Diridh
Gleann Mhairc
Carn a' Chlamain
A' Bhuidheanach Bheag
Coire Mhic-sith
Glen na Stalaic
Glen Tilt
Beinn a' Ghlò
Carn nan Gabhar
Glen Garry
Loch Garry
Fig. 2a
Fig. 2b
Fig. 2c
Fig. 2d
Fig. 2e
Fig. 2f
GB
Chronological constraint
Younger Dryas glaciers
Speculative plateau ice
Ice-dammed lakes
Sissons (1974)
Elevation (m OD)
1250 - 1500
1000 - 1250
750 - 1000
500 - 750
250 - 500
0 - 250

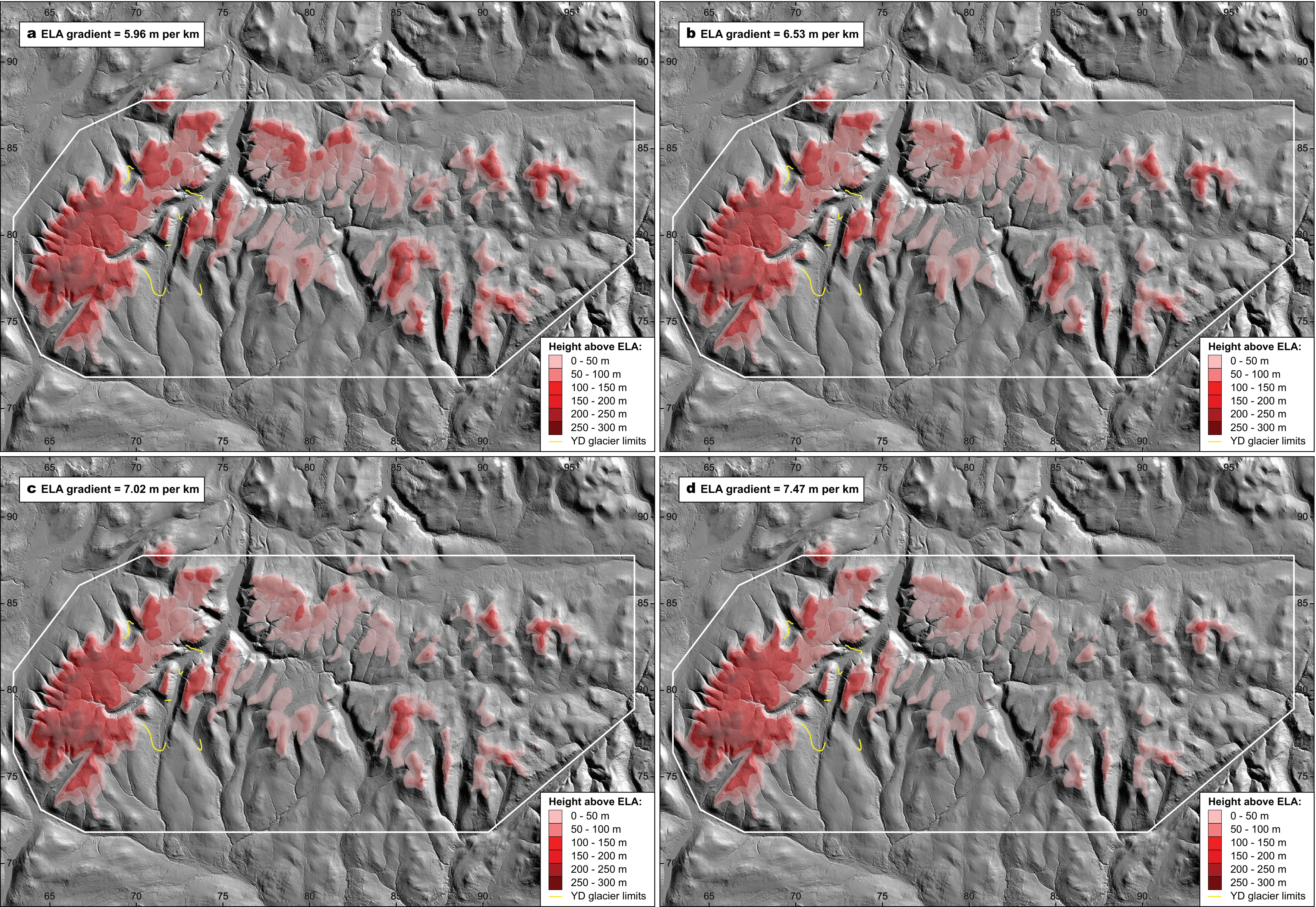


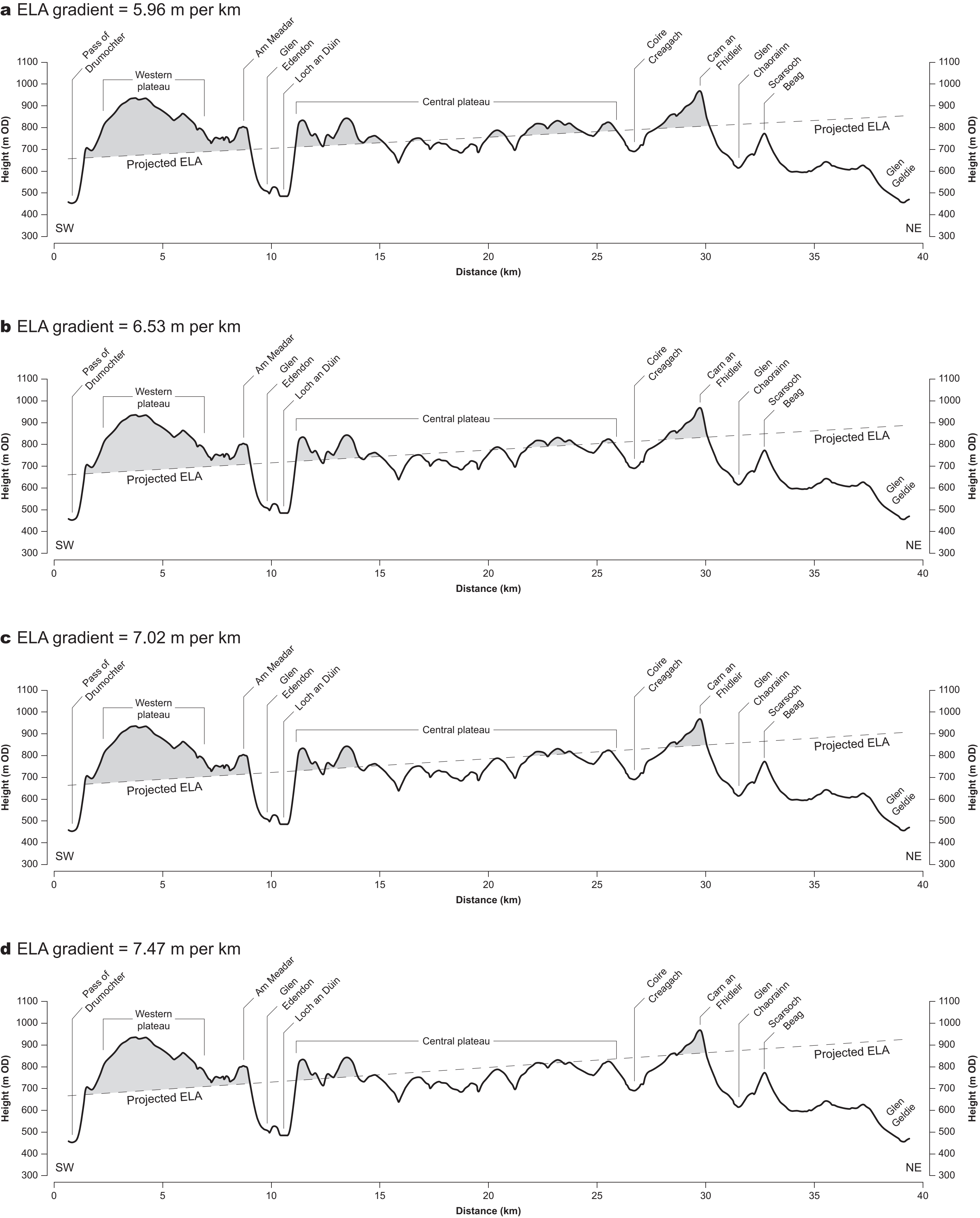
a
b
c
Type A moraines
Type B moraines
Type C moraines
Dissected/gullied glaciogenic material
Ice-contact deposits
Glaciolacustrine sediments
Striae
Roche moutonnées
Drift limits
Glacial meltwater channels
Rockslope failures (RSF)
Solifluction deposits
Mountain-top detritus
Frost-shattered bedrock outcrops
Scree/talus deposits
Terraces
High-level terraces
Debris cones
Alluvial fans
Water bodies
Steep rock faces
Contour lines (50 m)
Summits (height in m OD)
Boundary of mapped area
An Dùn
Loch an Dùin
Cama Choire
Glen Edendon
Glen Chulaibh
Gaick Pass
A' Chaoirnich
Glen Tarf
Sròn na Macranaich
Meall Dubh-chlais
Dùn Mor
Bràigh Coire na Conlaich
Glen Tilt
500 m
1 km


d
Loch an Dùn
867
769
793
803
Glen Dearg
Glas Choire
Glen Mhuilinn
474
569
462
1 km
e
782
Beinn Gharbh
932
786
973
1008
Beinn Dearg
791
Coire Bhig
727
931
Glen Bruar
545
547
568
657
1 km
f
1 km
River Feshie
River Eidart
733
Glen Geldie
734
Glen Chaorainn
Scarsoch Bheag
774
754
899
Meall Tionail
994
Carn an Fhidhleir
1006
An Sgarsoch
837
839
906
967
853

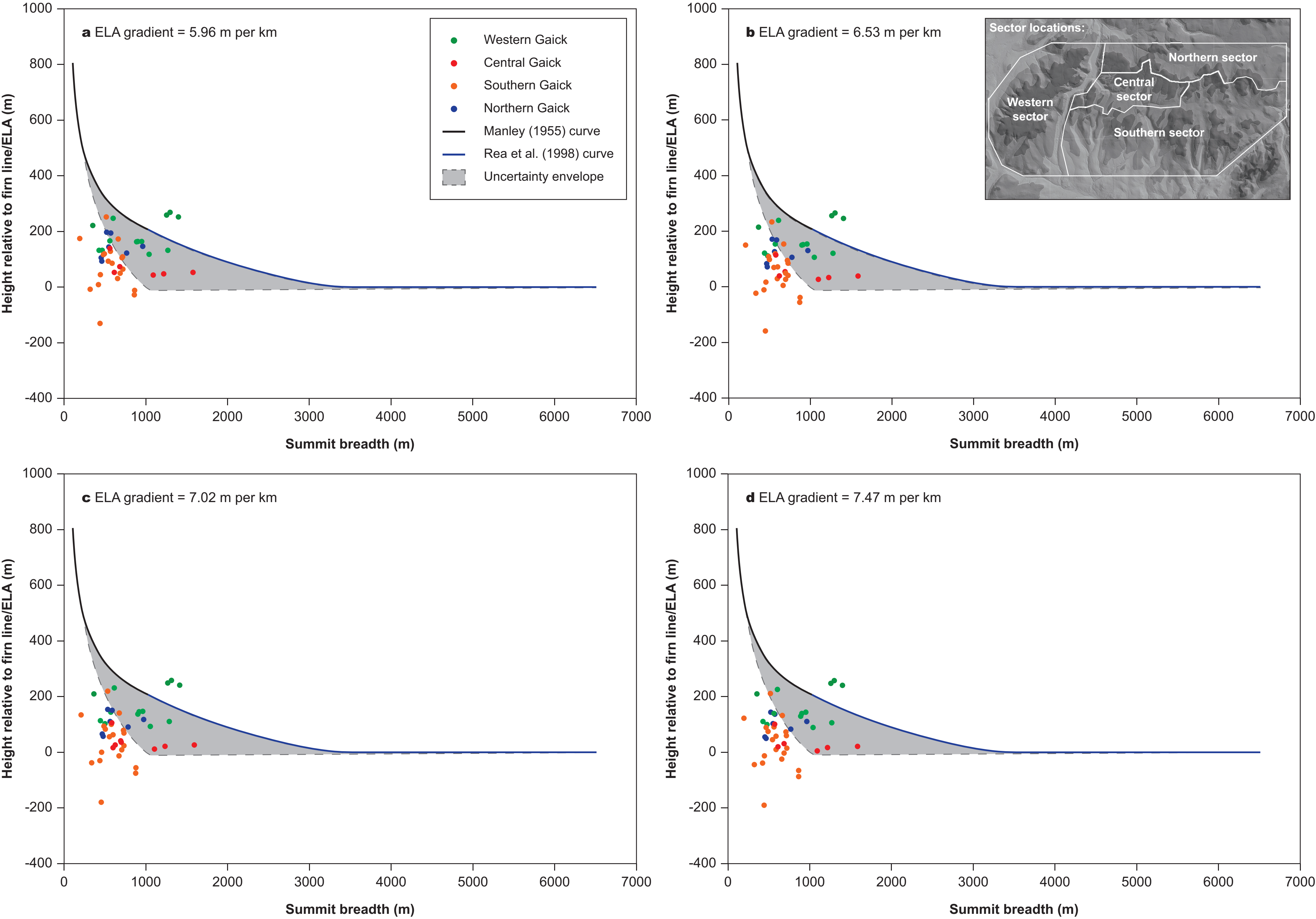


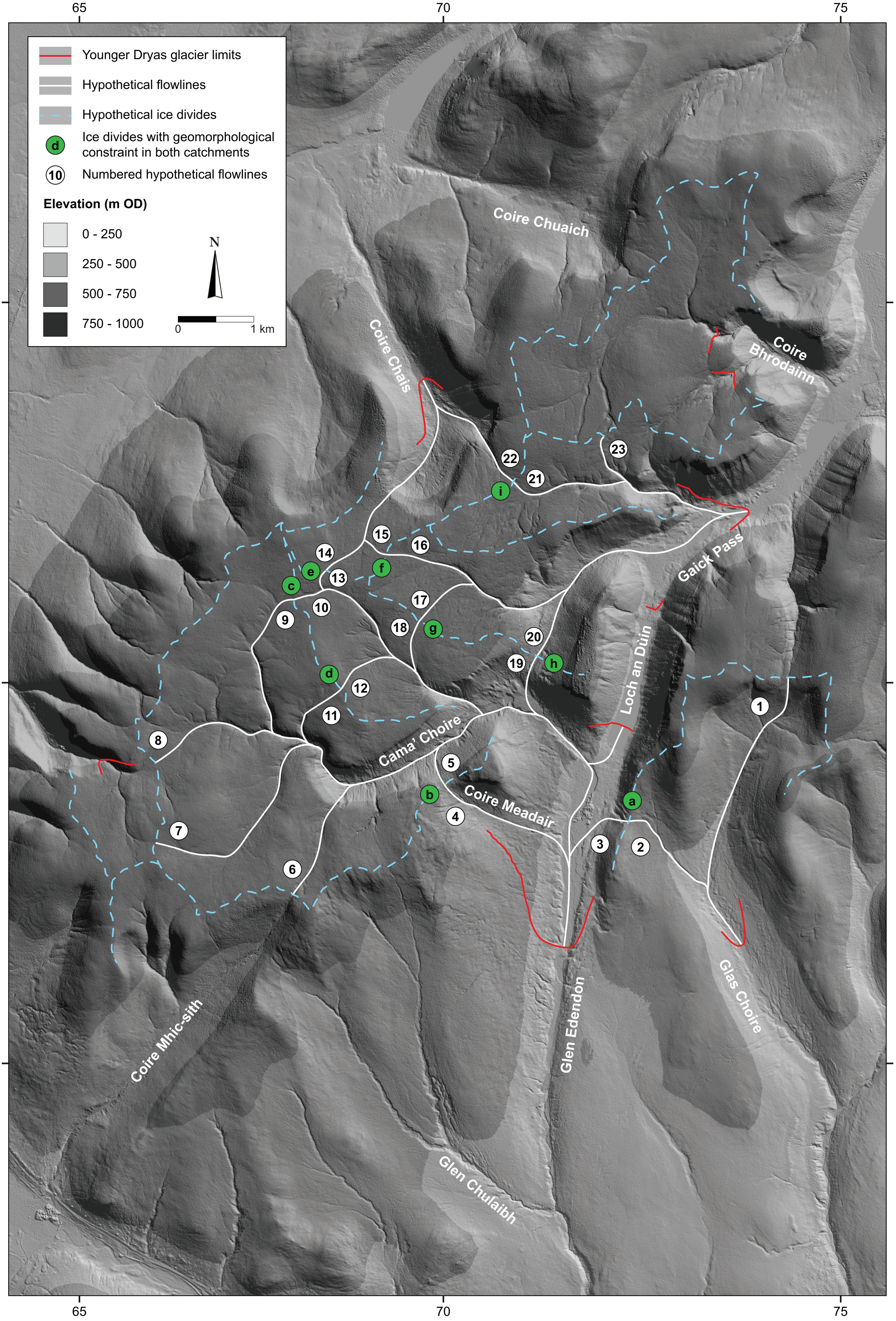
Type A moraines
Type B moraines
Type C moraines
Younger Dryas limits
Loch Cuaich
Coire Chuaich
Coire Bhrodainn
L. Bhrodainn
Coire Chais
Gaick Pass
Loch an Dùin
Coire Chuirn
Coire Mhic-sith
Glen Edendon
Glas Choire
Glen Chulaibh
Allt Glas Choire
Area not mapped
Area not mapped
0
1 km
65
70
75
85
80
75


a ELA gradient = 5.96 m per km
b ELA gradient = 6.53 m per km
c ELA gradient = 7.02 m per km
d ELA gradient = 7.47 m per km
Height above ELA:
0 - 50 m
50 - 100 m
100 - 150 m
150 - 200 m
200 - 250 m
250 - 300 m
YD glacier limits
65
70
75
80
85
90
95

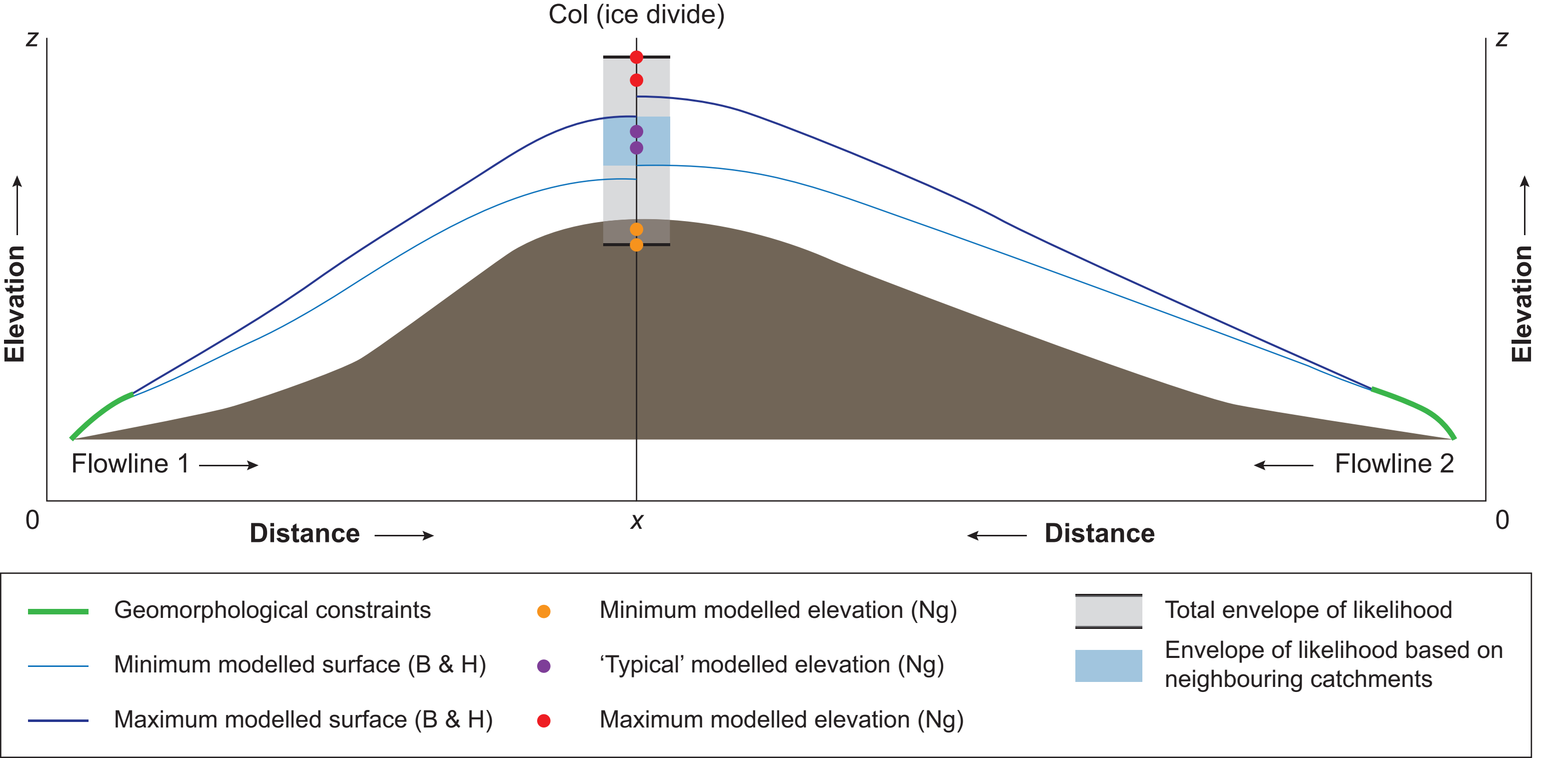


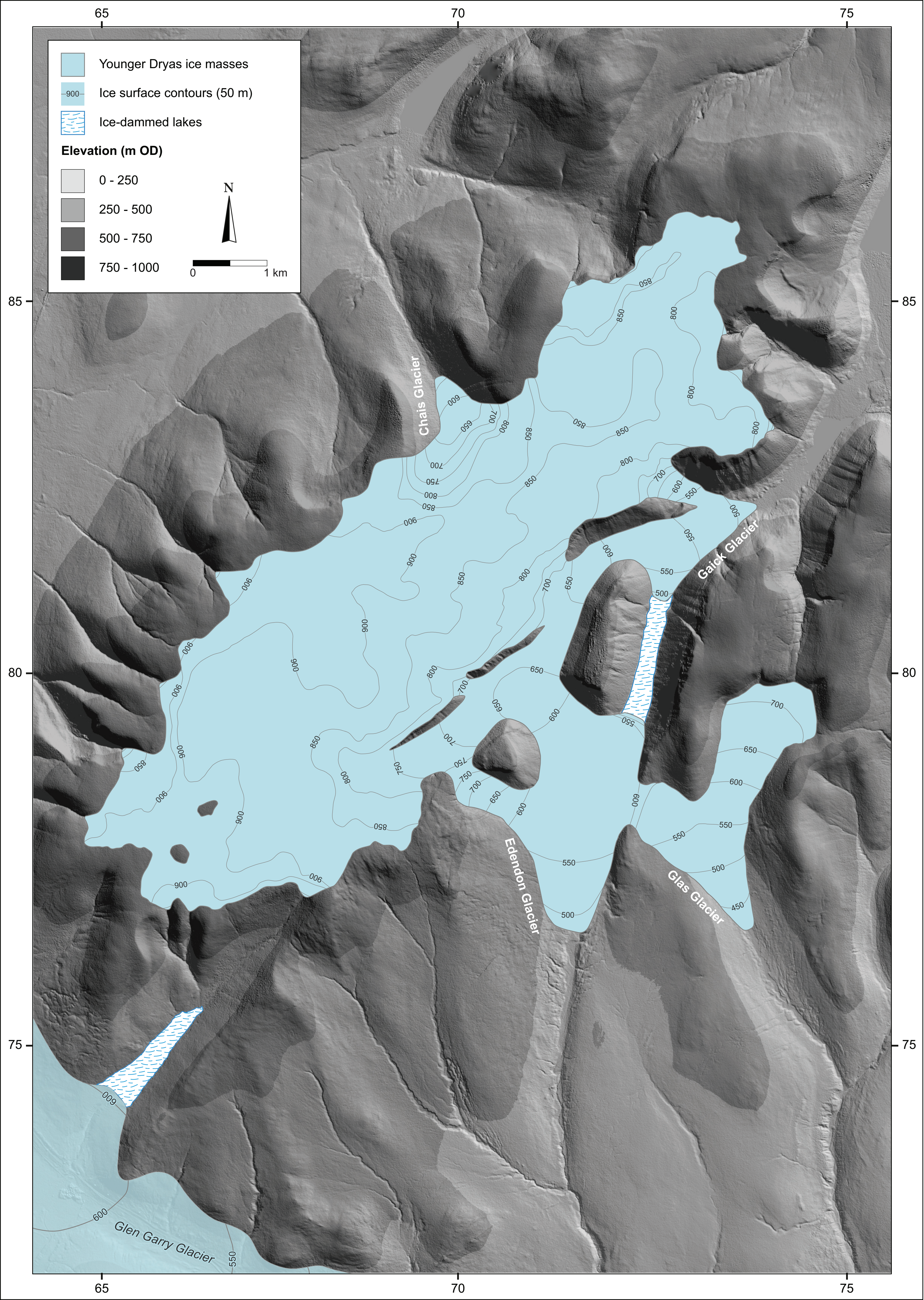
a ELA gradient = 5.96 m per km
b ELA gradient = 6.53 m per km
c ELA gradient = 7.02 m per km
d ELA gradient = 7.47 m per km
Pass of Drumochter
Western plateau
Am Meadar
Glen Edendon
Loch an Dùin
Central plateau
Coire Creagach
Carn an Fhidleir
Glen Chaorainn
Scarsoch Beag
Glen Geldie
Projected ELA
SW
NE
Height (m OD)
Distance (km)


a ELA gradient = 5.96 m per km
b ELA gradient = 6.53 m per km
c ELA gradient = 7.02 m per km
d ELA gradient = 7.47 m per km
Western Gaick
Central Gaick
Southern Gaick
Northern Gaick
Manley (1955) curve
Rea et al. (1998) curve
Uncertainty envelope
Sector locations:
Northern sector
Central sector
Western sector
Southern sector
Height relative to firn line/ELA (m)
Summit breadth (m)
1000
800
600
400
200
0
-200
-400
0
1000
2000
3000
4000
5000
6000
7000

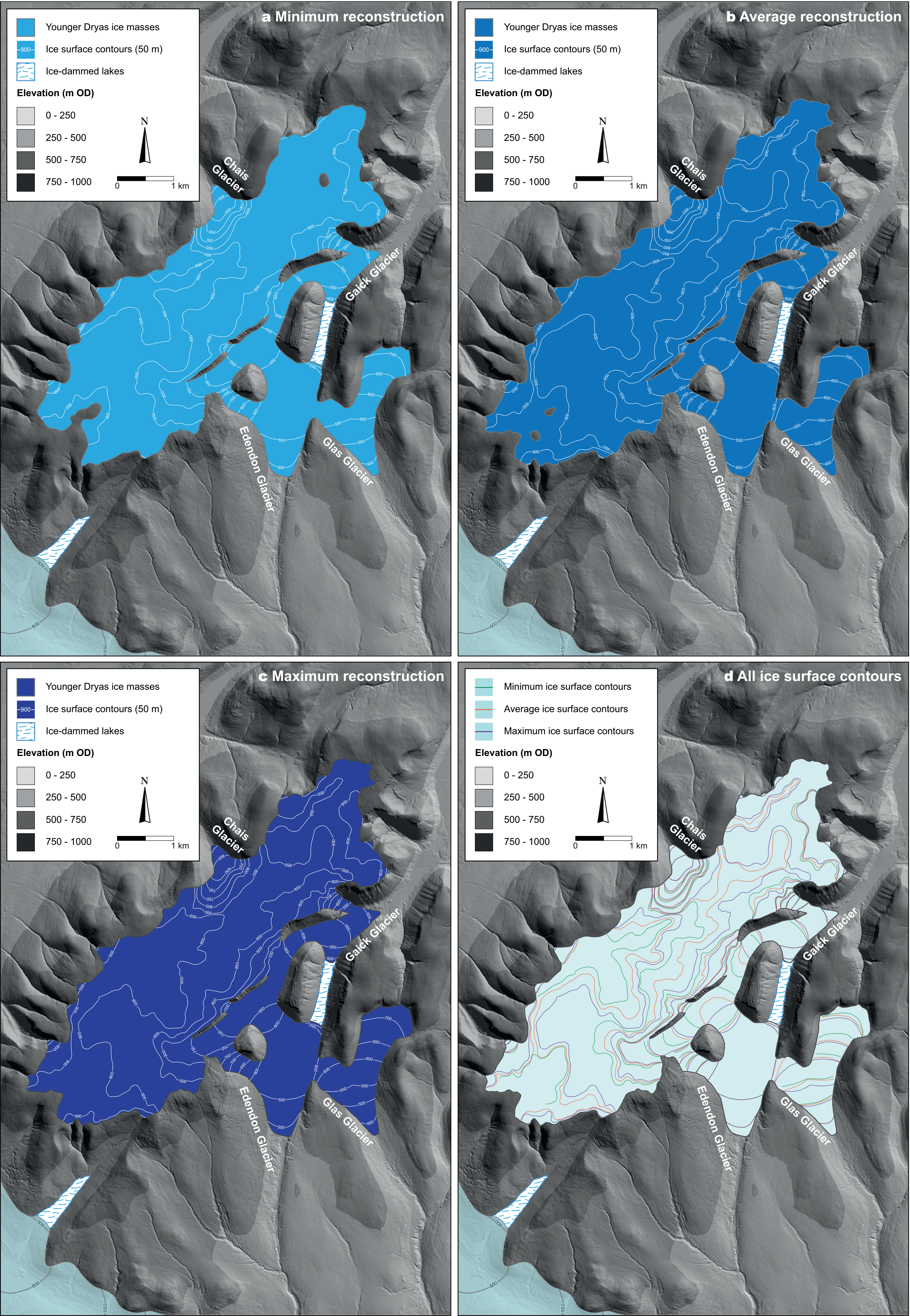


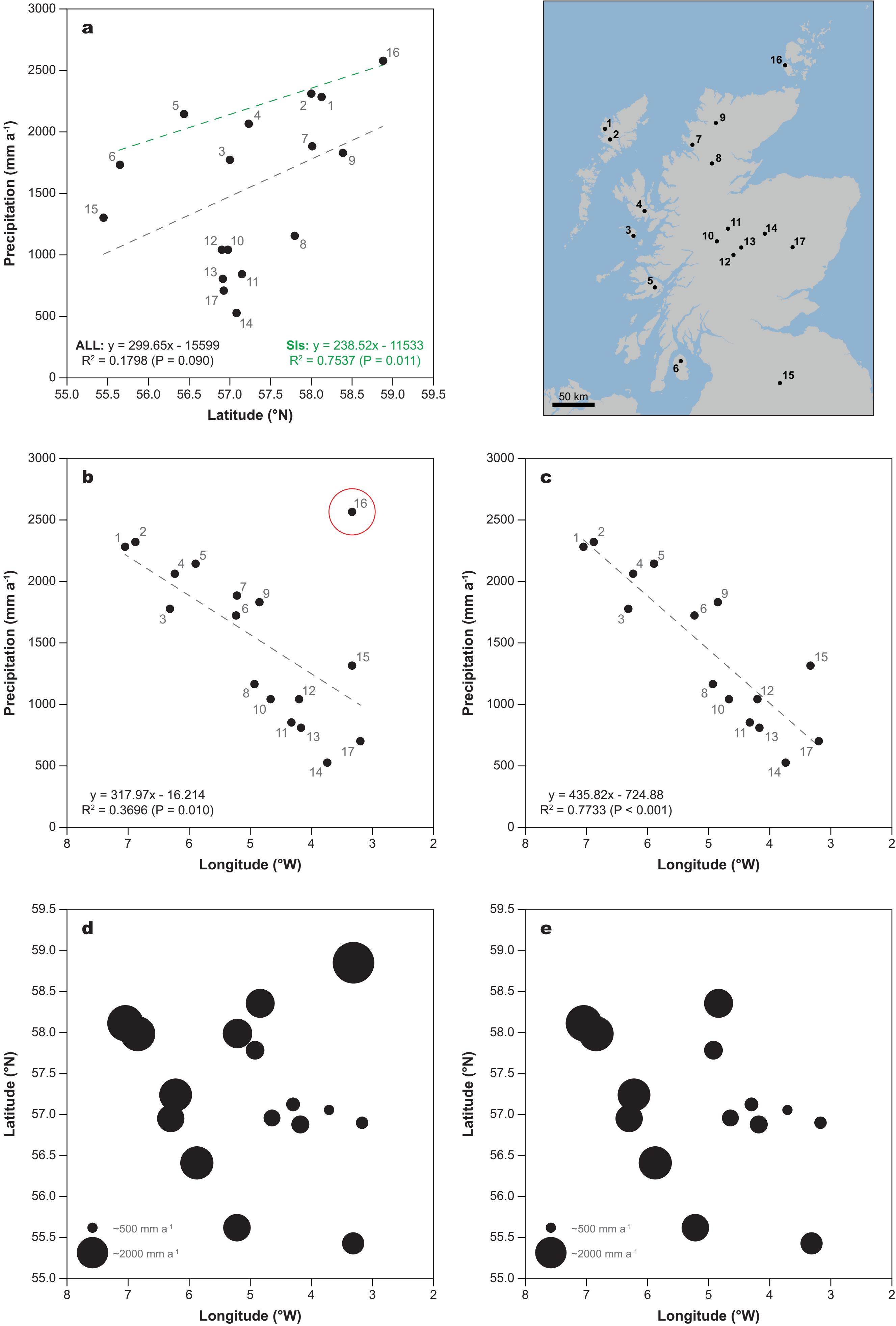
65
70
75
85
80
75
Younger Dryas glacier limits
Hypothetical flowlines
Hypothetical ice divides
Ice divides with geomorphological constraint in both catchments
Numbered hypothetical flowlines
Elevation (m OD)
0 - 250
250 - 500
500 - 750
750 - 1000
N
0
1 km
Coire Chuaich
Coire Chais
Coire Bhrodainn
Gaick Pass
Loch an Dùin
Cama' Choire
Coire Meadair
Glen Edendon
Glas Choire
Coire Mhic-sith
Glen Chulaibh


Col (ice divide)
z
z
Elevation
Elevation
Flowline 1
Flowline 2
0
0
x
Distance
Distance
Geomorphological constraints
Minimum modelled surface (B & H)
Maximum modelled surface (B & H)
Minimum modelled elevation (Ng)
'Typical' modelled elevation (Ng)
Maximum modelled elevation (Ng)
Total envelope of likelihood
Envelope of likelihood based on neighbouring catchments

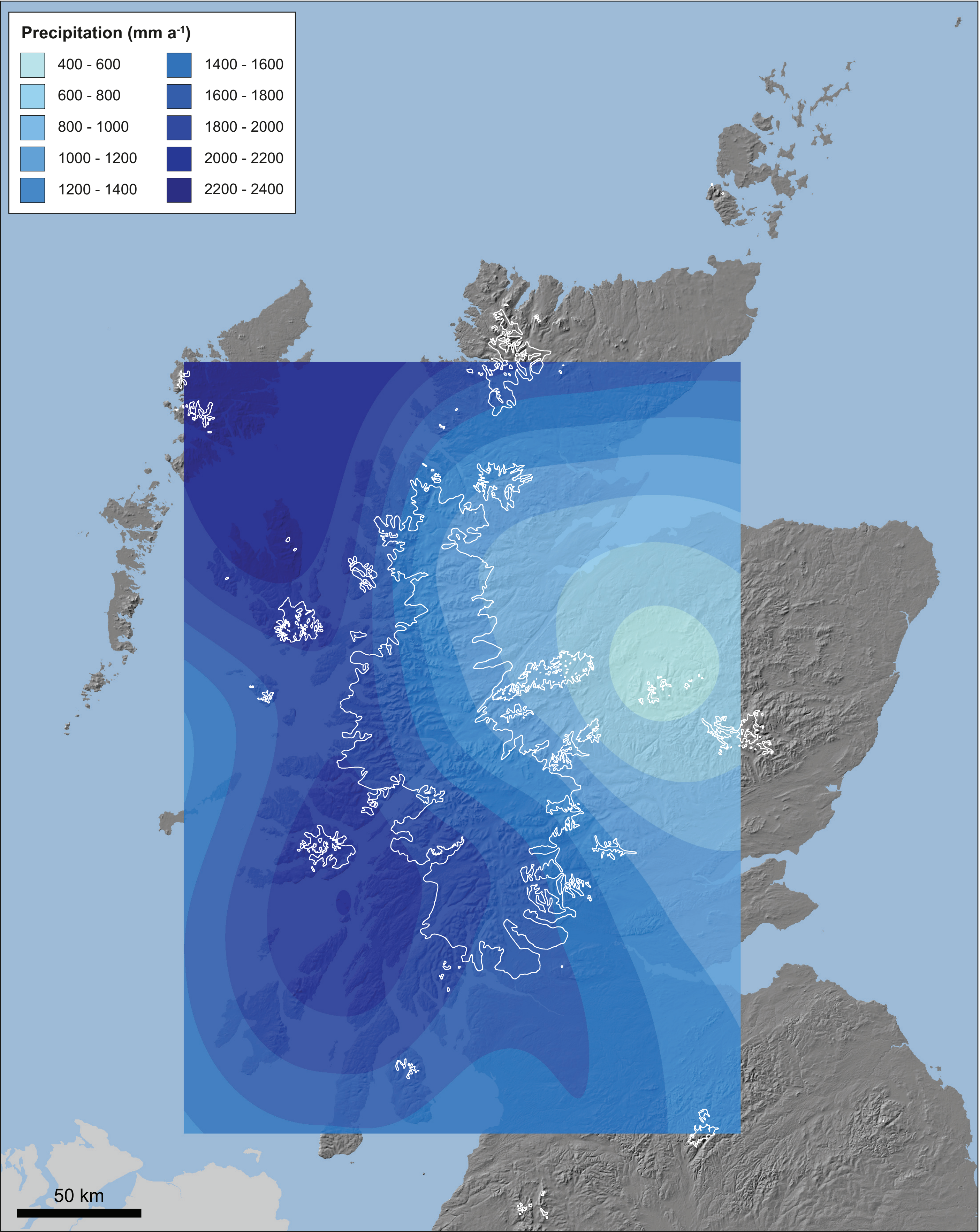


Younger Dryas ice masses
Ice surface contours (50 m)
Ice-dammed lakes
Elevation (m OD)
0 - 250
250 - 500
500 - 750
750 - 1000
N
0
1 km
Chais Glacier
Gaick Glacier
Edendon Glacier
Glas Glacier
Glen Garry Glacier
65
70
75
85
80


a Minimum reconstruction
b Average reconstruction
c Maximum reconstruction
d All ice surface contours
Younger Dryas ice masses
Ice surface contours (50 m)
Ice-dammed lakes
Minimum ice surface contours
Average ice surface contours
Maximum ice surface contours
Elevation (m OD)
0 - 250
250 - 500
500 - 750
750 - 1000
N
0
1 km
Chais Glacier
Gaick Glacier
Edendon Glacier
Glas Glacier



a
ALL: y = 299.65x - 15599
R² = 0.1798 (P = 0.090)
SIs: y = 238.52x - 11533
R² = 0.7537 (P = 0.011)
Precipitation (mm a⁻¹)
Latitude (°N)
50 km
b
y = 317.97x - 16.214
R² = 0.3696 (P = 0.010)
Longitude (°W)
c
y = 435.82x - 724.88
R² = 0.7733 (P < 0.001)
d
e
~500 mm a⁻¹
~2000 mm a⁻¹


Precipitation (mm a-1)
400 - 600
600 - 800
800 - 1000
1000 - 1200
1200 - 1400
1400 - 1600
1600 - 1800
1800 - 2000
2000 - 2200
2200 - 2400
50 km

Tables

Table 1. Summary of criteria used to distinguish the moraine types in the Gaick.

Criteria	Type A moraines	Type B moraines	Type C moraines
Shape / dimensions	~2–15 m high, ~10–85 m wide and 10–250 m long ridges and mounds; may be more pronounced than Type B and C moraines; may display consistent shapes / dimensions within a single assemblage; comparable to 'hummocky moraine' in other areas of Scotland	Wide range of dimensions; no clear difference between these moraines and Type A; morphology may be highly variable within a single assemblage; comparable to 'hummocky moraine' in other areas of Scotland	Wide range of dimensions but often take the form of subdued (<2 m high), broad moraine ridges; frequently manifest as larger moraines than Type A and B
Planimetric pattern	Inset transverse chains that form nested arcuate loops or chevron patterns	Inset transverse chains that form nested arcuate loops or chevron patterns; may be less clearly organised than Type A moraines	Topographically-discordant moraines (i.e. trend in an upvalley direction); cross-cut moraines; incompatible with radial flow from the Gaick
Location in valleys	Occur on valley slopes and trend obliquely downslope towards the valley centre; limited to upvalley parts of valleys	Restricted to a narrow band at the slope foot and do not trend obliquely downslope, OR characterise all parts of the catchment (i.e. Type A moraines are absent)	Occur exclusively in the downvalley parts of valleys and around the periphery of the Gaick
Lateral gradients	Steep average gradients (~15–20 m per 100 m)	Shallow average gradients (~5 m per 100 m)	Variable average gradients
Terrace associations	Single river terrace above the active floodplain; major river terrace may terminate abruptly at outer moraines	Distinct river terrace levels above the current floodplain or a single river terrace that terminates in upvalley parts (where Type A is present)	Distinct river terrace levels above the active floodplain, which includes 'high-level' terraces (up to 30 m above the floodplain)
Slope associations	Sediment-covered slopes inside outer moraines and moraine limit may grade to 'drift limit'; talus/scree is largely absent	Extensive talus/scree may occur on valley slopes and descend to upper moraine surfaces	*No distinguishable features*
Other associations	Meltwater channels occur between individual chains; mutually exclusive from dissected sediment masses	Assemblages of dissected material may intersperse / interrupt moraine suites and sharp boundaries between the two are absent	Assemblages of dissected material may intersperse / interrupt moraine suites and sharp boundaries between the two are absent

Table 2. Summary of the geomorphological evidence used to distinguish and delineate the Younger Dryas glacier limits in the Gaick. Moraine type[1]: A – B indicates that an upvalley transition from Type B moraines to Type A moraines was used to distinguish the Younger Dryas glacier limit.

Site/catchment	Grid reference	Younger Dryas limit	Moraine type [1]	Dissected material	Glaciolacustrine landforms	Glaciofluvial landforms	'Drift' limits	Meltwater channels	Terraces	Talus/ scree	Neighbouring catchments
Western Gaick:											
Coire Mhic-sith	652 743	√	√ A		√	√	√		√		
Glen Stalcair	691 720			√				√			
Glen Chulaibh	708 731			√				√			
Glen Edendon	713 765	√	√ B – A					√	√		
Loch an Dùin S	721 794	√	√ A					√		√	√
Loch an Dùin N	725 809	√	√ A		√					√	√
Gaick Pass	737 822	√	√ B – A				√			√	
Coire Chùirn	652 789	√						√		√	√
Coire Chais	695 839	√	√ B – A					√		√	
Coire Chuaich	709 864		√ B						√		
Coire Bhrodainn	733 845	√						√	√	√	√
Southern Gaick:											
Glen Gallaidh	722 747		√ B								
Glas Choire	736 765	√	√ B – A	√		√		√	√		
Glen Dearg	756 757			√					√		√
Glen Mhuilinn	763 763								√		√
Glen Chireachain	793 736								√		√
Coire Mhòir	794 754		√ B	√					√		√
Coire Bhig	808 755		√ B						√		√
Glen Bruar	827 769		√ B						√	√	√
Gleann Diridh	872 740		√ B						√	√	√
Gleann Mhairc	885 742		√ B						√	√	√
Glen Tarf S	972 827		√ B					√			√
Glen Tarf N	981 797		√ B	√				√	√		√
Central Gaick:											
Gharbh Ghaig	765 825		√ B					√	√	√	
Northern Gaick:											
Glen Bynack	976 846		√ B								√
Glen Geldie	915 876		√ C	√				√	√		√
Glen Chaorainn	922 852		√ B								√
Upper Feshie	915 876		√ C	√				√	√		√
Caochan Dubh	871 886		√ C								√
Am Breac–choire	898 853		√ C								√
Coire Creagach	876 838		√ C					√			√
Coire nan Cisteachan	859 836							√			√
Clais Damh	841 853		√ B								√
Coire Bhran	803 868		√ B						√		√

Table 3. Ice surface elevations (m) established using the two glacier surface profile models described in the text and any geomorphological constraints. The flowlines correspond to the numbered flowlines on Fig. 7. Ice surface elevations that were used to define the envelope of likelihood are highlighted in bold text. For explanation of how the envelope of likelihood was defined, please see text and Fig. 8.

Flowline	Ice mass (ice divide)	Col height (m)	Col grid reference	Benn and Hulton (2010)		Ng et al. (2010)			Envelope of likelihood	
				Minimum	Maximum	Minimum	Typical	Maximum	Minimum	Maximum
1	Glas Choire (Chaoirnich)	752	743 800	**709**	**718**	714	771	841	709 [1]	718 [1]
2	Glas Choire (Edendon)	571	722 781	588	605	NMA	NMA	NMA	592	604
3	Edendon (Glas Choire)	571	722 781	**592**	**604**	NMA	NMA	NMA		
4	Meadair (Cama' Choire)	737	698 786	**755**	**770**	713	770	827	755	770
5	Cama' Choire (Meadair)	737	698 786	737	776	NMA	NMA	NMA		
6	Edendon (Coire Mhic-sith)	860	678 772	**894**	**930**	826	925	970	894	930
7	Edendon (A'Bhuidheanach Bheag	920	660 778	**928**	**950**	874	985	1025	928	950
8	Edendon (Coire Chùirn)	827	660 789	**896**	**920**	811	899	948	896	920
9	Edendon (Mhic an Righ) 1	855	678 810	**906**	943	817	906	956	906 [2]	935 [2]
10	Mhic an Righ (Edendon) 1	855	678 810	901	**935**	771	901	962		
11	Edendon (Mhic an Righ) 2	865	684 799	**914**	932	832	870	941	914 [2]	927 [2]
12	Mhic an Righ (Edendon) 2	865	684 799	898	**927**	856	853	913		
13	Edendon (Coire Chais)	867	682 814	901	**945**	810	913	991	921 [2]	945 [2]
14	Coire Chais (Edendon)	867	682 814	**921**	969	797	876	967		
15	Coire Chais (Gaick Pass)	855	690 816	**914**	959	793	863	932	914	941
16	Gaick Pass (Coire Chais)	855	690 816	892	**941**	829	888	978		
17	Mhic an Righ (Gaick Pass)	817	694 808	834	874	**839**	923	968	839 [2, 3]	896 [2, 3]
18	Gaick Pass (Mhic an Righ	817	694 808	861	888	816	**896**	959		
19	Edendon (Gaick Pass)	590	710 803	635	**667**	NMA	NMA	NMA	655 [2]	667 [2]
20	Gaick Pass (Edendon)	590	710 803	**655**	672	NMA	NMA	NMA		
21	Coire an Dubh-chadha (Coire Chais)	791	707 826	838	**862**	730	806	905	839	862
22	Coire Chais (Coire an Dubh-chadha)	791	707 826	**839**	869	772	798	859		
23	Gaick Pass (Coire Bhrodainn)	805	718 832	**825**	**854**	742	803	888	825	854

NMA: no modern analogue in the Ng et al. (2010) dataset.

[1] Envelope of likelihood is taken from the Benn and Hulton (2010) model, since the watershed is at the plateau rim and there is no clear empirical evidence for plateau ice.

[2] Surface profile models from the Loch an Dùin (S) limit to this col also conform to this envelope of likelihood.

[3] Results from the Ng et al. (2010) model are used to define in order to increase confidence, given the narrow range of values from the Benn and Hulton (2010 model).

Table 4. Equilibrium line altitudes (ELAs) calculated for the Gaick Icefield and the main catchments using the ArcGIS ELA calculation toolbox (Pellitero et al., 2015). Values in bold indicate ELAs used for subsequent palaeoclimatic calculations.

Ice mass	ELA1 AABR = 1.67 (m)	ELA2 AABR = 1.8 (m)	**ELA3 AABR = 1.9 (m)**	ELA4 AABR = 2.0 (m)	ELA5 AAR = 0.5 (m)	ELA6 AAR = 0.6 (m)
Minimum reconstruction:						
Glas glacier	569	567	565 ± 17	563	570	558
Edendon glacier	754	749	746 ± 34	742	851	821
Gaick glacier	725	721	717 ± 31	715	804	783
Chais glacier	783	780	778 ± 23	775	841	814
Bhrodainn glaciers	799	799	799 ± 2	798	803	802
Gaick Icefield	*744*	*739*	*736 ± 31*	*733*	*811*	*800*
Average reconstruction:						
Glas glacier	578	576	574 ± 18	572	588	569
Edendon glacier	763	758	755 ± 34	751	867	832
Gaick glacier	733	728	725 ± 31	722	813	790
Chais glacier	789	786	783 ± 23	781	845	824
Bhrodainn glaciers	821	821	820 ± 6	819	840	826
Gaick Icefield	*758*	*754*	*751 ± 31*	*747*	*843*	*811*
Maximum reconstruction:						
Glas glacier	585	582	581 ± 17	579	595	578
Edendon glacier	769	764	761 ± 35	757	883	846
Gaick glacier	740	736	732 ± 33	729	830	795
Chais glacier	798	795	792 ± 24	790	853	830
Bhrodainn glaciers	825	824	823 ± 7	823	847	837
Gaick Icefield	*764*	*759*	*756 ± 32*	*752*	*850*	*819*
All reconstructions:						
Glas glacier	578 ± 9	576 ± 7	**574 ± 26**	572 ± 9	588 ± 18	569 ± 11
Edendon glacier	763 ± 9	758 ± 9	**755 ± 43**	751 ± 9	867 ± 16	832 ± 14
Gaick glacier	733 ± 8	728 ± 8	**725 ± 40**	722 ± 7	813 ± 17	790 ± 7
Chais glacier	789 ± 9	786 ± 9	**783 ± 33**	781 ± 9	845 ± 8	824 ± 10
Bhrodainn glaciers	821 ± 21	821 ± 21	**820 ± 23**	819 ± 21	840 ± 37	826 ± 24
Gaick Icefield	*758 ± 14*	*754 ± 15*	***751 ± 46***	*747 ± 13*	*843 ± 32*	*811 ± 11*

Table 5. Precipitation values and associated uncertainty for the Gaick Icefield at the average ELA and sea level, calculated using the AABR = 1.9 ± 0.81 ELA. The results of both the Ohmura et al. (1992) and Golledge et al. (2010) relationships are displayed, where S-type = summer-type precipitation, N-type = neutral-type precipitation and W-type = winter-type precipitation. Palaeoprecipitation values are presented for both a mean July temperature (T_j) at sea level of 8.5 ± 0.3ºC (Benn and Ballantyne, 2005) and a summer sea-level temperature of 6.38ºC (Golledge, 2008).

Temp. (ºC)	Effective precipitation at the ELA (mm a^{-1})				Effective precipitation at sea level (mm a^{-1})			
	Ohmura et al.	Golledge et al.			Ohmura et al.	Golledge et al.		
		S-type	N-type	W-type		S-type	N-type	W-type
8.5 ± 0.3	1795 ± 539	1763 ± 657	1260 ± 527	1008 ± 461	1177 ± 333	1156 ± 410	826 ± 331	661 ± 292
6.38	1109 ± 421	864 ± 474	617 ± 396	494 ± 357	727 ± 265	567 ± 305	405 ± 256	324 ± 232

Table 6. Currently available annual precipitation estimates for Younger Dryas sites in Scotland, arranged from west to east. Palaeoprecipitation estimates are based on published ELAs and were recalculated by Boston et al. (2015) using the methods described in Section 5.5. Additionally, palaeoprecipitation values have been recalculated in the present study for Skye (Ballantyne et al. (2016), the Cairngorms (Standell, 2014) and the Tweedsmuir Hills (Pearce, 2014). Note, the existing estimate for the Southeast Grampians has been used in the absence of any other palaeoclimatic data for the area. Bold values indicate estimates used for subsequent analyses.

Site	Ice mass	Reference	ELA (m)	ELA method	Precipitation at sea level (mm a^{-1})			
					Ohmura et al. (1992)	Golledge et al. (2010)		
						S-type	N-type	W-type
1	Uig, Lewis	Ballantyne (2006)	212	AABR (1.8)	2754 ± 349	3216 ± 421	**2297** ± 358	1838 ± 327
2	Harris	Ballantyne (2007a)	204	AABR (1.8)	2786 ± 349	3259 ± 421	**2328** ± 358	1862 ± 326
3	Skye	Ballantyne et al. (2016)	277	AABR (1.8)	2524 ± 323	2907 ± 395	**2076** ± 331	1661 ± 299
4	Rhum	Ballantyne and Wain-Hobson (1980)	357	AWMA	2221 ± 356	2500 ± 426	**1786** ± 363	1429 ± 330
5	Mull	Ballantyne (2002)	250	AABR (1.8)	2605 ± 351	3016 ± 424	**2154** ± 360	1723 ± 328
6	Arran	Ballantyne (2007b)	371	AABR (1.8)	2174 ± 357	2437 ± 428	**1741** ± 363	1393 ± 330
7	Ben More Coigach	Chandler and Lukas (2017)	328	AABR (1.9)	2817 ± 425	2665 ± 417	**1903** ± 346	1523 ± 310
8	Beinn Dearg	Finlayson et al. (2011)	576	AABR (1.8)	1562 ± 360	1639 ± 425	**1171** ± 361	937 ± 329
9	NW Highlands	Lukas and Bradwell (2010)	339	AABR (1.8)	2283 ± 356	2582 ± 427	**1844** ± 362	1475 ± 330
10	Creag Meagaidh	Finlayson (2006)	625	AABR (1.8)	1436 ± 360	1478 ± 423	**1056** ± 360	845 ± 328
11	Monadhliath	Boston et al. (2015)	714	AABR (1.9)	1224 ± 409	1211 ± 480	**865** ± 400	692 ± 360
	- Western		610	AABR (1.9)	1465 ± 423	1515 ± 498	**1082** ± 413	866 ± 370
	- Central		705	AABR (1.9)	1279 ± 422	1280 ± 494	**914** ± 410	731 ± 368
	- Eastern		777	AABR (1.9)	1088 ± 407	1043 ± 471	**747** ± 395	596 ± 354
12	Drumochter	Benn and Ballantyne (2005)	626	AABR (1.8)	1434 ± 360	1475 ± 423	**1054** ± 359	843 ± 328
13	*Gaick*	*This study*	*751*	*AABR (1.9)*	*1177 ± 333*	*1156 ± 410*	***826*** *± 331*	*661 ± 292*
14	Cairngorms [1]	Standell (2014)	918	AABR (1.9)	852 ± 286	760 ± 342	**543** ± 278	434 ± 246
15	Tweedsmuir Hills [2]	Pearce (2014)	527	AABR (1.9)	1725 ± 334	1853 ± 416	**1323** ± 340	1059 ± 302
16	Hoy, Orkney	Ballantyne et al. (2007)	141	AABR (1.8)	3046 ± 344	3615 ± 416	**2582** ± 354	2066 ± 323
17	SE Grampians	Sissons and Sutherland (1976)	790	AWMA	1060 ± 358	1008 ± 413	**720** ± 352	576 ± 322

[1] Based on the valley glacier scenario of Standell (2014) and excluding ‘candidate’ glaciers (potential but uncertain Younger Dryas glaciers).
[2] ELA value for the main Tweedsmuir Icefield and based on the surface profile modelling scenario of Pearce (2014).

Table 7. Summary of the results from regression analyses undertaken to examine the influence of geographic location (latitude and longitude) on precipitation values in Scotland during the Younger Dryas. Only negligible differences are evident between the different precipitation types.

Statistical test	Ohmura et al. (1992)		Golledge et al. (2010) S-type		Golledge et al. (2010) N-type		Golledge et al. (2010) W-type		r^2 range
	r^2 value	p value	r^2 value	p value	r^2 value	p value	r^2 value	p value	
Latitude regression (ALL)	0.1928	0.078	0.1798	0.090	0.1798	0.090	0.1797	0.090	0.013
Latitude regression (Scottish Islands)	0.7529	0.011	0.7534	0.011	0.7537	0.011	0.7541	0.011	0.001
Longitude regression (ALL)	0.3594	0.011	0.3694	0.010	0.3696	0.010	0.3694	0.010	0.010
Longitude regression (excl. Orkney and BMC)	0.7731	<0.001	0.7767	<0.001	0.7733	<0.001	0.7769	<0.001	0.004
Multiple regression (ALL)	0.4670	0.012	0.4662	0.012	0.4663	0.012	0.4661	0.012	0.001
Multiple regression (excl. Orkney and BMC[1])	0.7991	<0.001	0.8017	<0.001	0.7944	<0.001	0.8019	<0.001	0.007

[1] Orkney and Ben More Coigach are excluded from this analysis as they were strongly influenced by topoclimatic factors (see text)