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Palaeoclimatic reconstruction from Lateglacial (Younger Dryas Chronozone) cirque glaciers in Snowdonia, North Wales

Jacob M. Bendle, Neil F. Glasser *

Centre for Glaciology, Institute of Geography and Earth Sciences, Aberystwyth University, Aberystwyth, Ceredigion SY23 3DB, UK

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ABSTRACT

The cirques of Snowdonia, North Wales were last occupied by glacier ice during the Younger Dryas Chronozone (YDC), *c.* 12.9–11.7 ka. New mapping presented here indicates 38 small YDC cirque glaciers formed in Snowdonia, covering a total area of 20.74 km^2. Equilibrium line altitudes (ELAs) for these glaciers, calculated using an area–altitude balance ratio (AABR) approach, ranged from 380 to 837 m asl. A northeastwards rise in YDC ELAs across Snowdonia is consistent with southwesterly snow-bearing winds. Regional palaeoclimate reconstructions indicate that the YDC in North Wales was colder and drier than at present. Palaeotemperature and annual temperature range estimates, derived from published palaeoecological datasets, were used to reconstruct values of annual accumulation and 'winter balance plus summer precipitation' using a degree-day model (DDM) and non-linear regression function, respectively. The DDM acted as the best-estimate for stadial precipitation and yielded values between 2073 and 2687 mm a^{-1} (lapse rate: 0.006 °C m^{-1}) and 1782–2470 mm a^{-1} (lapse rate: 0.007 °C m^{-1}). Accounting for the potential input of windblown and avalanched snow onto former glacier surfaces, accumulation values dropped to between 1791 and 2616 mm a^{-1} (lapse rate: 0.006 °C m^{-1}) and 1473–2390 mm a^{-1} (lapse rate: 0.007 °C m^{-1}). The spatial pattern of stadial accumulation suggests a steep precipitation gradient and provides verification of the northeastwards rise in ELAs. Glaciers nearer the coast of North Wales were most responsive to fluctuations in climate during the YDC, responding to sea-ice enforced continentality during the coldest phases of the stadial and to abrupt warming at the end of the stadial.

1. Introduction

The period of climatic amelioration that terminated the Dimlington Stadial glaciation of Britain (*c.* 26–13 ka; Rose, 1985) was marked by a brief return to near full glacial conditions during the Younger Dryas Chronozone (YDC) of *c.* 12.9–11.7 ka (Rasmussen et al., 2006; Lowe et al., 2008). In upland Britain, the interpretation of former glacier limits and deposits of inferred YDC affinity has formed the basis for reconstructions of palaeoglacier dimensions (e.g. Carr, 2001; Gray, 1982; Sissons, 1974; Sutherland, 1984), from which equilibrium line altitudes (ELAs) and palaeoclimatic variables have been derived (Carr and Coleman, 2007). The most significant outcomes of this research have been: (i) recognition of spatial patterns in former ELAs; and (ii) identification of the likely temperature and annual precipitation regimes that supported regional ice mass development (e.g. Ballantyne, 2002b, 2007a,b; Benn and Ballantyne, 2005). Moreover, empirical palaeoclimatic data are necessary to inform and constrain model simulations that aim to understand the climatic conditions coupled with past glacial activity (e.g. Golledge et al., 2008; Golledge and Hubbard, 2005; Finlayson et al., 2011).

Renewed interest in the extent of YDC glaciation in Scotland has seen the development of a significant body of palaeoclimatic data (Golledge, 2010 and references therein). This data has permitted a more informed understanding of contemporaneous climatic sensitivity, through the analysis of regional atmospheric patterns (e.g. moisture sources and pathways) inferred from ELAs and reconstructed precipitation, and by comparison with data obtained from other palaeoenvironmental proxies (e.g. pollen or coleoptera). Despite several important contributions (Hughes, 2002, 2009), an inventory of equal scope does not yet exist for Wales, and as a consequence, our understanding of glacier–climate interactions during the lattermost phase of glacial activity is incomplete. This paper attempts to address these shortfalls using geomorphological mapping and glacier reconstruction over the high ground of Snowdonia, North Wales. Specifically, the study aims to: (1) establish the geomorphological evidence for the extent of locally

* Corresponding author. Tel.: +44 01970 622785; fax: +44 01970 622659.
E-mail address: nfg@aber.ac.uk (N.F. Glasser).

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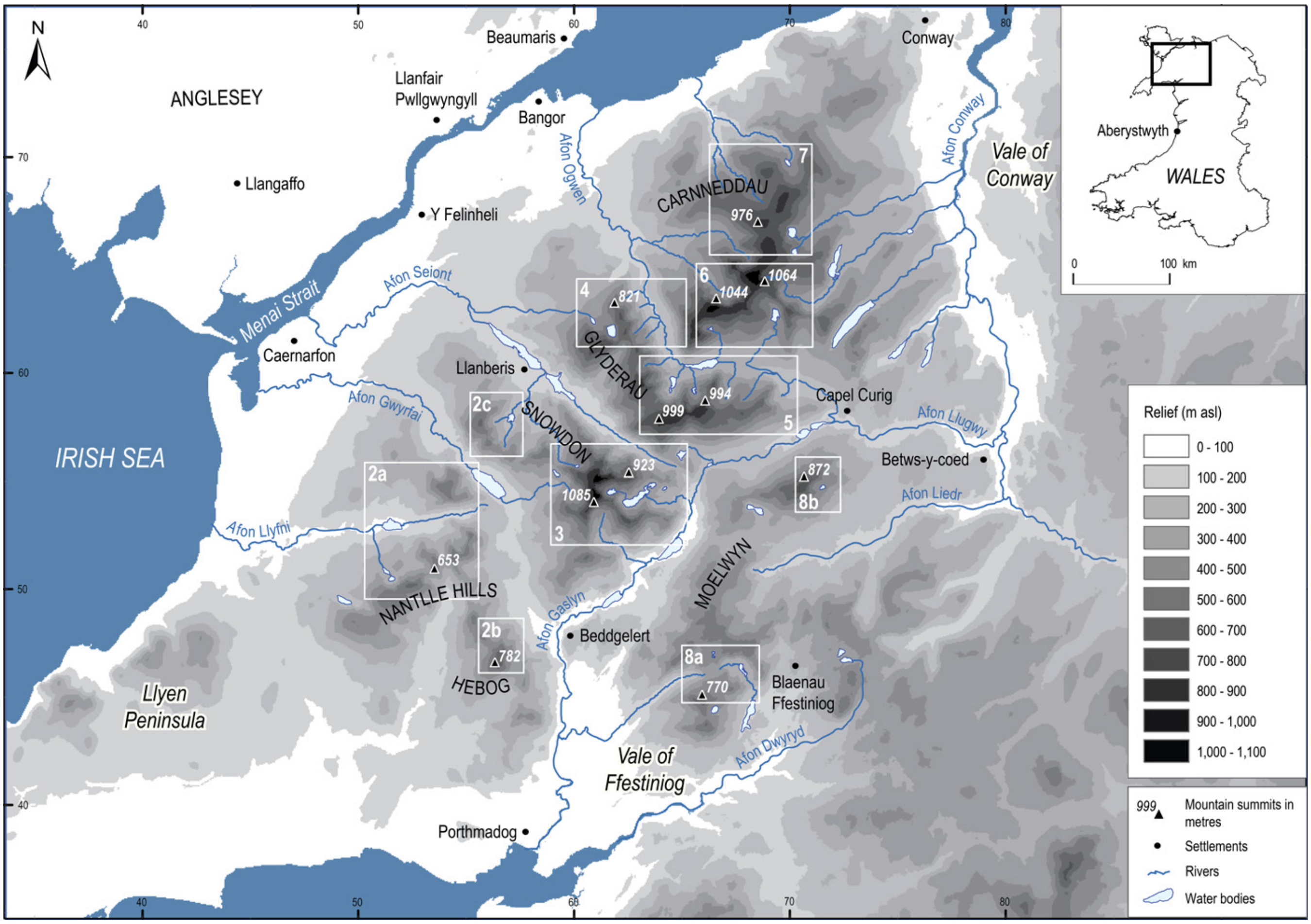

Fig. 1. Location and relief of Snowdonia. Numbered boxes represent the areas mapped in this study (Figs. 2–8). Inset shows national context.

nourished glaciers in Snowdonia; (2) reconstruct the three-dimensional form of inferred glaciers and establish the timing of their existence and; (3) examine the palaeoclimatic significance of the reconstructed glaciers.

2. Study site

Snowdonia (Fig. 1; 53°01′–53°13′N; 04°13′–03°58′W) possesses several high relief massifs, which are underlain by Ordovician volcanic and sedimentary rocks that rise to 1085 m above sea level (m asl). Bounded to the north and west by the Menai Straight and Irish Sea, to the east by the Vale of Conway and to the south by the Vale of Ffestiniog, the highest peaks are contained within three main massifs (Snowdon, Glyderau and Carnneddau) separated by the northward-draining glacial troughs of Llanberis Pass and Nant Ffrancon. Lower relief satellite ranges emanate to the south (Moelwyn) and west (Nantlle Hills and Hebog) of these larger, more prominent peaks. Glaciation through the Pleistocene has left behind a landscape marked by deep glacial troughs and high relief cirques, ice-scoured bedrock outcrops, and thick sheets of valley-floor diamict. During the last (Dimlington Stadial) glacial maximum a terrestrially based ice cap, with dispersal centres situated over the uplands of North-Central Wales, enveloped the area to become confluent with the British-Irish ice sheet (Jansson and Glasser, 2005; McCarroll and Ballantyne, 2000). Large outlet glaciers developed close to the final stages of deglaciation and drained the main ice dome northwards through major troughs located in northern and eastern Wales (Jansson and Glasser, 2005). Some workers have also proposed that small cirque glaciers reformed following dissolution of the last ice sheet (Gray, 1982; Seddon, 1957; Unwin, 1970, 1975). In the most recent and widely accepted study of local glaciation, Gray (1982) concluded that the landforms within high-level cirque basins were formed during the YDC.

3. Methods

3.1. Geomorphological mapping

The extent of local glaciation in Snowdonia was established through detailed geomorphological mapping over an area of approximately 180 km^2. As sources, we used colour vertical aerial photographs and Intermap Technologies NEXTMap Britain topographic data. Topographic data were converted to multi-azimuth (45°, 90° and 315°) relief-shaded digital terrain models (~1 m vertical and ~5 m horizontal resolution) to improve landform visualisation and minimise azimuth biasing (Smith et al., 2006; Smith and Clark, 2005). Landforms mapped digitally (e.g. glacial lineaments and mounds) were verified in the field.

Field mapping was carried out using 1:25,000 Ordnance Survey base maps. The main features recorded included former ice limits (e.g. end moraines and drift and/or boulder limits), ice movement direction indicators (e.g. striae, roches moutonneés, ice-moulded bedrock) and relict periglacial features (e.g. frost-weathered debris, scree and talus slopes, solifluction lobes and protalus ramparts) assumed to have formed outside the limits of local ice.

Landforms mapped in the field were combined with those mapped digitally and assimilated within ARCMap (ARCGIS) 9.3.

3.2. Glacier and equilibrium line altitude (ELA) reconstructions

Glacier limits were reconstructed based on the geomorphological mapping. The three-dimensional extent of individual glaciers was determined by constructing ice surface contours at 50 m intervals approximately normal to ice flow direction indicators (e.g. striae), following well-established procedures outlined in earlier studies (e.g. Ballantyne, 1989, 2002b; Carr, 2001; Sissons, 1974, 1977). In areas where glacier delimitation was more problematic (e.g. near headwalls), reconstructions were informed by the present-day topography and by analogy with modern glaciated basins.

The equilibrium line altitude (ELA) of a glacier is the altitude where annual accumulation and ablation are balanced. Contemporary glacier ELAs are closely associated with regional ablation-season temperature and accumulation-season precipitation (Osmaston, 1975, 2005; Sissons and Sutherland, 1976; Sutherland, 1984), and therefore provide a sensitive link between glaciers and climate. For that reason, estimations of former glacier ELAs offer valuable data for palaeoclimatic inferences. To permit comparison with previous work, ELAs were calculated using the established area-weighted mean altitude (AWMA; Sissons, 1974), accumulation area ratio (AAR; Porter, 1975, 2001) and area–altitude balance ratio (AABR; Benn and Gemmell, 1997; Furbish and Andrews, 1984; Osmaston, 1975, 2005; Rea, 2009) approaches. The relative strengths and weaknesses of these approaches are reviewed elsewhere (e.g. Benn and Lehmkuhl, 2000; Benn et al., 2005; Osmaston, 2005). Following previous studies, AARs of 0.5 and 0.6 and AABRs of 1.67, 1.8 and 2.0 were selected for calculations.

Glacier ELAs reflect not only the input of direct precipitation but also the addition of snow blown from adjacent terrain, so potential snowblow and avalanche sources were inferred following procedures developed by Benn and Ballantyne (2005). The delimitation of potential snowblowing sources relied on the assumption that snow-bearing winds were dominantly sourced from a southwesterly quadrant (180–270°) during the most recent phase of British glaciation, as demonstrated in previous studies (e.g. Sissons, 1979, 1980a,b; Ballantyne, 1989, 2002b). Ratios between the combined snowblow and avalanche areas and individual glacier areas were computed to account for the inevitably larger snowblow and avalanche sources of larger glaciers.

4. Results and interpretations

4.1. Synthesis of the geomorphological evidence

The results of the geomorphological mapping are shown in Figs. 2–8. The size of the study area and the abundance of landform evidence preclude a systematic description of each site in this paper. The reader is referred to Figs. 2–8 for greater detail and a broad overview of the mapping is provided below.

Evidence for former glaciers is apparent within the high-level glacial cirques throughout the study area. Within these basins a clear landsystem contrast exists across the inferred glacier limits (Fig. 9; Ballantyne and Harris, 1994; Benn and Ballantyne 2005; Lukas and Bradwell, 2010). Former ice limits are marked by distinct end and lateral moraines or the horizontal limits of glacially transported boulders ('boulder limit') and glaciogenic sediment (Fig. 9c; 'drift limit'). On higher ground, former glacial limits are often less clear; however, periglacial scree and talus deposits commonly descend to the upper ice limit (Fig. 9a), where on occasion they are truncated by drift deposits. Where apparent, the gradient of lateral moraines coincide with the altitude of other strands of geomorphological evidence (e.g. periglacial trimlines) when projected upvalley, usually to within 5–10 m.

Inside the glacial limits, ice-scoured bedrock (Fig. 9b), undulating glaciogenic drift and patchy boulder deposits dominate. Recessional moraines occur in the downvalley area of many cirques (Fig. 9a) but rarely extend farther upvalley. A notable variation in moraine size also occurs across the study area. Those cirques that exhibit comparatively large moraines (e.g. Cwm Coch; Fig. 4) however are commonly characterised by steep rock-slopes or cliffs that would have overlooked former glacier surfaces and enhanced the supply of supraglacial debris. Outside the glacier limits, periglacial phenomena are common (Fig. 9d). Frost-weathered debris covers the highest mountain summits, and protalus, scree and talus cones are found in greater abundance and thickness.

The regional consistency of the observed morphostratigraphic relationships implies that a single distinct phase of restrictive glaciation was responsible for formation of the observed landforms. Overall, the results of our mapping suggest that the last ice masses to occupy Snowdonia took the form of small, locally nourished and topographically constrained cirque glaciers. The identification of 38 glaciers in total builds on earlier mapping surveys (Gray, 1982; Seddon, 1957; Unwin, 1970, 1975), which presented evidence for a maximum of 35 independent ice masses.

4.2. The age of local palaeoglaciers

Constraining the age of this period of local cirque glaciation is imperative if the glacier limits are to act as a basis for palaeoclimatic reconstruction. The landforms contained within the cirques of Snowdonia have been interpreted as relating to several phases of glaciation by previous authors (Seddon, 1957, 1962; Unwin, 1970, 1975). However, the available dating evidence and morphostratigraphic relationships between mapped landforms support glacier occupancy during the YDC (Gray, 1982).

First, sites at Cwm Dwythwch (SH 573582) and Nant Ffrancon (SH 637633), 0.8 and 0.65 km northeast of the nearest inferred ice margins respectively, have yielded full Lateglacial stratigraphic sequences (Seddon, 1962), indicating that these sites lay outside YDC glacier limits. Conversely, sediment cores extracted from Melynllyn (SH 702657; Walker, 1978) and several small enclosed basins at Cwm Llydaw (SH 632543) and Cwm Cywion (SH 632604; Ince, 1983), inside the inferred ice limits, have revealed basal pollen stratigraphic sequences characteristic of the YDC-Holocene transition and basal radiocarbon ages of approximately 10,000 ^{14}C year BP. This would suggest that glaciers occupied these sites during the YDC. Further, samples taken from boulders covering moraine crests that dam Llyn Idwal (SH 646599) have yielded ^{36}Cl exposure ages of 12.9 ± 2.0 and 11.6 ± 1.3 ka (Phillips et al., 1994), consistent with a YDC age for these moraines.

Where chronometric dating evidence is unavailable, morphostratigraphic principles point towards similar conclusions (cf. Lukas, 2006). First, relict *in situ* periglacial features only occur beyond the mapped glacier limits. As such features were last active during the YDC (Ballantyne and Harris, 1994) their absence inside the inferred ice limits suggests that these areas must have been ice-covered at that time. Second, a clear contrast in boulder frequency exists across many of the inferred ice limits, as described at YDC glacier sites elsewhere in Britain (e.g. Finlayson, 2006; Lukas, 2006; Thorp, 1986). Moreover, boulders found within the ice limits are almost exclusively of local lithological provenance, implying that ice was both sourced from, and remained confined to, individual cirques. This contrasts with the earlier and more widespread ice cap glaciation of the Dimlington stadial, for which far-travelled erratics have been

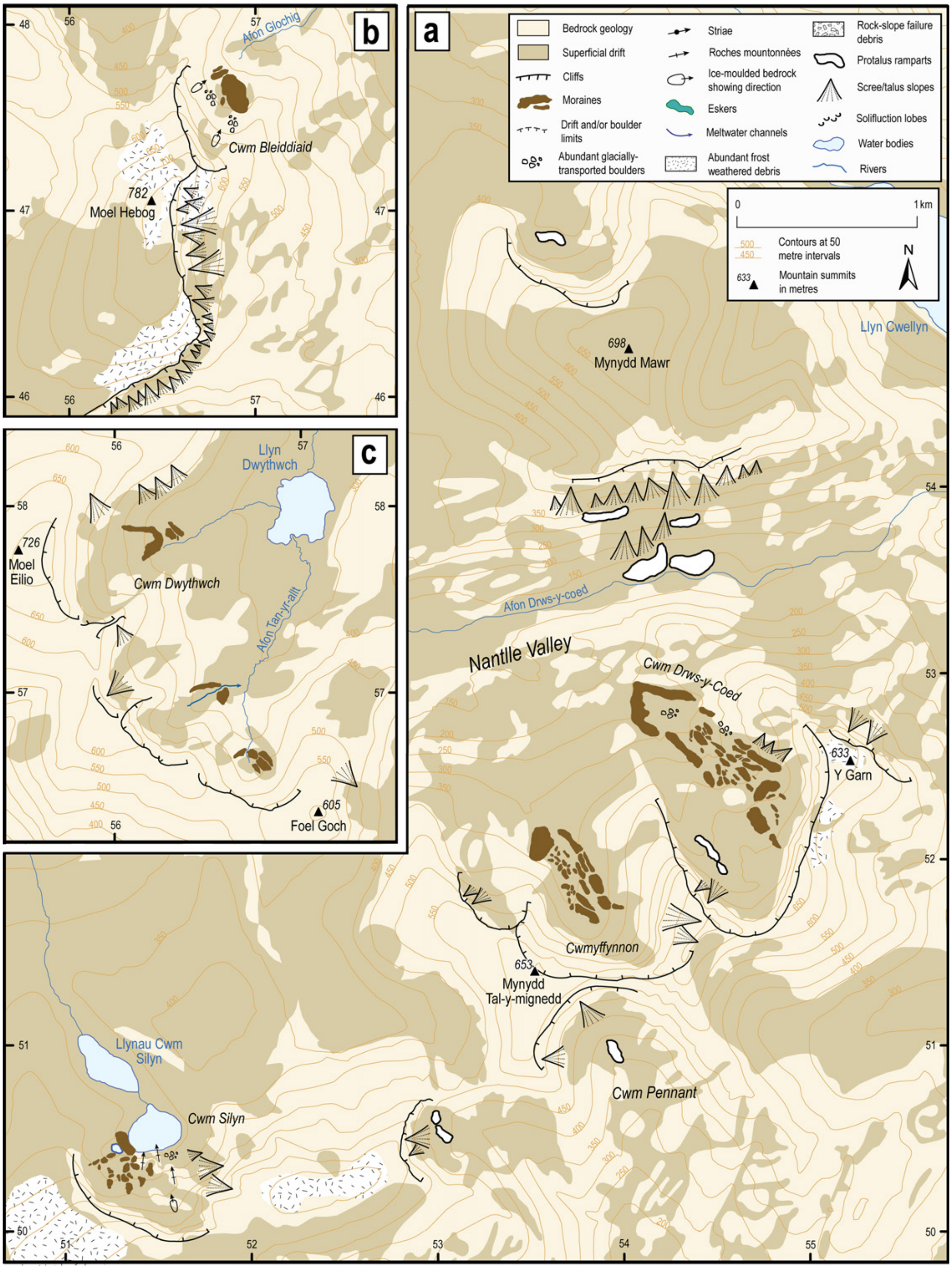

Fig. 2. Geomorphological map of (a) the Nantlle Hills and areas around, (b) Cwm Bleiddiaid and (c) Cwm Dwythwch. Cartographic symbology applies for all further geomorphological maps.

widely identified (e.g. Rowlands, 1979; Warren et al., 1984). Finally, the extent of thick glaciogenic sediment is believed to correspond to areas of former YDC ice coverage (e.g. Ballantyne, 1989, 2002b; Lukas, 2006). Consistent with these observations, thick sediment sheets terminate at the limits of several local glaciers in Snowdonia. Based on the convergence of chronometrically dated and morphostratigraphic evidence, a YDC age can therefore be inferred with some confidence for the glacial limits depicted in this study.

4.3. Glacier and equilibrium line altitude reconstructions

The geomorphological and age-specific evidence discussed above were used to reconstruct 38 topographically constrained cirque glaciers of inferred YDC age (Fig. 10). Assuming that the maximum ice limits were reached simultaneously, the YDC glaciation of Snowdonia occupied a total area of 20.74 km^2. A glacial area of 7.93 km^2 developed in the cirques of the Snowdon massif, owing mainly to the influence of the relatively large glacier

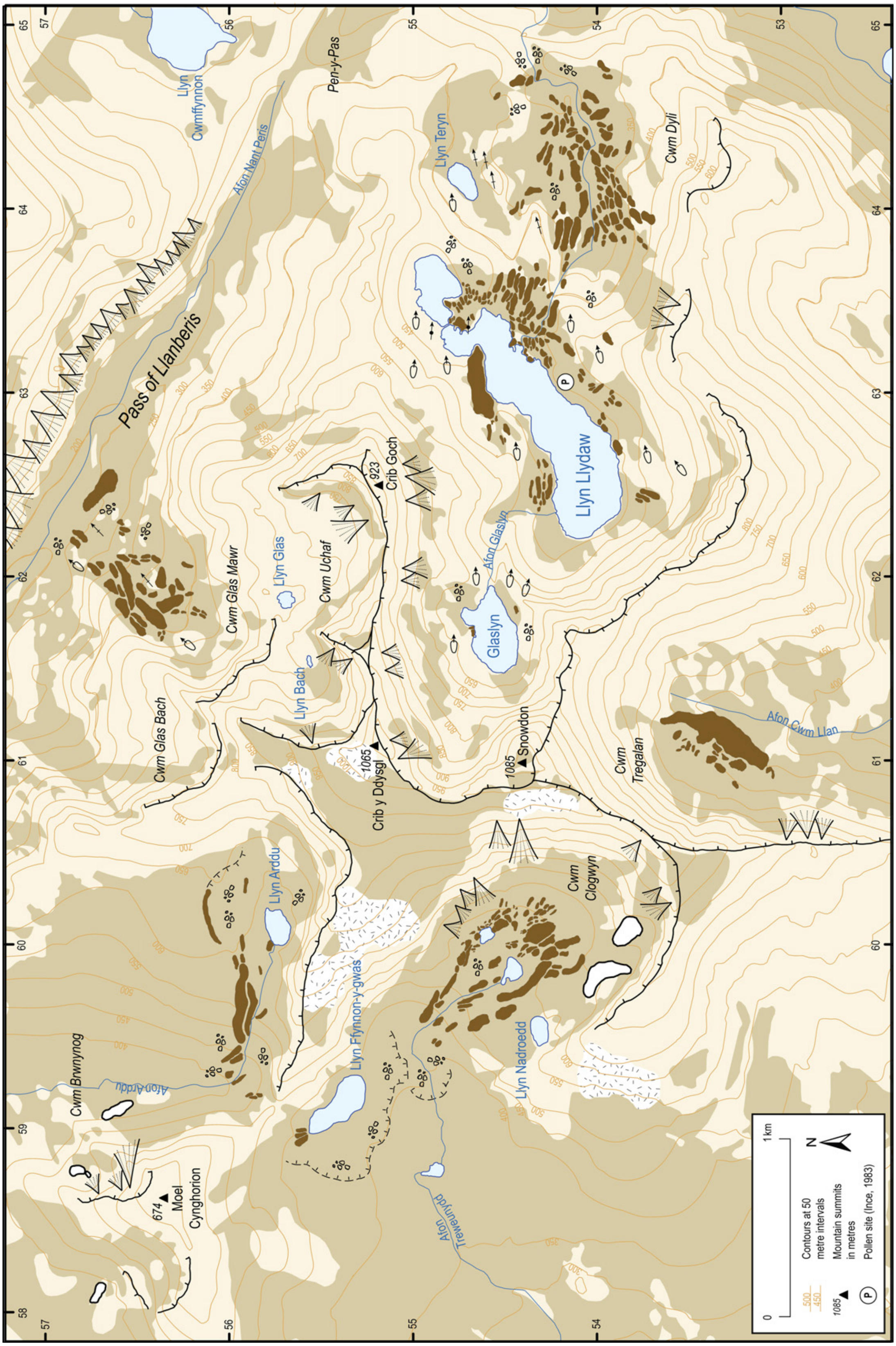

Fig. 3. Geomorphological map of the area around Snowdon (1085 m asl).

that occupied the Llydaw basin. To the northeast, ice coverage reduced, with 5.38 and 5.67 km^2 restricted to the cirques of the Glyderau and Carneddau massifs respectively. The few glaciers identified outside of the main mountain massifs account for only a minor amount (1.76 km^2) of the total ice-covered area.

In order to compare the reconstructed ice masses depicted here (Fig. 10) with those of equivalent age identified in other upland areas of Britain, and for the derivation of contemporaneous palaeoclimatic variables, former glacier ELAs were calculated. Reconstructed ELAs (Table 1) indicate that alternative approaches

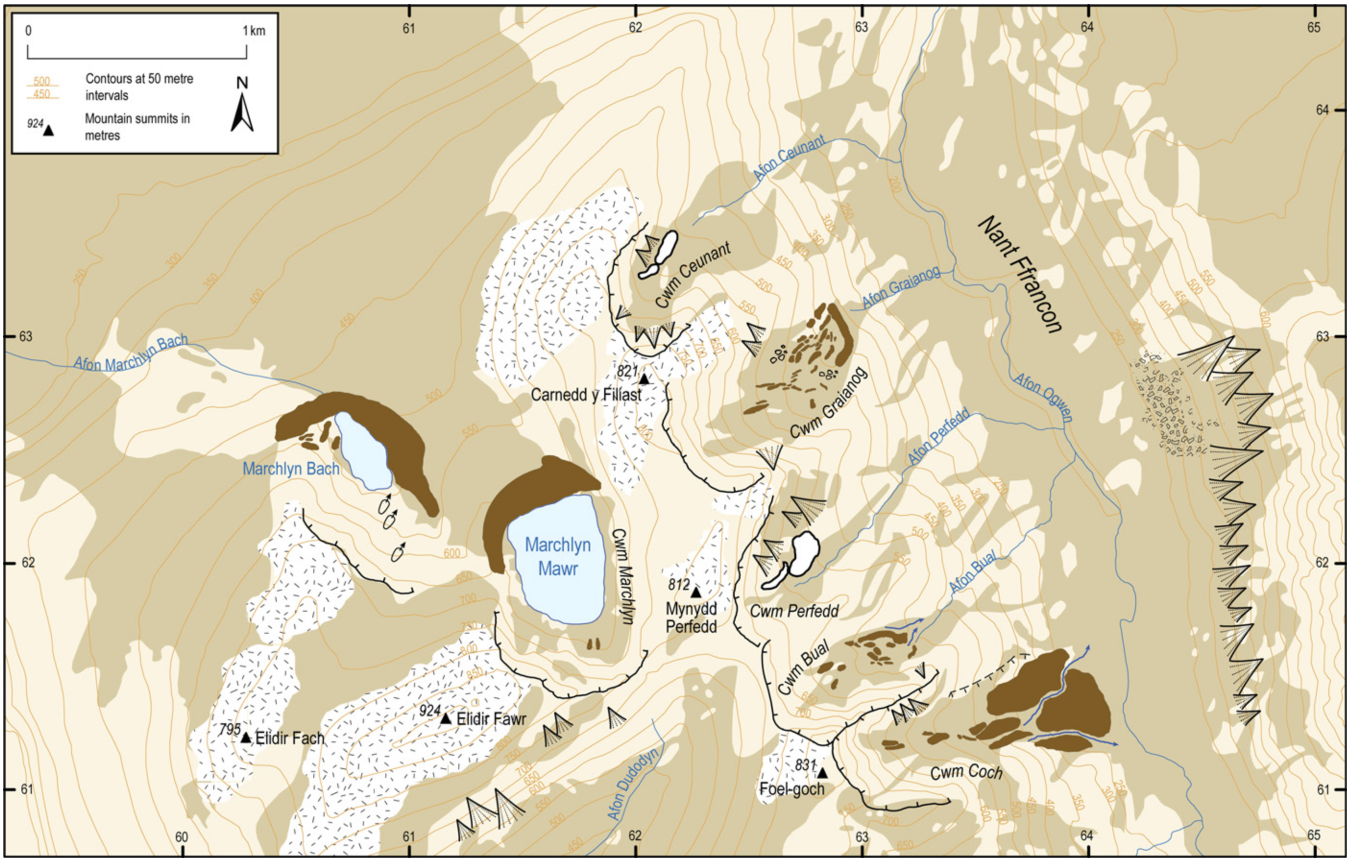

Fig. 4. Geomorphological map of the areas around Marchlyn Mawr and Nant Ffrancon.

yielded similar results. As the AABR method is the most reliable in approximating the true climatic ELA (Osmaston, 2005), and as the chosen AABRs fall close to values defined for contemporary mid-latitude maritime glaciers (Rea, 2009), an AABR value of 1.8 (Table 1) is taken to represent the best-estimate ELA. Values of AABR 1.8 vary markedly across the study area and reveal a general northeastwards rise in the altitude of ELAs from 380 m asl at Cwm Drws-y-Coed in the Nantlle Hills to 837 m asl at Cwm Caseg in the eastern mountains of the Carneddau. Calculated mean regional ELAs are also testament to this observation. Our best-estimate total mean ELA of 571 m asl is somewhat lower than the 600 m asl calculated by Gray (1982). The difference can be partly explained by the alternative approach to ELA calculation (cf. AABR 1.8 and AWMA in Table 1), but also reflects the revised glacier limits.

Despite the consistent northeastwards rise in glacier ELAs, suggesting a possible northeastwards decline in snowfall during the YDC, it is important to consider the role of windblown and avalanched snow accumulation before regional ELA trends are determined (Carrivick and Brewer, 2004; Mitchell, 1996). While the results (Fig. 11) demonstrate a less dominant control on YDC glacier ELAs than reported elsewhere (e.g. Ballantyne, 2007a; Ballantyne et al., 2007), their significance can be demonstrated nonetheless. Extrapolation of the regression line equation for the regions displaying the strongest relationships between combined snowblow and avalanche areas and regional ELA, suggest that for a hypothetical glacier with no snowblow or avalanche potential, the regional 'climatic' ELA would have been located at 654 m asl over the Moelwyns and 770 m asl over the Carneddau, significantly higher than their standard ELA equivalents in Table 1. This factor may explain the apparent failure of glacier development in several seemingly favourable cirques (cf. Ballantyne, 2007a), as these basins had floors below the regional climatic ELA and supported only limited snowblow and avalanche sources.

5. Palaeoclimatic reconstructions

5.1. Palaeotemperature and palaeoprecipitation: results

The reconstructed ELAs of YDC glaciers in Snowdonia (Table 1) provide an opportunity to retrodict palaeoclimate. For comparative purposes, palaeoprecipitation was estimated using the non-linear precipitation/temperature relationship of Ohmura et al. (1992), which has been widely employed in British glacier-derived climatic reconstructions (e.g. Benn and Ballantyne, 2005; Finlayson, 2006; Lukas and Bradwell, 2010). The function is founded upon empirically derived climatic data from the ELAs of a global sample of contemporary glaciers, and is expressed as:

$$\mathrm{Pa} = 645 + 296T_3 + 9T_3^2 \quad (1)$$

where Pa is the sum of winter balance and summer precipitation (mm a^{-1}) and T_3 is the three-month (June–August) mean summer temperature (°C) in a free air atmosphere at the ELA. Based on coleopteran data from Llanilid, South Wales (SS 944818, 60 m OD), mean July temperatures (T_j) of 10.5 °C for the coldest part of the YDC (Walker et al., 2003), were converted to T_3 on the presumption that (cf. Benn and Ballantyne, 2005):

$$T_3 = 0.97\,T_j \quad (2)$$

More recently, the use of a global dataset for glacier-derived climate reconstructions has come under scrutiny. Several studies

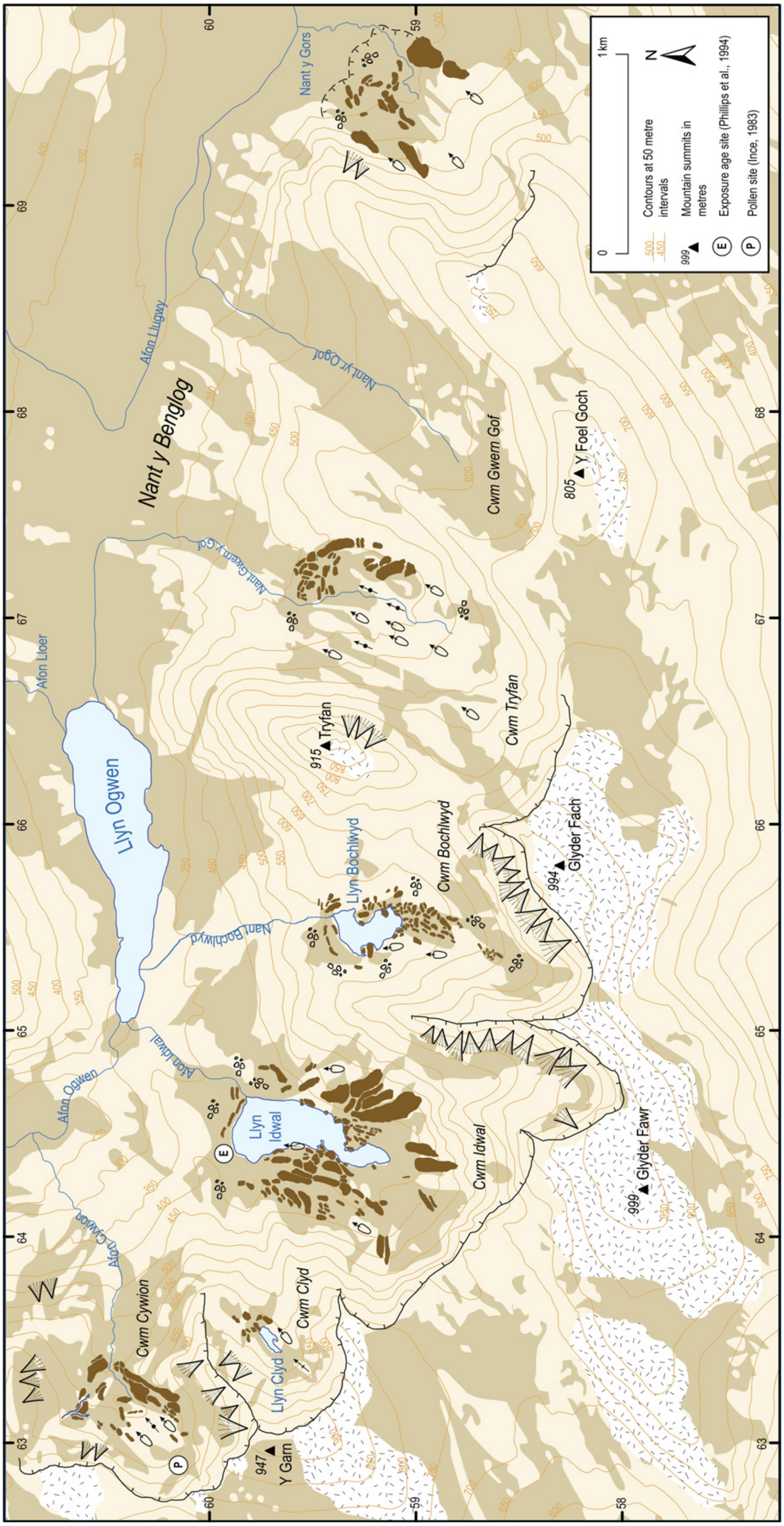

Fig. 5. Geomorphological map of the Nant y Benglog valley.

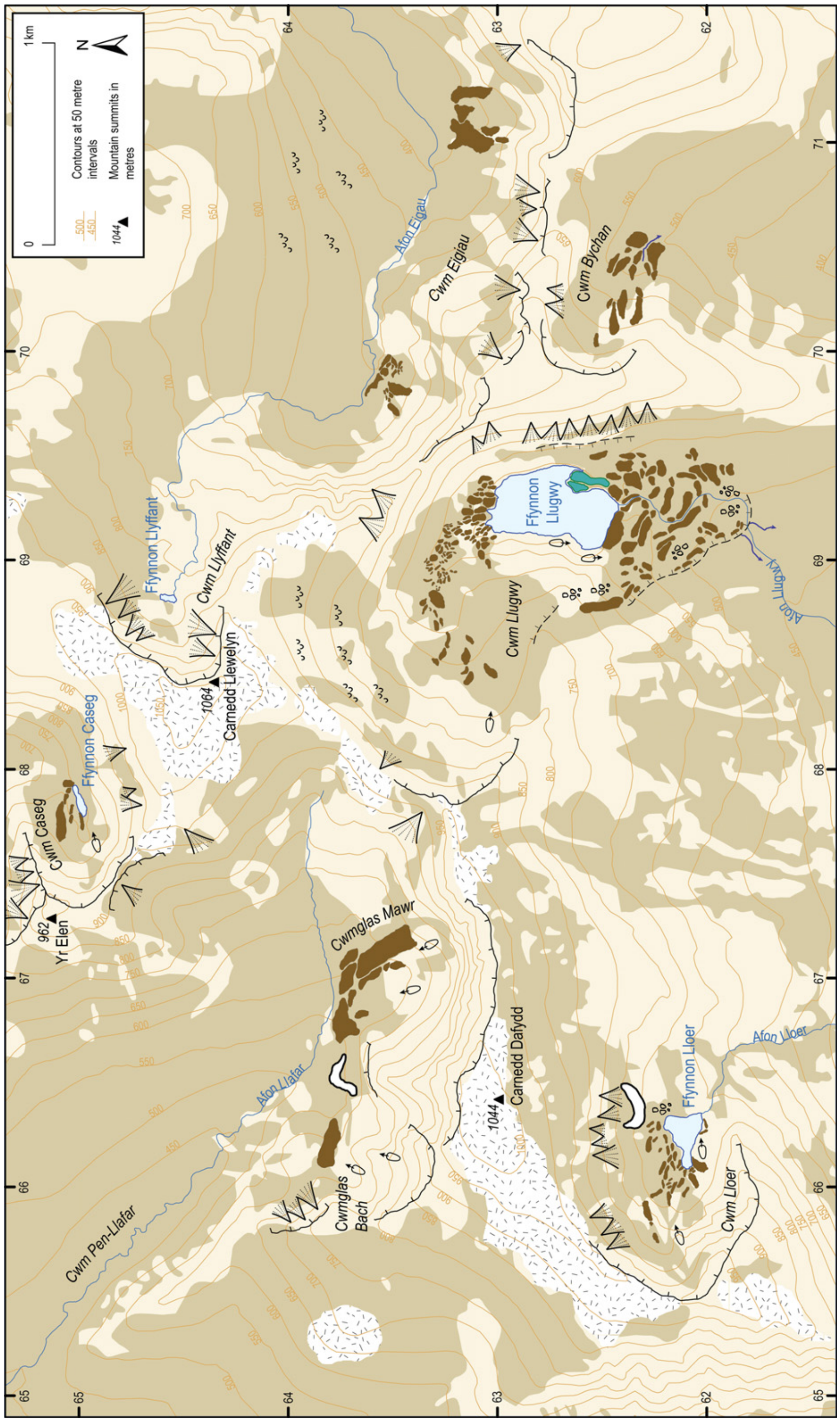

Fig. 6. Geomorphological map of the areas surrounding Carnedd Dafydd (1044 m asl) and Carnedd Llewelyn (1064 m asl).

have reported that shifts in the climatic seasonality of temperature and precipitation were particularly prominent during the YDC (e.g. Denton et al., 2005; Lie and Paasche, 2006; Thomas et al., 2008) and exerted strong control on the mass balance characteristics of glaciers (Golledge et al., 2010). Palaeoclimatic reconstructions that employ the global precipitation/temperature function of Ohmura et al. (1992) disregard the influence of regional climatic seasonality and as a consequence may lead to erroneously high palaeoprecipitation totals (Golledge et al., 2010). Therefore, a degree-day model (DDM) approach (cf. Brugger, 2006; Hughes, 2008), which allows for a user-defined input of annual temperature range (Hughes and Braithwaite, 2008), was employed to estimate the accumulation required to offset annual glacier melting. Using a mean annual temperature of −6 °C for the coldest part of the YDC

Fig. 7. Geomorphological map of the areas around Foel Grach (976 m asl) and Cwm Anafon.

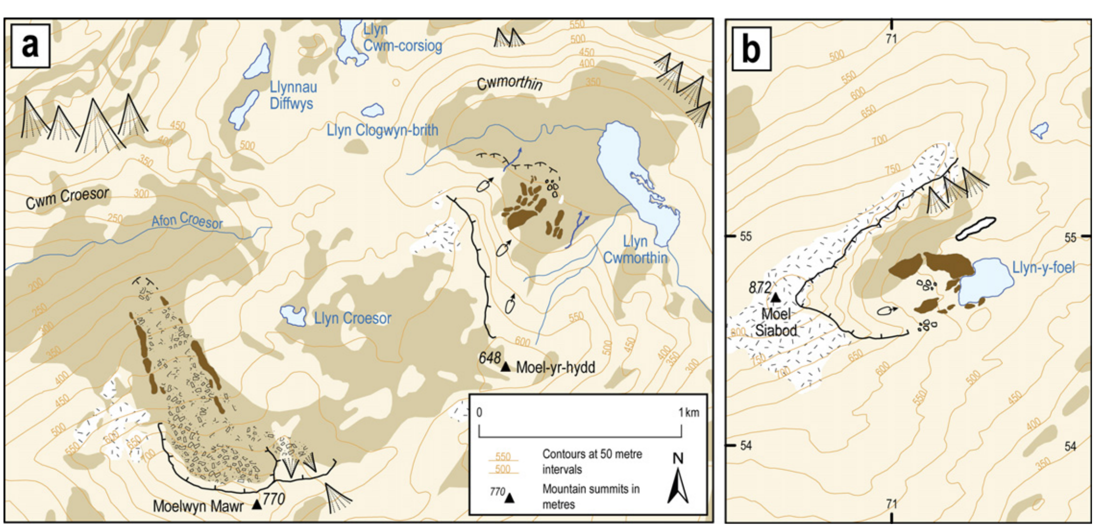

Fig. 8. Geomorphological map of (a) the areas surrounding Moelwyn Mawr (770 m asl) and (b) Cwm Siabod.

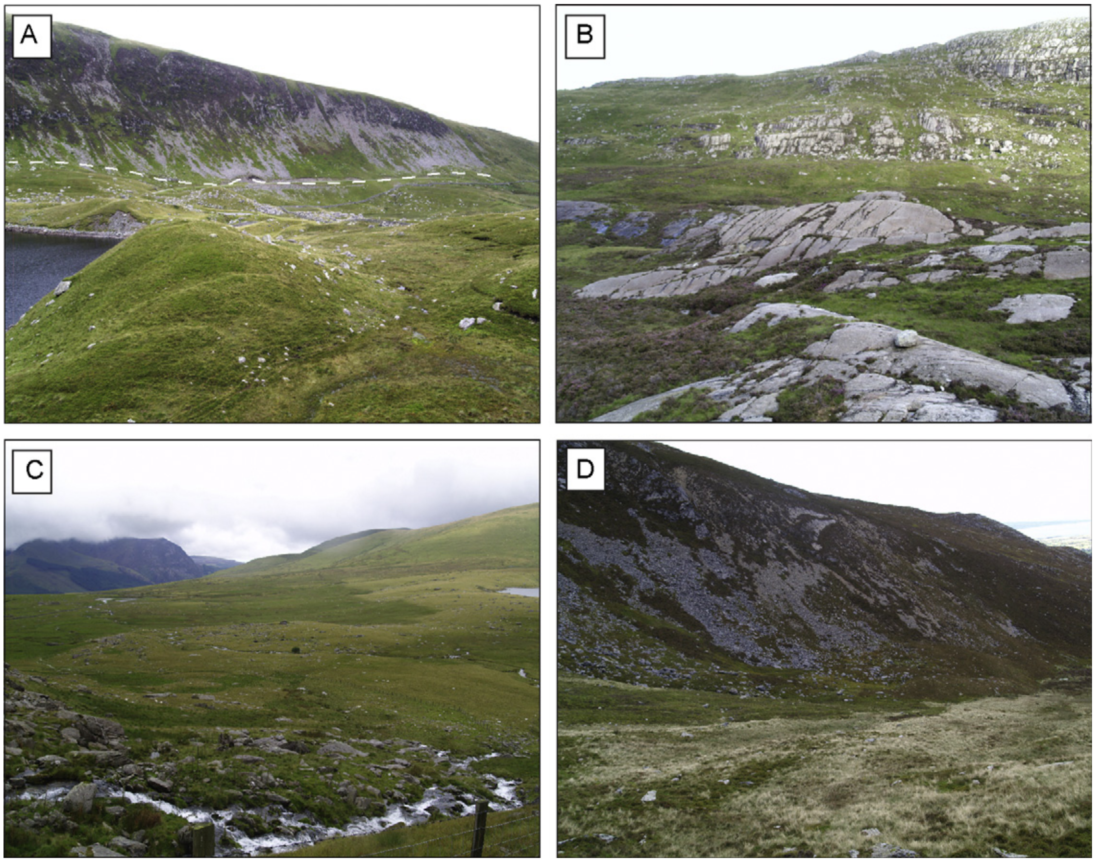

Fig. 9. Field evidence displaying the landsystem contrast observed inside and outside of the inferred glacier limits. (A) Recessional moraines and boulder deposits on the floor of Cwm Llugwy (below dashed line) contrast with mature talus accumulations at higher elevation on the eastern valley-side (above dashed line). (B) Ice-scoured and striated bedrock outcrops at Cwm Tryfan typical of the bedrock outcrops found inside many inferred glacier limits. Ice flow was from right to left. (C) Abruptly terminating drift limits in Cwm Clogwyn and neighbouring Cwm Ffynnon-y-Gwas. (D) Talus slopes and cirque floor protalus ramparts in Cwm Ceunant are representative of sites that were not glaciated during the YDC.

in North Wales (Isarin et al., 1998; their Fig. 2), daily temperature means were calculated as:

$$T_d = A_y \sin\left(\frac{2\pi d}{\lambda - \Phi}\right) + T_a \tag{3}$$

where T_d is the mean daily air temperature (°C), A_y is the amplitude of annual temperature (i.e. half of the annual temperature range), d is the day of the year (1–365), λ is the period (365 days), Φ is the phase angle (1.93) and T_a is the mean annual air temperature (°C). Adopting a degree-day factor of 4.1 ± 1.5 mm day^{-1} K^{-1} (cf. Braithwaite, 2008) and an annual temperature range (34 °C) akin with YDC palaeoenvironmental proxies (e.g. Isarin et al., 1998; their Fig. 4), daily snowmelt could be calculated and summed to give an estimate of annual accumulation (mm a^{-1}). Assuming environmental lapse rates of 0.006–0.007 °C m^{-1}, 'winter balance plus summer precipitation' and annual accumulation at the mean regional ELAs of all YDC cirque glaciers (Table 1) was calculated using Eqs. (1)–(3) (Table 2).

For reasons given above, the results of the DDM approach (Eq. (3)) are considered to be most representative of potential annual accumulation during the YDC in Snowdonia. Ignoring the additional influence of snowblow and avalanche input, regional accumulation at the ELA during the YDC varied from 2073–2687 mm a^{-1} to 1782–2470 mm a^{-1}, employing environmental lapse rates of 0.006 and 0.007 °C m^{-1}, respectively. Results derived from the regional climatic ELAs were somewhat lower, ranging from 1791 to 2616 mm a^{-1} (lapse rate: 0.006 °C m^{-1}) to 1473–2390 mm a^{-1} (lapse rate: 0.007 °C m^{-1}).

5.2. Interpretation of the palaeoclimatic data

5.2.1. Stadial precipitation

Meteorological data from the Capel Curig station (SH 720582, 216 m OD) show that present-day (1971–2009) precipitation averages around 2650 mm a^{-1} (Meteorological Office, 2010), but much higher values, with annual peaks reaching *c.* 4300 mm a^{-1}, have been reported over the higher mountains (Dore et al., 2006; Meteorological Office, 1977), and imply that annual accumulation during the YDC was less that at present, pointing to a colder, slightly more arid climate for this period. Also of interest is the spatial pattern in YDC accumulation. At present, mean annual precipitation is greatest around the summit of Snowdon (1085 m asl) and declines to the southwest and east (Meteorological Office, 2010). The regionalised values of annual accumulation derived from former YDC glaciers propose an alternative scenario; a strong stadial precipitation gradient, which complements the northeastwards rise in stadial ELAs (Table 1) and probably reflects the northeasterly navigation of snow-bearing winds across Snowdonia during the YDC and effective snow capture by the cirque basins located to the southwest. In this scenario, the glaciers to the northeast would have been starved of accumulation relatively, providing verification for the higher ELAs calculated in these areas.

5.2.2. Palaeoclimatic inferences from landform evidence

A notable feature of the landform evidence is that some of the inferred glacier limits consist of an outer moraine or drift and/or boulder limit that encloses several recessional moraines (Figs. 2–8). It has been widely accepted that such deposits are the product

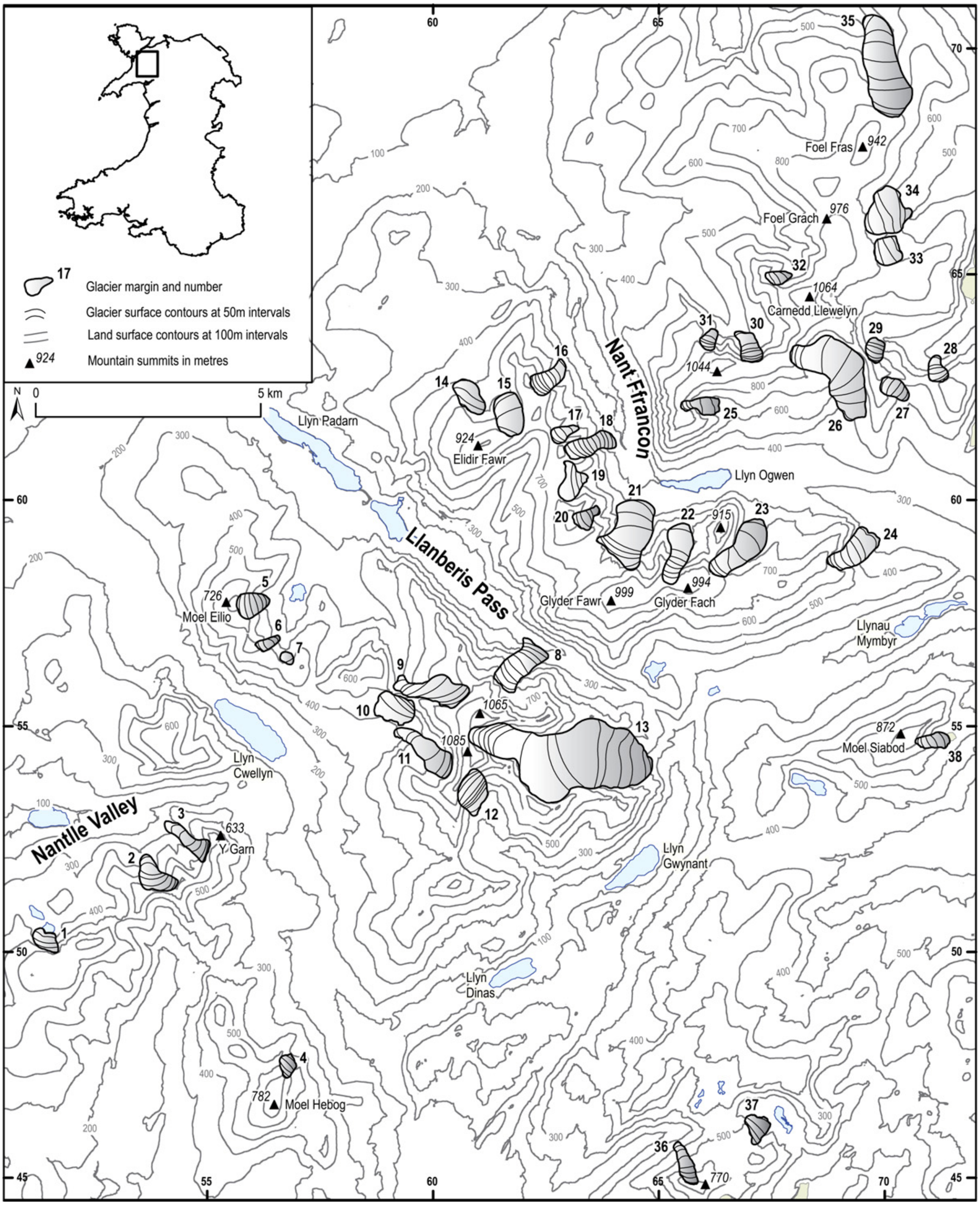

Fig. 10. Reconstructed dimensions of Younger Dryas Chronozone (YDC) palaeoglaciers in Snowdonia.

of limited periods of glacier stabilisation or readvance, which interrupted an overall trend of glacier recession (e.g. Benn et al., 1992; Bennett and Boulton, 1993). In Snowdonia, these features are largely confined to the downvalley area, with the evidence for deglaciation generally becoming patchy further upvalley. This would perhaps suggest that for some glaciers, initial ice margin retreat was punctuated by minor standstills or readvances and implies that these glaciers remained active and close to climatic equilibrium during the early phases of recession. Conversely, the paucity of recessional moraines in the mid- to upvalley areas of most former glaciers suggests that much of the deglacial phase was uninterrupted by significant readvance (cf. Benn et al., 1992). Finally, some of the smaller glaciers lack recessional moraines altogether suggesting that after attaining maximum extent these glaciers underwent sustained negative mass balance and receded rapidly once deglaciation had begun.

Although the recessional moraines found in some cirques may closely reflect transitional climatic change late in the YDC, sediment supply may have equally influenced moraine distribution. It is likely that paraglacial sediment, which accumulated in the interval between ice-sheet deglaciation and the YDC (cf. Anderson and Harrison, 2006; Ballantyne, 2002a; Harrison et al., 2010), became re-entrained as small YDC glaciers advanced over the modified landscape. The general upvalley reduction in the

Table 1
Reconstructed areas and equilibrium line altitudes (ELAs) of Younger Dryas Chronozone (YDC) palaeoglaciers in Snowdonia.

Glacier (name and number)	Area (km^2)	ELA (m asl)	AAR		AABR		
		AWMA	0.5	0.6	1.67	1.8	2.0
Nantlle Hills and Hebog glaciers							
1 Cwm Silyn glacier	0.2	442	439	425	429	427	425
2 Cwm-y-Ffynnon glacier	0.41	498	503	471	479	476	472
3 Cwm Drws-y-Coed glacier	0.33	401	391	370	382	380	376
4 Cwm Bleiddiaid glacier	0.14	516	516	502	503	502	499
All Nantlle and Hebog glaciers	*1.08*	*464*	*462*	*442*	*448*	*446*	*443*
Snowdon glaciers							
5 Cwm year Hafod glacier	0.32	549	555	531	530	527	523
6 Cwm Dwythwch (i) glacier	0.11	480	480	466	467	465	463
7 Cwm Dwythwch (ii) glacier	0.06	484	481	476	476	475	474
8 Cwm Du'r Arddu glacier	0.7	648	659	640	627	624	620
9 Cwm Glas Mawr glacier	0.66	414	418	388	388	383	378
10 Cwm Ffynnon-y-Gwas glacier	0.46	463	450	441	453	452	450
11 Cwm Clogwyn glacier	0.68	536	540	529	522	520	517
12 Cwm Tregalan glacier	0.46	677	708	662	651	648	642
13 Llydaw glacier	4.48	597	591	573	570	567	561
All Snowdon glaciers	*7.93*	*539*	*542*	*523*	*520*	*518*	*514*
Glyderau glaciers							
14 Marchlyn Bach glacier	0.35	565	562	547	555	554	552
15 Marchlyn Mawr glacier	0.5	703	706	696	693	692	690
16 Cwm Graianog glacier	0.33	576	571	533	549	545	539
17 Cwm Bual glacier	0.15	618	622	605	605	603	601
18 Cwm Coch glacier	0.41	504	512	458	468	463	456
19 Cwm Cywion glacier	0.32	689	707	695	677	675	672
20 Cwm Clyd glacier	0.18	777	784	772	762	760	757
21 Cwm Idwal glacier	1.2	525	519	503	507	505	502
22 Cwm Bochlywyd glacier	0.57	699	646	636	677	674	670
23 Cwm Tryfan glacier	0.8	597	580	568	581	579	576
24 Cwm-y-Gors glacier	0.57	410	399	362	390	387	383
All Glyderau glaciers	*5.38*	*606*	*601*	*580*	*588*	*585*	*582*
Carneddau glaciers							
25 Cwm Lloer glacier	0.24	753	739	726	739	737	735
26 Cwm Llugwy glacier	1.67	684	696	675	663	660	656
27 Cwm Bychan glacier	0.22	655	659	638	640	637	634
28 Cwm Eigiau (i) glacier	0.18	522	533	508	504	502	498
29 Cwm Eigiau (ii) glacier	0.18	624	626	601	607	605	602
30 Cwmglas Mawr glacier	0.28	713	689	657	692	689	685
31 Cwmglas Bach glacier	0.13	702	685	667	686	684	680
32 Cwm Caseg glacier	0.13	850	849	833	838	837	834
33 Melynllyn glacier	0.31	708	709	697	698	696	694
34 Cwm Dulyn glacier	0.74	657	649	638	644	642	639
35 Cwm Anafon glacier	1.59	626	629	602	606	603	599
All Carneddau glaciers	*5.67*	*681*	*678*	*658*	*665*	*663*	*660*
Moelwyn glaciers							
36 Moelwyn Mawr glacier	0.26	548	542	526	529	526	522
37 Cwmorthin glacier	0.23	414	405	397	403	402	400
38 Cwm Siabod glacier	0.19	612	603	585	598	596	593
All Moelwyn glaciers	*0.68*	*525*	*517*	*503*	*510*	*508*	*505*
All glaciers	*20.74*	*590*	*588*	*568*	*573*	*571*	*568*

quantity and size of recessional moraines may therefore in part reflect the progressive exhaustion of readily available paraglacial sediment over time (cf. Ballantyne, 2002a).

6. Discussion

We have presented geomorphological evidence for glacier advance during the YDC of *c.* 12.9–11.7 ka. In general, the reconstructed glacier dimensions vindicate the accuracy of Gray's (1982) mapping. However, Gray (1982) mapped a total of 35 glaciers, but we have identified 38, which includes new evidence from sites in the Moelwyns and at Moel Hebog. The debris accumulations at Moelwyn Mawr and Cwm Bochlwyd, which have been previously interpreted as fossil rock glacier (Lowe, 1993; Harrison et al., 2008; Rose, 2001) and sturzstrom (Harrison, 1992) deposits, are interpreted here as the product of rock slope failures onto active glacier surfaces during deglaciation. This possibility is supported by the morphology of these debris accumulations and their position exclusively inside bounding outer moraines.

Ballantyne (2002b) speculated that annual YDC precipitation totals on the Scottish west coast were significantly higher than at present. Subsequent studies have utilised proxy-based temperature records and the regression function of Ohmura et al. (1992) (Eq. (1)) to yield similar estimates of high annual snowfall at other YDC sites in western Scotland (e.g. Benn and Lukas, 2006; Lukas and Bradwell, 2010), with the findings cited as evidence for stormier, more humid conditions in these localities at this time. In the Grampian Highlands, reconstructed precipitation totals comparable with present-day suggest a marked stadial precipitation gradient existed across Scotland (Benn and Ballantyne, 2005; Finlayson, 2006). The consistent rise in stadial ELAs and decrease in palaeoprecipitation across Snowdonia suggests the predominance of southwesterly snow-bearing winds and that similar circulation patterns operated over Wales during the YDC. Patterns inferred from YDC glaciers in the Aran and Berwyn mountains of North Wales lend credence to this interpretation (Hughes, 2009) and uphold earlier speculation that the east to northeastwards passage of snow-bearing winds was a powerful control on YDC

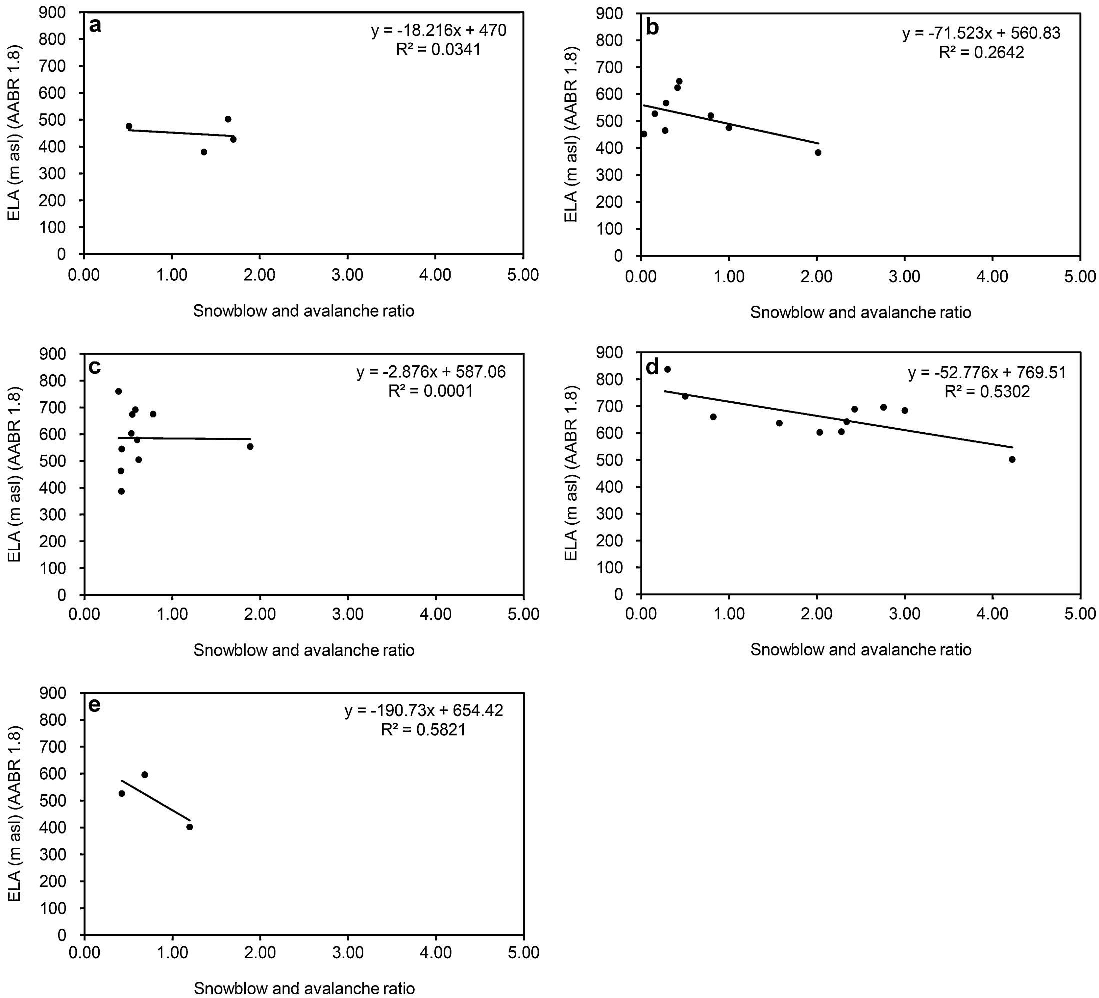

Fig. 11. Bivariate plots of correlation between snow-contributing ratios (snowblow and avalanche areas against glacier areas) and the equilibrium line altitudes (ELAs) of Younger Dryas Chronozone (YDC) palaeoglaciers which occupied (a) the Nantlle Hills and Moel Hebog range, (b) the Snowdon massif, (c) the Glyderau massif, (d) the Carneddau massif and (e) the Moelwyn massif.

glaciation in upland Britain (Ballantyne, 1989; Sissons, 1979, 1980a).

However, our results are not in complete agreement with earlier studies. Annual accumulation predicted using the DDM and independent estimates of palaeotemperature imply that the YDC in Wales was colder and drier than at present, contrary to some reconstructions in western Scotland, which suggest that up to 26% more precipitation fell than today (Ballantyne, 2002b; Benn and Lukas, 2006; Lukas and Bradwell, 2010). The possibility of lower annual snowfall totals during the YDC are more closely aligned with proxy-based evidence (e.g. Atkinson et al., 1987; Isarin and Renssen, 1999) and ice core palaeothermometry studies (Alley, 2000), which indicate that NW Europe was dominated by cold, dry winters resulting from extensive sea-ice formation in the North Atlantic, and relatively mild but short summer ablation seasons. If this was the case, and given that the limit of YDC sea-ice has been placed slightly south of Snowdonia at *c.* 50–52°N (Isarin et al., 1998; Renssen and Isarin, 1998; Renssen and Vandenberghe, 2003), YDC glaciers in North Wales are likely to have been highly sensitive to enhanced continentality and extreme winter cooling. The low winter temperatures and shortened summer ablation seasons associated with the formation of extensive sea-ice may have cancelled out reduced annual snowfall and, as a consequence, enabled YDC glaciers to survive in relatively drier conditions. Similarly, Finlayson et al. (2011) reconstructed low precipitation estimates over the Beinn Dearg massif in northwest Scotland and Golledge et al. (2010) recalculated earlier Scottish studies to yield lower values of stadial precipitation. Therefore, our findings support more recent glacier-based interpretations of YDC palaeoclimate in Britain, which suggest that climatic seasonality played a central role in forcing relatively drier conditions at the ELA of coeval ice masses.

However, the raw values of stadial accumulation in Snowdonia (Table 2) do not differ significantly from precipitation estimates at sites along the Scottish west coast (e.g. Ballantyne, 2007a; Lukas and Bradwell, 2010), despite the fact that these studies utilise the global dataset of Ohmura et al. (1992). The tendency of this function to overestimate palaeoprecipitation (cf. Golledge et al., 2010) thus implies a latitudinal contrast in stadial palaeoclimate and possibly slightly milder conditions over North Wales than in Scotland. There are several possible explanations for the regional differences in stadial palaeoclimate. First, Ballantyne (2006, 2007a,b) suggested a general southwards rise in YDC ELAs and implied ablation-season temperatures along the western seaboard of Britain. It might be expected therefore that Northwest Wales faced slightly milder summer conditions than in Scotland and possibly enhanced moisture availability (and thus accumulation) as a result. This factor may have been augmented by the annual recession of the winter sea-ice margin, which would have exposed Snowdonia to less arid conditions for longer before readvancing. Second, the wetter conditions over North Wales during the YDC may reflect proximity to the transient position of the oceanic polar

Table 2
Reconstructed palaeoclimate at the mean regional equilibrium line altitudes (ELAs) of Younger Dryas Chronozone (YDC) palaeoglaciers, calculated using the Ohmura et al. (1992) regression function (Eq. (1)) and a degree-day model (Eq. (3)).

	Regional ELA	Lapse rate (°C m^{-1})	T_3 (ELA) (°C)	Pa (mm a^{-1})
Ohmura et al. (1992)	446	0.006	7.97	3575
	446	0.007	7.53	3447
	518	0.006	7.55	3392
	518	0.007	7.05	3243
	585	0.006	7.16	3225
	585	0.007	6.59	3057
	663	0.006	6.70	3034
	663	0.007	6.06	2844
	508	0.006	7.61	3417
	508	0.007	7.11	3206
				Annual accumulation (mm a^{-1})
Degree-day model	446	0.006	6.52	2687
	446	0.007	6.08	2470
	518	0.006	6.09	2477
	518	0.007	5.57	2233
	585	0.006	5.69	2287
	585	0.007	5.10	2020
	663	0.006	5.22	2073
	663	0.007	4.56	1782
	508	0.006	6.15	2506
	508	0.007	5.64	2266
Degree-day model (regional 'climatic' ELA)	470	0.006	6.38	2616
	470	0.007	5.91	2390
	561	0.006	5.83	2354
	561	0.007	5.27	2096
	587	0.006	5.68	2281
	587	0.007	5.09	2014
	770	0.006	4.58	1791
	770	0.007	3.81	1473
	654	0.006	5.28	2097
	654	0.007	4.62	1809

front (*c.* 45–50°N; Ruddiman and McIntyre, 1981). The strong thermal gradient that existed south of the polar front probably encouraged enhanced interaction between competing air masses and stormier conditions with more abundant precipitation. Model simulations of the YDC in NW Europe support this notion and indicate patterns of strong zonal circulation and a high frequency of winter depressions (Isarin et al., 1998; Renssen et al., 2001). In reality, a combination of these factors probably conspired to force milder ablation seasons and increased precipitation at lower latitudes along the west coast of Britain. However, the possibility of slightly milder conditions over Snowdonia should not detract from the apparent increase in aridity during the YDC relative to the present-day.

The inferred severe winter cooling and reduced accumulation has implications for the dominant style of YDC glaciation in Snowdonia. Benn and Lukas (2006) presented detailed sedimentological and palaeoclimatic evidence in support of highly dynamic YDC glaciers in Scotland, and inferred extensive wet-based conditions and high annual mass turnover. However, under the climatic scenario presented here, YDC glaciers in Snowdonia would probably have been less dynamic than previously inferred (cf. Golledge et al., 2010). Short summer ablation conditions and sustained winter cooling may have encouraged YDC glaciers that behaved in similar fashion to modern high arctic polythermal glaciers, such as those which exist on Svalbard today (Hambrey et al., 1997). Indeed, studies in Britain have presented evidence for former ice dynamics not too dissimilar from existing Svalbard cirque glaciers. For instance, Sharp et al. (1989) showed that YDC glaciers in North Wales possessed very low basal sliding velocities (9 m a^{-1} or less) and could have moved largely by internal deformation alone. Polythermal glacier dynamics have also been inferred from moraine complexes of YDC age in Snowdonia based on morphological comparisons with modern high arctic moraine assemblages (Graham and Midgley, 2000; Hambrey et al., 1997). However, further work is necessary to determine the dynamics and style of YDC glaciers in Snowdonia and establish possible modern analogues.

Abrupt cooling at the onset of the YDC has traditionally been put forward as the stimulus for localised glaciation in upland Britain. However, the build-up of individual glaciers is more likely to reflect complicated interactions between precipitation and temperature over a considerably longer timescale. The development of extensive winter sea-ice during the YDC may have hindered glacier build-up due to enhanced continentality and relatively low accumulation, especially in the more eastern cirques of Snowdonia that were regionally drier. In contrast, build-up early in the YDC (or perhaps in the preceding interstadial) would have been favoured by improved moisture availability and gradual cooling. Therefore, YDC glaciers in Snowdonia probably attained their maximum extent nearer the beginning of the stadial and stabilised when more severe winter temperatures prevailed. This scenario corresponds with pollen evidence from Northwest England (Walker, 2004) and terrestrial stable isotope data from South Wales (Walker et al., 2003), which suggests that the mid-to-late YDC was characterised by lower temperatures and increasingly drier conditions than earlier in the stadial. This situation might have altered in the latter stages of the YDC, for which independent proxies indicate that temperatures rose abruptly (e.g. Atkinson et al., 1987; Walker et al., 2003) and ablation seasons lengthened (Bakke et al., 2009). The recessional moraines in the downvalley area of some former glaciers probably reflect the reaction to sustained warming during this short time-slice and a switch to a more dynamic but oscillatory deglacial state characterised by higher annual mass turnover. However, compared to examples from the Scottish Highlands (e.g. Bennett and Boulton, 1993) and English Lake District (McDougall, 2001), where large suites of recessional moraines commonly reach the valley heads of former YDC glaciers, the relative paucity of recessional moraines within the cirques of Snowdonia suggest that glaciers in North Wales reacted more rapidly to changing temperature at the end of the YDC. This may be explained by the higher ablation-season temperatures experienced at lower latitudes during this period (cf. Ballantyne, 2006, 2007a,b) and the shorter response times of small, 'marginal' glaciers to changes in temperature and precipitation (Carrivick and Brewer, 2004).

Overall, inferences drawn from reconstructed cirque glaciers in Snowdonia have improved our understanding of regionalised palaeoclimate and glacier–climate interactions during the YDC in North Wales. However, further investigation is necessary at lesser studied YDC sites in Central and Southern Wales before the response of mountain glaciers to late Pleistocene climatic change can be evaluated in a national context. The methods used here would be effective in establishing the palaeoclimatic significance of former YDC glaciers in these areas.

7. Conclusions

Geomorphological mapping of Snowdonia, North Wales has enabled the identification and reconstruction of 38 small cirque glaciers. The geomorphological and available chronometric dating evidence support a readvance during the YDC after dissolution of the last Welsh Ice Cap. ELAs reconstructed using an AABR approach range from 380 to 837 m asl and reveal a clear northeastwards rise across the study area. The trend can be attributed to a strong stadial precipitation gradient caused by the northeastwards passage of snow-bearing winds and preferential snow capture in cirque basins to the west of Snowdonia. Stadial accumulation reconstructed using a DDM approach indicates a colder, drier climate than at present, probably as a result of extensive sea-ice formation and enhanced continentality. Our results show that small glaciers on the western coast of Britain were highly responsive to climate fluctuation through the YDC. Investigations at other sites along the west coast of Britain may therefore prove useful in assessing the response of mountain glaciers to late Pleistocene environmental change and in further developing our understanding of YDC palaeoclimate.

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