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Timing of glacial retreat in the Wicklow Mountains, Ireland, conditioned by glacier size and topography

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Abstract

Reconstructing the deglacial history of palaeo-glaciers provides vital information on retreat processes; information which can inform predictions of the future behaviour of many of the world's glaciers. On this basis, this paper presents 170 Schmidt Hammer exposure ages from moraine boulders and glacially-sculpted bedrock to reveal the post-Last Glacial Maximum (LGM) history of the Wicklow Mountains, Ireland. These data suggest that large ice masses survived for 4-7 ka after retreat of the Irish Sea Ice Stream and were sustained by summit ice-fields until ~16.6 ka. Post-LGM retreat was driven by climate and involved numerous short-term ice front oscillations (≤ 1 ka), with widespread moraine deposition during Heinrich Stadial 1. In contrast, marked asynchroneity in the timing of Younger Dryas deglaciation is closely linked to snow redistribution which demonstrates the sensitivity of small cirque glaciers (≤ 1 km^2) to local topography. This result has important implications for palaeoclimate reconstructions as cirque glacier dynamics may be (at least partly) decoupled from climate. This is further complicated by post-depositional processes which can result in moraine ages (e.g. ^{10}Be) which post-date retreat. Future palaeoclimate studies should prioritise cirques where snow contributing areas are small and where post-depositional disturbance of moraines is limited.

Keywords

- Schmidt Hammer exposure dating (SHED)
- Wicklow, Ireland
- Glacier chronology
- Topographic controls
- ^{10}Be dating

Introduction

Understanding how mountain scale ice-masses retreat is important if we are to predict the future behaviour of many of the world's glaciers. Fortunately, improvements in systematic and robust dating techniques and their application to glacial and glaciofluvial deposits mean that is now possible to gain vital information by reconstructing the retreat history of many of the world's former (palaeo) ice masses. In the British Isles, recent work has focused on understanding the dynamics of the Irish Ice Sheet (IIS) during the global Last Glacial Maximum (LGM; 23.3 – 27.5 ka; Hughes and Gibbard, 2015). This work has attempted to (1) establish the dimensions of the ice sheet at its maximum extent, (2) reveal the pattern and timing of retreat and (3) understand the configuration of the IIS within the larger British-Irish Ice Sheet (BIIS). With this in mind, recent studies have generated a wealth of geochronological data (e.g. ^{10}Be, ^{36}Cl, ^{14}C) to constrain the maximum extent of glaciation (e.g. Clark et al., 2009; Ballantyne and Stone, 2015; Barth et al., 2016), the timing of terrestrial and marine based ice sheet retreat (e.g. Sejrup et al., 2005; Clark et al., 2012), the dynamics of component ice caps and mountain glaciers (e.g. Ballantyne et al., 2006; Harrison et al., 2010; Barth et al., 2018) and the retreat history of the Irish Sea Ice Stream (ISIS) (e.g. Chiverrell et al., 2013; Smedley et al., 2017a), a major outlet of the BIIS (Smedley et al., 2017b). However, while considerable progress has been made, further work is necessary to understand the extent and retreat history of many mountain ice caps and glaciers in Ireland, particularly in areas where numerical ages are lacking, and to model their interaction with the IIS and marine based ice streams after the LGM. Understanding the fundamental climatic, topographic and glaciological processes which determine mountain glacier growth/decay, as revealed through reconstruction of retreat histories, provides critical information which can inform models of future glacier behaviour in response to anthropogenic forcing of climate.

In the Irish Sea Basin, ice caps centred on the mountains of Wales (Hughes et al., 2016), the Lake District (Wilson et al., 2013) and Wicklow, Ireland (Ballantyne et al., 2006) coalesced with the ISIS during the LGM and may have persisted after ISIS retreat. However, while the deglacial chronologies of the Welsh and Lake District ice caps are constrained by ^{10}Be, OSL and ^{14}C ages (Ballantyne et al., 2009; McCarroll et al., 2010; Glasser et al., 2012; Lloyd et al., 2013; Hughes et al., 2016; Smedley et al., 2017b), there is a paucity of geomorphological and geochronological evidence for post-LGM activity in the Wicklow Mountains. ^{10}Be ages from summits in Wicklow (Ballantyne et al., 2006) and the adjacent Blackstairs Mountains (Ballantyne and Stone, 2015) indicate summit deglaciation soon after the LGM (n = 5; 21.0 - 22.9 ka), while a single ^{36}Cl age from the Mottee Stone, a large glacial erratic (~150 tonnes) transported ~13 km SE from its Wicklow source area, indicates separation of terrestrial ice and the ISIS by 23.1 ± 2.2 ka (Bowen et al., 2002). This timeframe accords with Bayesian modelling of ice stream retreat, with the ISIS retreating ~80 km along the Irish coast during the interval 20.8 – 23.9 ka (Smedley et al., 2017b). Recent research has begun to establish the geomorphological context for deglaciation in the Wicklow Mountains (Knight et al., 2017). However, with the exception of isolated ^{14}C, ^{36}Cl and ^{10}Be ages from cirque moraines at Lough Nahanagan and Kelly's Lough (Colhoun and Synge, 1980; Bowen et al., 2002; Barth et al., 2018), the retreat history of the Wicklow Ice Cap is poorly constrained by numerical ages. To address this knowledge gap, this study presents 170 Schmidt Hammer (SH) exposure ages from cirque and valley moraines and from a summit overridden by ice at the LGM. These data provide (1) the first comprehensive retreat history for the Wicklow Ice Cap, (2) new information on the extent, timing and dynamics of Late Pleistocene mountain glaciation; information which complements a growing body of research in Ireland (e.g. Barth et al., 2016) and within the Irish Sea Basin (e.g. Hughes et al., 2016), and perhaps most significantly, (3) new insight into the climatic and topographic factors which conditioned post-LGM retreat.

Methods

To develop a deglacial chronology, sampling was focused on prominent moraines and boulder accumulations as these are the best geomorphological indicators of the dimensions of former mountain glaciers (Barr et al., 2017). Key sites along the main SW-NE axis of the mountain range were targeted for Schmidt Hammer exposure dating (SHED; Tomkins et al., 2016) including glacially-deposited boulders on prominent cirque moraines (>400 m) at Kelly's Lough (KL), Lough Nahanagan (LN), Mullaghcleevaun (MC) and Upper Lough Bray (ULB). Moraines targeted for SHED exhibit good spatial coherence (Kirkbride and Winkler, 2012) as they are generally matrix-poor, boulder-rich and feature clearly defined moraine crests, although the outer cirque moraine at Lough Nahanagan is degraded (Colhoun and Synge, 1980). This moraine is ~1 km in length, broadly convex in cross-profile form and consists of unsorted granite, sand and gravel deposits with entrained glacially smoothed granite boulders (1 - 4 m diameter; Colhoun and Synge, 1980). Multiple nested moraines and boulder accumulations are preserved within the inferred YD glacial limit at each site, which conforms to a pattern of active oscillatory retreat (Bickerdike et al., 2017). In addition, samples were obtained from valley moraines (250 – 400 m, c. 2 - 4 km from cirque headwalls) at Carrawaystick Brook (CB), Upper Glendasan (UGD), Lough Brook (LB) and Glenmacnass Waterfall (GW) and from ice-moulded bedrock and erratic boulders from the summit of Carrigshouk (CS; 571 m), which was overridden by ice at the LGM (Fig. 1; Table 1). 20 surfaces were sampled at each site (Carrawaystick Brook; n = 10) and 170 surfaces were sampled in total, comparable to previous applications of SHED in the Mourne Mountains, Northern Ireland (Barr et al., 2017).

30 R-values were generated per surface (Niedzielski et al., 2009). Sampled boulders were of sufficient size (> 25 kg; Sumner and Nel, 2002; Demirdag et al., 2009) and all sampled surfaces were free of surface discontinuities (Williams and Robinson, 1983) and lichen (Matthews and Owen, 2008). All sampled surfaces were quartz-rich, medium-coarse grained Caledonian granite (GSI, 2013; Bedrock Geology; Scale 1:500,000), with no clear spatial variability in grain size or rock composition. Although smaller scale geological maps indicate some variability between sampled sites (GSI, 2016; Bedrock Geology; Scale 1:100,000), the predominant style of weathering is sub-aerial, as evidenced by granular disintegration of the rock surface (André, 2002; Tomkins et al., 2018b). Although weathering rate variability cannot be excluded as an explanation for contrasting exposure age distributions across geological boundaries, it appears unlikely that these surfaces would weather at significantly different rates given their comparable grain size (1-5 mm), quartz content (~20%), degree of lichen colonisation and phenocryst size (≤ 30 mm). Moreover, given the long-timescales of exposure (≥ 11 ka) and limited climatic variability across the relatively small mountain range (~220 km^2), any differences in surface R-Values due to lithology will likely be significantly smaller than the effect of variable exposure age. This interpretation is supported by large spatial scale ^{10}Be-SH calibration curves from the British Isles (Tomkins et al., 2016; 2018a; n = 54; $R^2 = 0.94$, $p < 0.01$) and the Pyrenees (Tomkins et al., 2018b; n = 52; $R^2 = 0.96$, $p < 0.01$) which indicate that the primary control on surface R-Values is cumulative exposure to sub-aerial weathering. *Instrument calibration* (Correction Factor = 1.017) and *age calibration* (Correction Factor = 0.992) were performed using the SHED-Earth online calculator (http://shed.earth) following the recommendations of Dortch et al. (2016) and Tomkins et al. (2018a). SH exposure ages and 1σ uncertainties were calculated based on the arithmetic mean for each surface (Mean of 30 R-values) and based on the updated calibration curve of Tomkins et al. (2018a) which includes ^{10}Be dated surfaces from Blackstairs Mountain, Wexford (n = 2; Ballantyne and Stone, 2015) and Bloody Foreland, Donegal (n = 6; Ballantyne et al., 2007; Clark et al., 2009). These data fit the trend established at calibration sites in Scotland and NW

England and indicate that errors in SH exposure age estimates due to climatic variability appear unlikely (Barr et al., 2017). Calibration site exposure ages are calculated using the online calculators formerly known as the CRONUS-Earth online calculator (http://hess.ess.washington.edu/math/, Wrapper script 2.3, Main calculator 2.1, constants 2.3, muons 1.1; Balco et al., 2008) and are based on the time-dependent Lm scaling (Lal, 1991; Stone, 2000), the Loch Lomond Production Rate (LLPR; Fabel et al., 2012; 4.02 ± 0.18 atoms g^{-1} a^{-1}) and assuming 0 mm ka^{-1} erosion. The LLPR is constrained by independent ^{14}C ages (MacLeod et al., 2011) and is the most widely used local production rate in the British Isles. However, the results presented here will be subject to recalibration in light of future refinement of local production rates. To ensure consistency when comparing the results of this study with independent numerical ages from Wicklow and other mountain massifs, ^{10}Be ages from Ballantyne et al. (2006), McCarroll et al. (2010), Hughes et al., (2016) and Barth et al. (2018) were recalibrated following the methods described above. Reported ^{36}Cl ages have not been recalibrated (Bowen et al., 2002; Ballantyne et al., 2009) due to incomplete sample information.

Reported SH exposure ages are interpreted to reflect the cumulative exposure of rock surfaces to sub-aerial weathering. Based on the assumption that sampled rock surfaces are glacial in origin e.g. boulders deposited on moraines or bedrock surfaces eroded sub-glacially, the onset of exposure can be considered contemporaneous with deglaciation. However, boulders can reside sub-aerially for considerable periods after glacial retreat (Hughes et al., 2016), leading to rock surface shielding, minimal sub-aerial weathering and higher Schmidt Hammer R-Values. These surfaces would generate SH exposure ages which post-date retreat and instead, likely reflect the timing of boulder exhumation and the stabilisation of the moraine ridge (Hallet and Putkonen, 1996). Given the growing consensus that moraine ages are more likely influenced by post-glacial instability than prior exposure (Heyman et al., 2011), the most cautious approach is to interpret SHED data as minimum limiting ages (Briner et al., 2005). The influence of prior exposure on surface R-Values is currently unclear, with limited data on the depth-dependence of sub-aerial weathering in granitic surfaces. For ^{10}Be, surface erosion of 3-5 m is necessary to remove the accumulated in-situ cosmogenic signal (Gosse and Phillips, 2001; Hughes et al., 2016; Briner et al., 2016). However, the depth of erosion required to 'reset' the surface for Schmidt Hammer testing is not known. If the required depth is comparable to ^{10}Be, then is it possible that surfaces could retain a weathering signature from a previous period of exposure. While it is clear that further work is needed to address this uncertainty, and the uncertainty introduced by moraine stabilisation processes, these issues can be mitigated by collecting statistically large datasets and by analysing the distribution of calculated ages (e.g. Dortch et al., 2013; Murari et al., 2014) to identify outlier ages which are compromised by geological uncertainty. For each sampled site (n = 9), probability density estimates (PDEs) were produced and modelled to separate out the highest probability Gaussian distribution (Fig. 2; Dortch et al., 2013). Using the KS density kernel in MATLAB (2015) and a dynamic smoothing window based on age uncertainty, PDE peaks and tails were separated into individual Gaussian distributions, the sum of which integrates to the cumulative PDE at 1000 iterations to obtain the best fit. The re-integrated PDE (made from the isolated Gaussians) goodness of fit is indicated graphically (Dortch et al., 2013). This analytical method has been employed in studies using ^{10}Be (c.f. Dortch et al., 2013; Murari et al., 2014) to account for negative or positive skew of datasets and to identify ages that are too young (*moraine degradation*; Heyman et al., 2011) or too old, respectively (*inheritance*; Hallet and Putknonen, 1996). Full sample information for the 170 sampled surfaces sampled can be found in the Supplementary Dataset.

Based on the results of SHED, three dimensional reconstructions of cirque glaciers were generated using the GLaRe tool (Pellitero et al., 2016; Basal shear stress = 100 kPa; Step length = 10 m) and used to estimate palaeo equilibrium-line altitudes (ELAs). Valley glaciers were also reconstructed for individual catchments using this method although ELAs were not calculated for these ice configurations as geochronological data are not available for all glacier outlets. ELAs were estimated using the GIS tool of Pellitero et al. (2015), applying the area-altitude balance ratio method (AABR = 1.9 ± 0.81; Rea, 2009). ELAs are controlled by climate (Ohmura et al., 1992; Hughes and Braithwaite, 2008) but are also strongly influenced by non-climatic factors (Table 2), such as the supply of snow and ice from indirect sources (Mitchell, 1996; Kern and Laszlo, 2010). To assess the impact of 'redistributed' snow and ice, combined snow and avalanche contributing areas (A_c) were calculated (c.f. Ballantyne, 2007a,b; Barr et al., 2017; Dominant wind direction W/SW = 210 - 300°, Avalanche slopes ≥ 25°) and compared to total glacier surface areas (A_g). The A_c/A_g ratio is a proxy for the potential contribution of redistributed snow to glacier accumulation.

Results

Gaussian exposure age distributions for each site (Table 1) are in correct stratigraphic order in individual glacier catchments, are broadly consistent with comparable deglacial chronologies across the British Isles (Clark et al., 2012), and clearly differentiate cirque and valley moraines, with deposition during the Younger Dryas (YD; 11.7 - 12.9 ka) and Oldest Dryas respectively (GS-2.1a; 14.7 - 17.5 ka). Moreover, these datasets are chronologically robust (Kirkbride and Winkler, 2012), with well-dated moraine sequences in Glenmalur, Glendasan and Glenmacnass (Fig. 1), and provide a framework for a wider morphostratigraphic deglacial chronology for the Wicklow Mountains (Knight et al., 2017). At cirque sites, SHED indicates deglaciation by 12.31 ± 0.51 ka at Upper Lough Bray, 12.00 ± 0.44 ka at Kelly's Lough 11.40 ± 0.13 ka at Mullaghcleevaun and 10.93 ± 0.26 ka for the outer moraine at Lough Nahanagan (Colhoun and Synge, 1980). In contrast, valley moraines were deposited at 16.46 ± 0.58 ka at Glenmacnass Waterfall, 16.21 ± 0.60 ka at Upper Glendasan, 15.48 ± 0.35 ka at Carrawaystick Brook and 15.41 ± 0.30 ka at Lough Brook. Finally, SHED indicates the emergence of Carrigshouk by 16.64 ± 0.82 ka. This date provides a minimum age for wider summit deglaciation in the Wicklows Mountains due to its comparatively low elevation (571 m) and central position on the range divide.

Independent radiometric ages from Lough Nahanagan (^{14}C, ^{36}Cl, ^{10}Be) and Kelly's Lough (^{10}Be) can be used to verify the results of SHED (Colhoun and Synge, 1980; Bowen et al., 2002; Barth et al., 2018). Unfortunately, geochronological data from Lough Nahanagan are clearly conflicting and indicate moraine deposition at either 11.5 - 11.6 ka (n = 2; ^{14}C), 17.9 ± 1.0 ka (n = 1; ^{36}Cl) or between 9.7 - 21.7 ka (n = 3; ^{10}Be). While the limited number of samples (n = 6) prevents statistically robust identification and rejection of erroneous results (Tomkins et al., 2018b), and in turn, independent verification of SHED data at this site, the observed age scatter does highlight the importance of pre- or post-depositional processes at Lough Nahanagan, with uncertainty introduced by moraine stabilisation (Hallet and Putkonen, 1996), nuclide inheritance (Putkonen and Swanson, 2003), or a combination of both. At Kelly's Lough, ^{10}Be ages (n = 6; Barth et al., 2018) are also likely influenced by geological uncertainty and are non-normally distributed, with ages of 9.5 - 9.7 ka (n = 2), 11.2 - 11.7 ka (n = 3) and one outlier age of 137 ± 7 ka. This 'old' outlier does not conform to the 2σ test (Dortch et al., 2013) and matches the typical signature of nuclide inheritance observed in analysis of large ^{10}Be datasets (Dortch et al., 2013; Murari et al., 2014). Excluding this outlier returns a reduced dataset (n = 5) with a mean age of 10.6 ± 0.5 ka (Arithmetic mean ± Standard Error of the Mean; SEM) following the analytical steps of Barth et al. (2018) and standard procedures for interpreting moraine age information i.e. the timing of deglaciation is determined as the mean of moraine boulder ages (Briner et al., 2005). However, this interpretation is based on the assumption of rapid moraine stabilisation after ice retreat. Moraine exposure ages (e.g. ^{10}Be) relate to the emplacement or exhumation of surfaces and the onset of exposure to cosmic radiation (Gosse and Phillips, 2001). In many geomorphological settings, these events may be contemporaneous with glacial retreat but not necessarily so (c.f. Hallet and Putkonen, 1996). Thus, the distribution of boulder ages on a single moraine more accurately represents the process of moraine stabilisation (Putkonen and Swanson, 2003) and individual ages are best interpreted as minimum-limiting ages (Briner et al., 2005). As a result, the greatest boulder age is hypothesised to most closely match the true age of a moraine, although only under an assumption of no prior exposure (Putknonen and Swanson, 2003). Based on this reasoning, the oldest ages from Kelly's Lough are likely more representative of the timing of deglaciation, with moraine deposition at 11.44 ± 0.12 ka (Arithmetic mean of 3 oldest samples ± SEM) or more conservatively, at 11.65 ± 0.74 ka (Oldest sample; CRS-14-3; Barth et al., 2018). This interpretation is supported by PDE analysis (n = 5; Dortch et al., 2013) which returns a peak Gaussian exposure age distribution of 11.44 ± 0.51 ka. These estimates overlap within uncertainty with the SH Gaussian exposure age distribution of 12.00 ± 0.44 ka (Fig. 2C) and confirm deglaciation during the late-YD or early Holocene. These independent ^{10}Be ages join a growing body of evidence that SHED can generate accurate ages for glacial landforms (Rode and Kellerer-Pirklbauer, 2011; Ffoulkes and Harrison, 2014; Tomkins et al., 2016; 2018a; 2018b).

Reconstructed cirque glaciers (Table 2) range in size from 0.35 km^2 (ULB) to 1.10 km^2 (LN) while snow contributing areas (A_c) range from 0.12 km^2 (ULB) to 1.07 km^2 (LN). At Lough Nahanagan and Mullaghcleevaun, extensive upland plateaus to the west and south (210 - 300°) account for large snow contributing areas ($A_c \geq 1$ km^2). In contrast, restricted upslope areas within the glacier drainage basin likely limited the potential for significant snow redistribution ($A_c \leq 0.5$ km^2) at Kelly's Lough and Upper Lough Bray. AABR ELAs for cirque glaciers range from 513 m (LN) to 648 m (KL) and show no clear spatial clustering. Finally, reconstructed valley glaciers range in size from 4.96 km^2 (UGD) to 12.46 km^2 (GW) and demonstrate a progressive reduction in total glacier area (A_g) throughout the period 15.4 - 16.5 ka.

Discussion

Firstly, these data demonstrate that significant ice masses persisted in the Wicklow Mountains after the LGM, with large valley glaciers (Length: ~4 km, Area: ~12.5 km^2) present until ~16.5 ka; 4-7 ka after retreat of the ISIS (Smedley et al., 2017b). In contrast, lowland (23.1 ± 2.2 ka; Bowen et al., 2002) and summit deglaciation (n = 3; 21.0 – 21.9 ka; Ballantyne et al., 2006) was coeval with ISIS retreat (20.8 – 23.9 ka; Smedley et al., 2017b), SH exposure ages from the summit of Carrigshouk (571 m) indicate that summit ice fields were present on the range divide until 16.64 ± 0.82 ka. However, distal summits were ice free as early as 21.9 ± 1.1 ka (Djouce Mountain, 725 m), 21.2 ± 1.1 ka (Scarr, 641 m) and 21.0 ± 1.1 ka (Kanturk, 523 m) and evidence a significant time lag in summit deglaciation (~4.4 ka). Collectively, ^{10}Be and SHED ages indicate rapid downwastage of the Wicklow Ice Cap soon after the LGM and a transition to summit ice fields which sourced discrete outlet glaciers (e.g. Glenmacnass, Glendasan, Glenmalur; Fig. 1); some of which persisted until at least ~15.4 ka and likely through until the onset of Greenland Interstadial 1 (GI-1; 12.9 - 14.7 ka;

Rasmussen et al., 2014). Deglaciation of Carrigshouk at ~16.6 ka marks a shift to topographically confined ice flow, with glaciers sourced from high elevation cirques, and likely reflects a time-progressive response to reduced moisture availability and winter aridity during this interval (Kelly et al., 2010).

This pattern of ice retreat, involving post-LGM downwastage of the ice cap and a transition to alpine-style valley glaciation, is consistent with numerical ages and geomorphological evidence from comparable mountain ice caps in the Irish Sea Basin and across Ireland. In Wales, summit ^{10}Be ages record rapid and spatially uniform downwastage of the Welsh Ice Cap soon after the LGM, with summits (≥ 600 m) exposed as nunataks at 19 - 20 ka (Hughes et al., 2016). This timeframe accords with summit ^{10}Be ages from Wicklow (Ballantyne et al., 2006) and the Blackstairs Mountains (Ballantyne and Stone, 2015). However, ^{14}C ages from proximal Welsh lowlands (15.82 ± 0.39 cal. ka BP; Lowe, 1981; Reimer et al., 2013) show that large alpine-style valley glaciers (Length: ~3.5 km), likely sourced from high elevation cirques, were present for ~4 ka after initial summit emergence (Hughes et al., 2016). This retreat history is matched by ^{10}Be and ^{36}Cl ages from the Lake District, which record substantial downwastage (< 750 m) of the Lake District ice cap on the Scafell massif by 17.3 ± 1.1 ka (^{36}Cl; Ballantyne et al., 2009). Despite this, a large valley glacier (Length: ~5 km), sourced from Scafell, was still present in Wasdale until 16.7 ± 0.9 ka (^{10}Be; McCarroll et al., 2010). In the central Lake District, a moraine age from Watendlath extends this period of alpine-style glaciation until 15.2 ± 0.9 ka (^{10}Be; Wilson et al., 2013). Collectively, these numerical ages show that while ice caps in the Irish Sea Basin underwent significant downwastage after the LGM, large valley glaciers persisted throughout the post-LGM period; likely until the onset of GI-1.

This pattern of ice retreat is also consistent with geomorphological evidence from the Kerry-Cork ice cap (KCIC) in SW Ireland (Ballantyne et al., 2007; Barth et al., 2016) although ^{10}Be ages from cirque moraines in the Macgillycuddy's Reeks indicate that extensive (i.e., ice sheet and ice cap) glaciation had terminated by 24.5 ± 1.4 ka (Barth et al., 2016). This timeframe is significantly earlier than comparable ice caps in Wicklow and throughout the Irish Sea Basin (Ballantyne et al., 2009; Hughes et al., 2016). However, cirque glaciers were present as recently as 16.7 - 16.9 ka (^{10}Be; n = 2; Harrison et al., 2010). While there is ongoing debate regarding the chronology of glaciation in SW Ireland (Barth et al., 2016; Knight, 2016) and the extent and configuration of the IIS and the KCIC during and after the LGM (Anderson et al., 1998; Rae et al., 2004; Harrison et al., 2010; Ballantyne et al., 2011), numerical ages from cirque moraines in the Macgillycuddy's Reeks (Harrison et al., 2010) support a prolonged period of post-LGM mountain glaciation, consistent with SHED, ^{10}Be, ^{36}Cl and ^{14}C ages in other mountain massifs. These data support a period of renewed or continuous mountain glaciation after the LGM, with significant ice masses (Length: ≤ 5 km) recorded in the Wicklow Mountains at 15.4 - 16.5 ka, in Wales until 15.82 ± 0.39 cal. ka BP (Lowe, 1981), in the Lake District at 15.2 - 16.7 ka (Ballantyne et al., 2009; Wilson et al., 2013) and in SW Ireland at 16.7 - 16.9 ka (Harrison et al., 2010). These data represent a growing body evidence for substantial glaciation during the post-LGM period and support a model of gradual, oscillatory retreat of mountain glaciers after the LGM. Moreover, these numerical ages accord with wider evidence for post-LGM disintegration of the BIIS into component ice caps (Clark et al., 2012).

Secondly, the geomorphological record indicates that post-LGM deglaciation involved numerous oscillations of glacier termini during the long-term retreat phase (~8 ka), with valley and cirque moraines deposited during the Oldest Dryas and Younger Dryas respectively (Fig. 2A). Correlative Gaussian exposure age distributions from valley moraines across the mountain range are indicative of a period of widespread moraine deposition, related to stabilisation or re-advance of valley glaciers at 15.4 - 16.5 ka. Ice-marginal moraines provide direct evidence of former ice margin positions (Svendsen et al., 2004) but determining whether glaciers are stationary or re-advancing cannot be determined solely from moraine chronology. Moreover, while moraines can be used as indirect proxies for palaeoclimate (Benn and Ballantyne, 2005), a multitude of non-climatic factors can also influence patterns of moraine distribution, formation and preservation (Barr and Lovell, 2014), and therefore introduce complexity to links between periods of glacial deposition and wider climatic trends (Blaauw et al., 2007). However, the interval 15.4 - 16.5 ka is coeval with the peak ice rafted debris flux (Bard et al., 2000; Eynaud et al., 2009) and reduced sea surface temperatures (Bard et al., 2000) during Heinrich Stadial 1 (Fig. 2B; HS1) and the re-advance of the Irish Ice Sheet (IIS) and the ISIS during the Killard Point Stadial (~16 - 17.1 ka; McCabe et al., 2007; Clark et al., 2012). This period of glacier stabilisation or re-advance in Ireland during the Oldest Dryas was matched further down the North-East Atlantic margin in Spain (Palacios et al., 2017) and could reflect a direct response to North Atlantic climate perturbations (HS1) with short-term oscillations of the ice front (≤ 1 ka) during the long-term post-LGM retreat phase. These chronological data match recent morphostratigraphic assessments of glacial geomorphology in the Wicklow Mountains which support a widespread pattern of sustained retreat interrupted by minor glacier readvance (Knight et al., 2017). Valley glacier retreat was synchronous across the Wicklow Mountains, as demonstrated by progressive deglaciation from low to high elevation (Fig. 3A; $R^2 = 0.9116$; $p = 0.045$). These data are indicative of climate-controlled retreat with independent outlet glaciers responding synchronously to reduced moisture availability (Kelly et al., 2010), irrespective of contrasting glacier aspects, source areas or glacier extents. Oldest Dryas valley glaciers were extensive (≤ 12.5 km^2), sustained by ice fields and prior to ~16.6 ka, overtopped low-lying summits (~571 m). As a result, the potential for significant redistribution of snow and avalanche material (Ac > Ag) was limited, particularly during periods of winter aridity (Kelly et al., 2010). Therefore, while topography likely influenced the retreat pattern in individual valleys, post-LGM retreat (~15 - 17 ka) was primarily driven by climate.

In contrast, marked asynchroneity in the timing of final YD deglaciation (Fig. 3; 11.4 - 12.3 ka) is unrelated to cirque elevation (Fig. 3A; $R^2 < 0.01$, $p = 0.97$), palaeo-ELA (Fig. 3E; $R^2 = 0.04$, $p = 0.81$) or site latitude ($R^2 = 0.10$, $p = 0.69$). If regional climate was the primary control on cirque glacier survival, then the timing of deglaciation would be expected to: (1) correlate with elevation, (2) be ELA dependent, or (3) show some relationship with temperature as a function of site latitude. These variables show little or negligible correlation with SH derived deglacial ages ($R^2 \leq 0.1$) and are not statistically significant at $p = 0.05$ ($p \geq 0.69$). These data suggest that climate was not the dominant control on the timing of final YD deglaciation. Instead, glacier retreat was strongly controlled by local topography and the redistribution of wind-blown snow and avalanche material (Fig. 3B; $R^2 > 0.99$, $p < 0.01$). Combined snow and avalanche contributing areas (A_c) range from just 0.119 km^2 at Upper Lough Bray to 1.071 km^2 at Lough Nahanagan. For glaciers with large A_c areas, topography may exert the primary control on glacier formation and survival, and may account for the comparatively late-deglaciation of Lough Nahanagan and Mullaghcleevaun during the early-Holocene. By comparison, glaciers with small A_c areas, where the potential for redistribution of snow and avalanche material is limited, may respond quasi-synchronously to climate warming. For example, the early deglaciation of Upper Lough Bray at 12.31 ± 0.51 ka is coeval with a gradual rise is summer air temperatures after ~12.5 ka (Brooks and Birks, 2000) which was likely sufficient to raise the 'climatic' ELA above cirque elevations ($A_c/A_g = 0$; Barr et al., 2017) and initiate mass wastage. In contrast, abundant snow redistribution ($A_c \geq 1$), conditioned by existing topographic configurations (extensive upland plateaus), was likely sufficient to locally supress the 'local' (non-climatic) ELA and promote glacier survival at other sites. However, the contribution of redistributed snow to glacier accumulation almost certainly diminished throughout the YD as summer air temperatures increased rapidly towards the onset of the Holocene (Brooks and Birks, 2000), thus limiting snowpack preservation. These data indicate that while regional climate provides the baseline conditions for glacier growth and decay, cirque glacier oscillations may primarily reflect the influence of topography. In this scenario, macro-topographic configurations condition glaciers to be sensitive to cold and/or wet climate and provide a first-order control on glaciation by (1) facilitating glacier initiation, and (2) enabling a lagged response to warming climate.

However, there is a weak correlation between glacier size (A_g) and deglaciation age (Fig. 3C; R^2 = 0.77, p = 0.12), although the size variation between the smallest (ULB; 0.35 km^2) and largest reconstructed glacier (LN: 1.10 km^2) is minimal (~0.75 km^2). As such, significant within-mountain range variation in glacier response times is not anticipated (Raper and Braithwaite, 2009). This may account for the weak correlation between A_c/A_g ratios and deglaciation ages (Fig. 3D; R^2 = 0.58, p = 0.24) although the observable trend demonstrates the probability of early deglaciation for glaciers with small A_c/A_g ratios. Based on this reasoning, we conclude that for small YD glaciers, local topoclimatic controls can be more significant than wider regional climate in determining cirque glacier functioning, and in particular, the timing of final deglaciation. Avalanches accumulation area and deglaciation age correlations have important implications for palaeoclimate reconstructions based on dating of cirque moraines (e.g. using ^{10}Be or SHED), as cirque glacier ELAs can be non-representative of the regional climate and consequently glacier dynamics are likely to be decoupled from climatic changes occurring in the North Atlantic region. This phenomenon is observed today in the behaviour of small glaciers in marginal glaciated settings such as the Italian Alps (Colucci, 2016) and other Mediterranean mountains (Hughes, 2018), which demonstrates the sensitivity of glaciers to macro-topography (Mitchell, 1996; Allen, 1998; Benn and Lehmkuhl, 2000; García-Ruiz et al., 2000; López-Moreno et al., 2006a, 2006b; Kern and Laszlo, 2010), particularly in marginal glaciated regions (Chueca and Julián, 2004; Mills et al., 2009). These data from Ireland also show that the impact of topography on glacier functioning is most significant when glaciers are small (≤ 1 km^2), resulting in clear asynchroneity in deglaciation (Fig. 3A; R^2 = < 0.01), and provide further evidence that the climatic integrity of cirque glaciers may be limited (Kirkbride and Winkler, 2012). In contrast, large glaciers (~12.5 km^2), with limited potential for snow redistribution, have been shown to respond synchronously to climate forcing (Fig. 3A; R^2 = 0.9116). Modelling studies have shown that topoclimatic variables (solar radiation/snow redistribution) can predict the style of deglaciation (moraine distribution) for small Younger Dryas glaciers (Coleman et al., 2009; Bickerdike et al., 2017). The chronological data presented here provides new evidence that topographic controls not only influence the style of deglaciation, but can determine the timing of final deglaciation, with clear within-mountain range variability.

A further challenge in linking cirque glacier oscillations to climatic fluctuations is the potential impact of moraine stabilisation (Hallet and Putknonen, 1996). This post-depositional process can result in moraine ages (e.g. ^{10}Be, SHED) which post-date glacial retreat. A 1-2 ka early stabilization period has been recorded for Alpine moraine sequences in Alaska and the Alps (Briner et al., 2005; Ivy-Ochs et al., 2006, 2008; Dortch et al., 2010a). In high-mountain and alpine environments, glaciers can produce distinctive asymmetric ice-contact fans that undergo rapid gullying and post-depositional reworking on their ice-proximal slopes (e.g. Hambrey et al., 2008; Lukas et al., 2012). However, these landforms are topographically and sedimentologically distinct from the low-relief, topographically concordant valley and cirque landsystems found in the Wicklow Mountains. Moreover, there has been comparatively little attention on the processes of moraine development in these low-relief environments, with analogue studies predominantly focused on 'hummocky moraine' landsystems (e.g. Benn and Lukas, 2006). In addition, there has been a relative paucity of research into moraine processes in smaller cirque type landsystems, likely under the assumption of rapid stabilisation after deglaciation. Recent work has highlighted the importance of self-censoring in cirque and valley environments due to obliterative overlap (Barr and Lovell, 2014) or ice-cored moraine degradation (Crump et al., 2017; Tonkin et al., 2017), while external-censoring due to slope instability may also provide a control on moraine stabilisation (Barr and Lovell, 2014). As a result, there is considerable uncertainty regarding the robustness of chronological datasets for cirque moraine systems (Kirkbride and Winkler, 2012).

To produce better-resolved glacier chronologies, researchers can account for moraine stabilisation through (1) morphostratigraphic comparison of moraine sequences, (2) Gaussian separation of exposure ages (Dortch et al., 2013), (3) assessment against independent geochronological data (e.g. ^{14}C) and (4) consideration of modern process studies and likely modern analogues for moraine assemblages when assessing site suitability for a give geochronological approach (e.g. Çiner et al. 2015). Based on these criteria, we infer that moraine stabilisation may be a key post-depositional process for the outer moraine at Lough Nahanagan. Firstly, this moraine is degraded (Colhoun and Synge, 1980) and is morphologically distinct from sampled cirque moraines at Mullaghcleevaun, Kelly's Lough and Upper Lough Bray which are tall (> 3 m), matrix-poor, boulder-rich, and feature clearly defined moraine crests (Fig. 4). The morphology of these large terminal moraines likely reflects the incorporation of Lateglacial rock slope failure (RSF) debris (Ballantyne et al., 2013). This debris may account for a significant proportion of the sediment budget of YD glaciers, particularly in cirques characterised by steep headwalls where RSF activity is enhanced.

Secondly, Gaussian separation of SHED data for Lough Nahanagan (n = 20) reveals a clear 'two-peak' probability density function (Fig. 2C), with Gaussian exposure age distributions of 10.93 ± 0.26 ka (n = 9) and 11.38 ± 0.26 ka (n = 9). The youngest age post-dates the YD by ~0.8 ka and is inconsistent with wider evidence for deglaciation of the British Isles by the YD/Holocene transition (MacLeod et al., 2011) although in isolation, this observation is insufficient to reject this age at this stage. In addition, this method highlights clear outlier ages (n = 3; 12.8 - 13.5 ka). These Gaussian distributions can be rejected as they are comprised of fewer than 3 ages (c.f. Fig. 3 in Dortch et al., 2013). Independent ^{14}C ages indicate deglaciation during the late YD and early Holocene (11.5 - 11.6 ka; Colhoun and Synge, 1980) while a single ^{36}Cl age suggests ice free conditions since 17.9 ± 1.0 ka (Bowen et al., 2002), although this age likely reflects prior exposure (*inheritance*) and is rejected from further analysis. New ^{10}Be ages from the outer moraine at Lough Nahanagan (Barth et al., 2018) are not internally consistent (n = 3; 9.7 - 21.7 ka) and are inconclusive with regards to the timing of deglaciation at this site. Based on these data, we conclude that the older Gaussian exposure age distribution of 11.38 ± 0.26 ka is more representative of final deglaciation as this age is consistent with previous ^{14}C ages and accounts for both the distinctive geomorphological assemblage at this site and the clear 'two-peak' distribution of SHED ages. This conclusion indicates that moraine stabilisation and boulder exhumation may account for the degraded moraine surface and comparatively '*young*' SHED ages.

These data provide further evidence that moraine ages are more likely influenced by post-glacial instability than prior exposure (Shanahan and Zreda 2000; Putkonen and Swanson 2003; Zech et al. 2005; Heyman et al. 2011; Applegate et al., 2012). As a result, the growth or decay of small cirque glaciers (< 1 km^2), as determined by radiometric methods (^{10}Be), may not only primarily reflect topographic controls, but may be profoundly influenced by post-depositional processes. The post-depositional evolution of YD moraine systems is largely unexplored at present and a clear co-benefit of the SHED approach is the insight it provides into these processes; insight that is not readily afforded by other geochronological approaches. Future research should carefully consider landform context (Barr and Lovell, 2014) and prioritise sampling of cirque environments where snow and avalanche contributing areas (A_c) are small (Warren, 1991; Mills et al., 2012; Barr and Lovell, 2014), where postglacial erosion is limited and where short transport distances promote the formation of matrix-poor boulder-rich moraines (Fig. 5; Pallàs et al., 2010). In these environments, snow redistribution is limited and moraines are more likely to stabilise rapidly after deglaciation. As such, these glaciers may respond quasi-synchronously to climatic fluctuations and may produce more robust palaeoclimatic reconstructions.

Conclusions

This study provides the first comprehensive glacial retreat history for the Wicklow Mountains, Ireland. 170 Schmidt Hammer exposure ages from cirque and valley moraines and from a summit overridden by ice at the LGM demonstrate that significant ice masses persisted for 4-7 ka after retreat of the Irish Sea Ice Steam and were sustained by summit ice-fields until ~16.6 ka. Post-LGM retreat involved numerous oscillations of glacier termini during the retreat phase, with widespread moraine deposition related to stabilisation or re-advance of valley glaciers during the Oldest Dryas, potentially in response to cooling during Heinrich Stadial 1 (HS1). However, these moraines reflect short-term oscillations (≤ 1 ka) of the ice front during the long-term retreat phase (~8 ka), which was driven by reduced moisture availability and winter aridity. These data match numerical ages (^{10}Be, ^{36}Cl, ^{14}C) from comparable mountains caps at the margins of the Irish Sea basin and in SW Ireland which support a model of widespread and persistent alpine glaciation during the post-LGM period. Significant ice masses (Length: ≤ 5 km) were present until the onset of Greenland Interstadial 1 and in some mountain massifs, for up to ~4 ka after initial summit emergence following the LGM. Valley glacier retreat in the Wicklow Mountains was driven by climate, with time-progressive deglaciation from low to high elevation ($R^2 = 0.9116$). In contrast, marked asynchroneity in the timing of Younger Dryas (YD) deglaciation (11.4 - 12.3 ka), unrelated to site elevation, latitude or equilibrium line altitude (ELA), is accounted for by macro-topography and the redistribution of snow and avalanche material, sufficient to locally supress the 'local' (non-climatic) ELA and promote glacier survival. Contrasting synchroneity in the timing of glacial retreat during these periods is conditioned by glacier size, with small YD glaciers (< 1 km^2) highly sensitive to local topographic controls. This result has important implications for palaeoclimate reconstructions based on dating of cirque moraines (e.g. ^{10}Be, SHED), as cirque glacier dynamics may be (at least partly) decoupled from climate. This is further complicated by post-depositional processes which can result in ages which post-date retreat. As a result, future palaeoclimate reconstructions should prioritise cirques where snow and avalanche contributing areas (A_c) are small and where the potential for post-depositional disturbance is limited (matrix-poor, boulder rich moraines).

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Figures

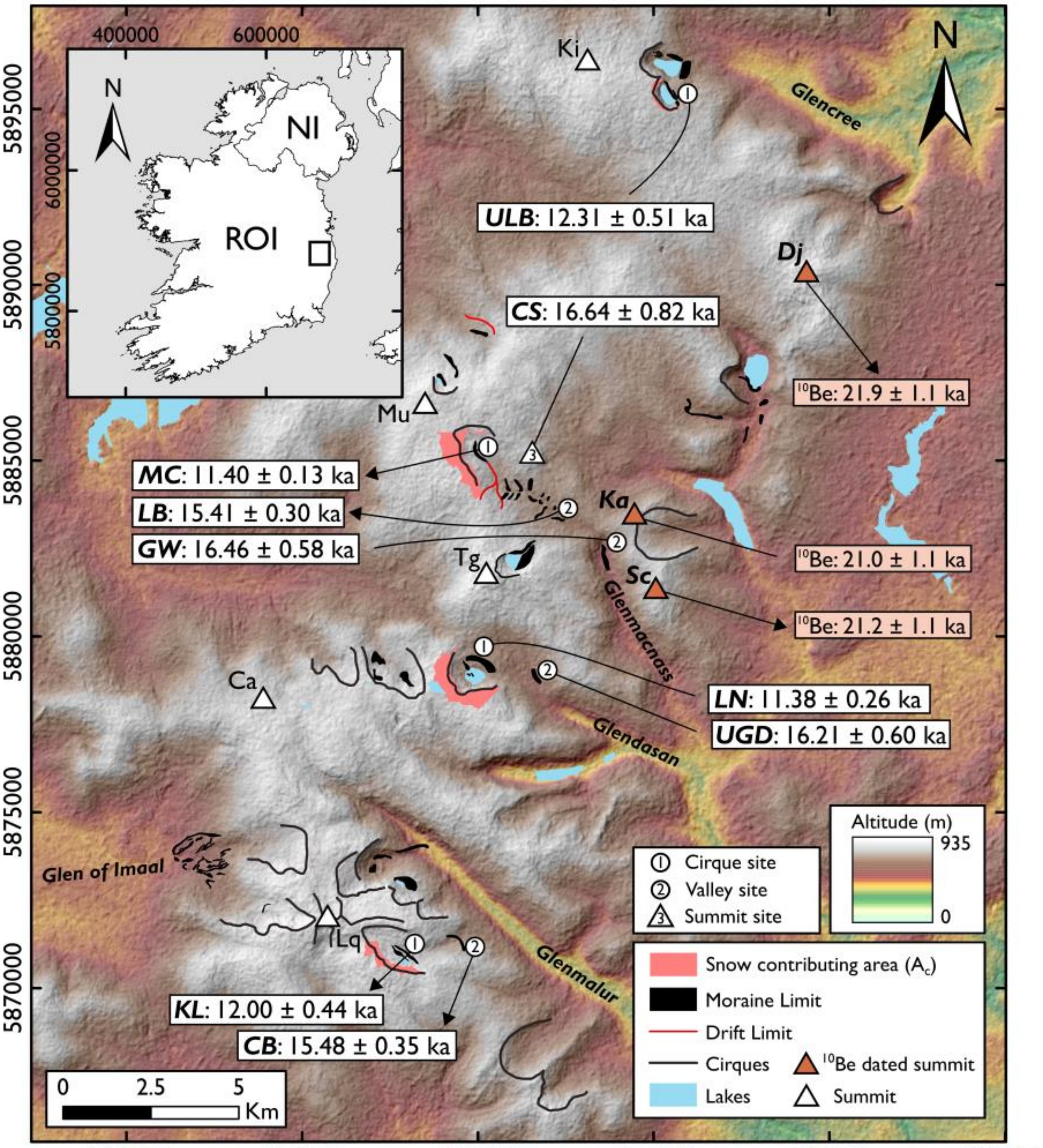

Figure 1. Generalised geomorphological map of the Wicklow Mountains. Moraines modified after Clark et al. (2017). ^{10}Be ages recalibrated from Ballantyne et al. (2006) using the online calculators formerly known as the CRONUS-Earth online calculator (Wrapper script 2.3, Main calculator 2.1, constants 2.3, muons 1.1; Balco et al., 2008) based on the Loch Lomond Production Rate (Fabel et al., 2012), the time-independent Lm scaling (Lal, 1991; Stone 2000) and assuming 0 mm ka^{-1} erosion. Ca: Camenabologue; Dj: Djouce Mountain; Ka: Kanturk; Ki: Kippure; Lq: Lugnaquillia; Mu: Mullaghcleevaun; Sc: Scarr; Tg: Tonelagee. Map projection: UTM WGS 1984.

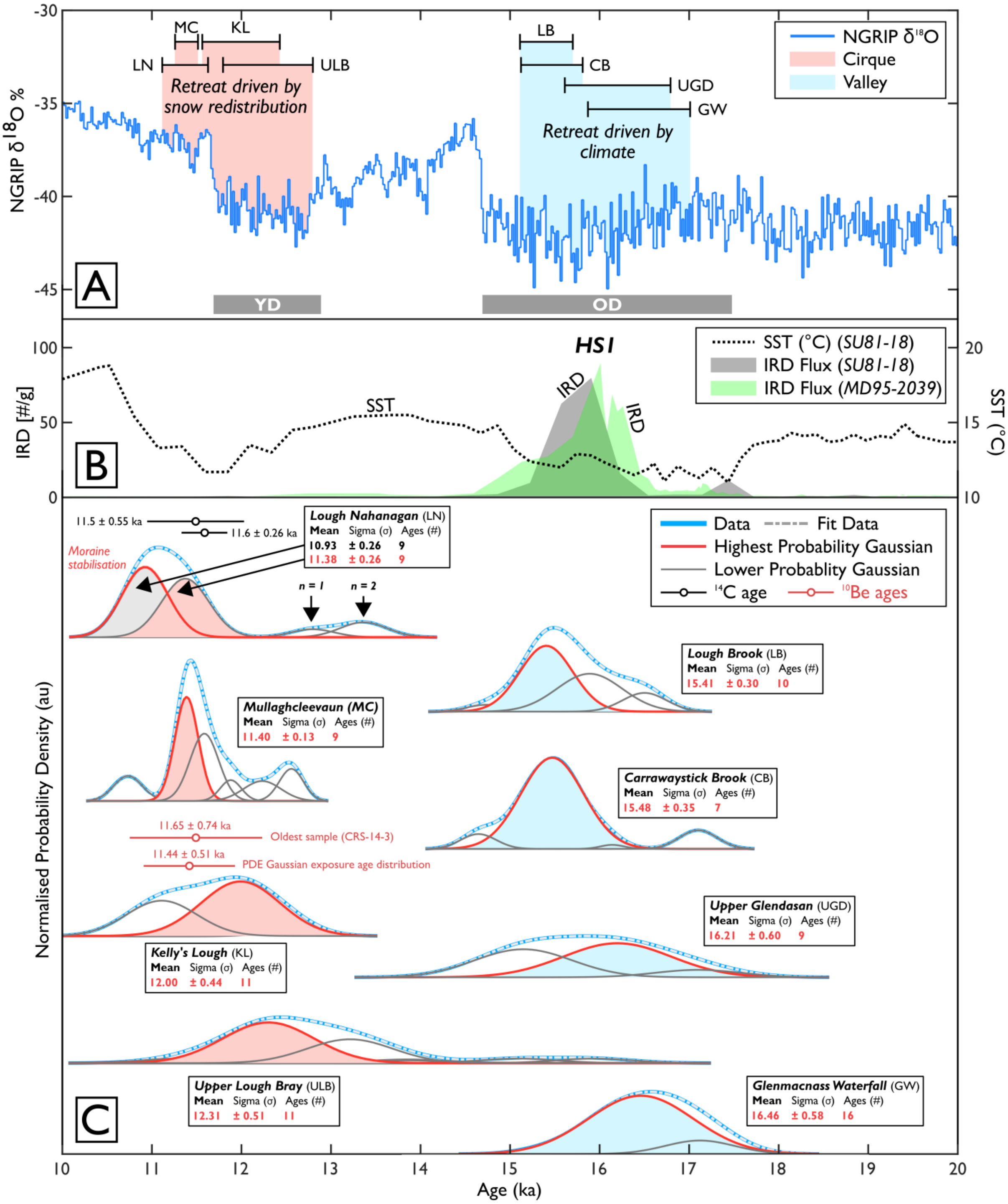

Figure 2. Gaussian ages related to deglaciation of the Wicklow Mountains. A: The NGRIP Oxygen Isotope Curve (Rasmussen et al., 2014) plotted against 1σ age boundaries for sampled cirque and valley sites. The Younger Dryas (YD) and Oldest Dryas (OD) periods are marked. B: Ice rafted debris (#/g) and sea surface temperature (°C) records from cores SUBI-18 (Bard et al., 2000) and MD95-2039 (Eynaud et al., 2009) in the North Atlantic. C: Gaussian models for sampled cirque and valley sites. For each site, the highest probability Gaussian is considered the most likely timing of deglaciation as all ages are younger than the Last Glacial Maximum (Dortch et al., 2013). At Lough Nahanagan, the oldest peak with more than 3 ages is selected (c.f. Dortch et al., 2013) as this moraine is degraded and morphologically distinct from other sampled cirque moraines (MC, KL, ULB). Moreover, this estimates matches previous ^{14}C ages (Colhoun and Synge, 1980). At Kelly's

Lough, the peak SHED Gaussian exposure age distribution is matched by independent [10]Be ages (Barth et al., 2018). This estimate overlaps with the peak [10]Be Gaussian exposure age distribution of 11.44 ± 0.51 ka (excluding outlier CRS-14-5d: 137 ± 7 ka) and more conservatively, the oldest sample of 11.65 ± 0.74 ka (CRS-14-3).

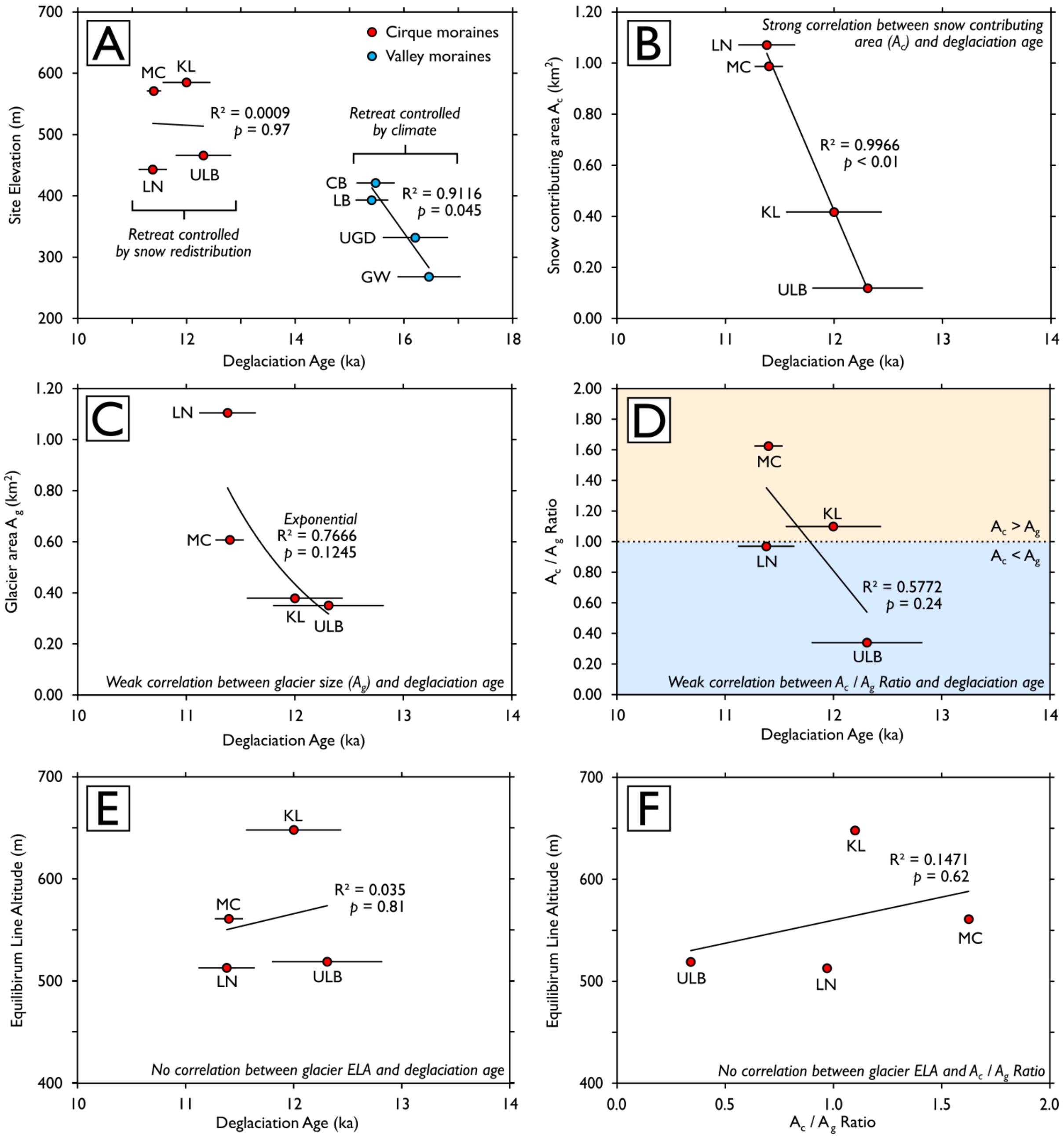

Figure 3. Topographic and climatic controls on the timing of cirque deglaciation in the Wicklow Mountains. A: Site Elevation; B: Snow contributing area (A_c); C: Glacier area (A_g); D: A_c/A_g ratio; E: ELA; F: ELAs and A_c/A_g ratio plots. These data show that for large valley glaciers, retreat is driven by climate with progressive deglaciation from low to high elevation (A). In contrast, marked asynchroneity in the timing of cirque deglaciation (A) is strongly controlled by snow redistribution (B). This asynchroneity is weakly correlated with glacier size (C) and A_c/A_g ratios (D) and is unrelated to ELA (E).

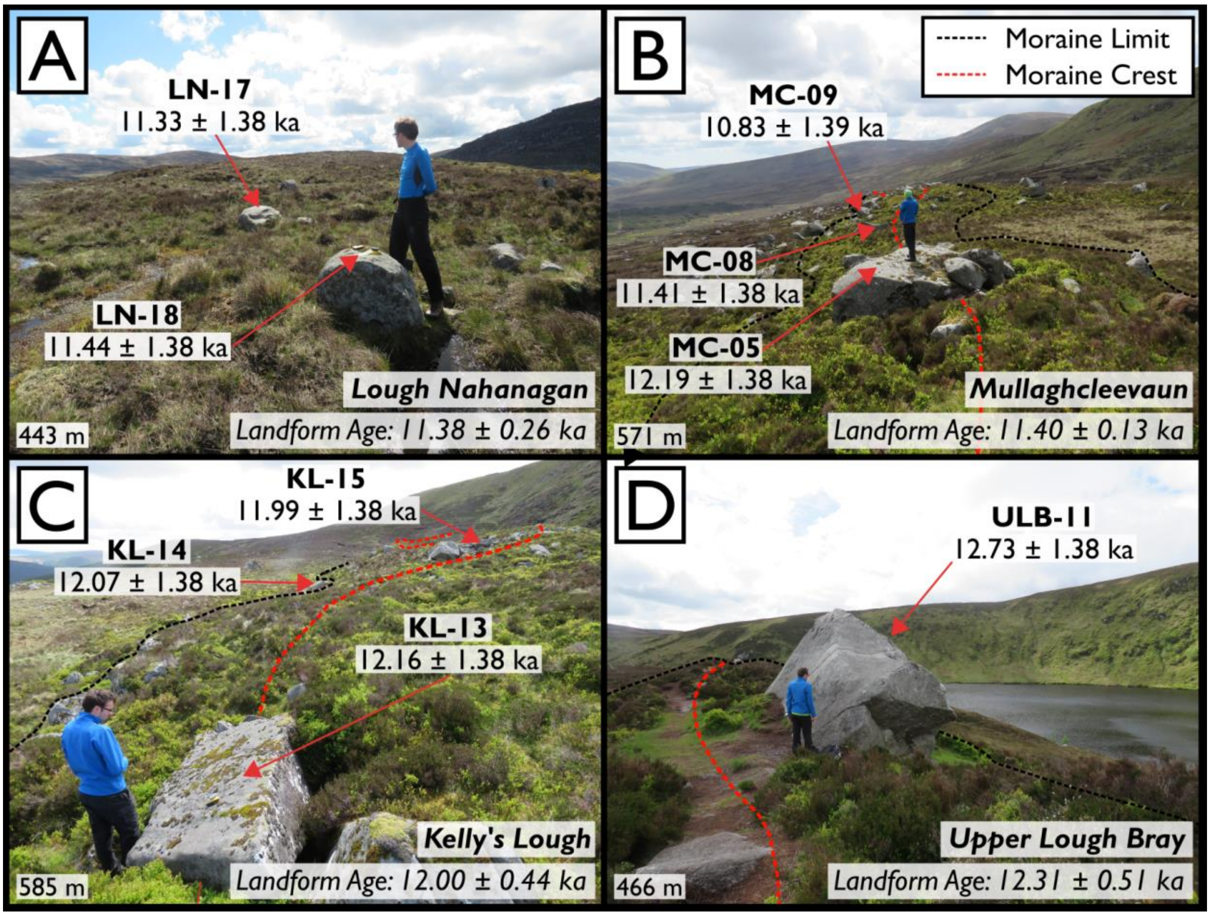

Figure 4. Sampled cirque moraines at Lough Nahanagan (A), Mullaghcleevaun (B), Kelly's Lough (C) and Upper Lough Bray (D). The outer cirque moraine at Lough Nahanagan is degraded (Colhoun and Synge, 1980) and is morphologically distinct from other cirque moraines which are sharp crested, boulder-rich and matrix-poor.

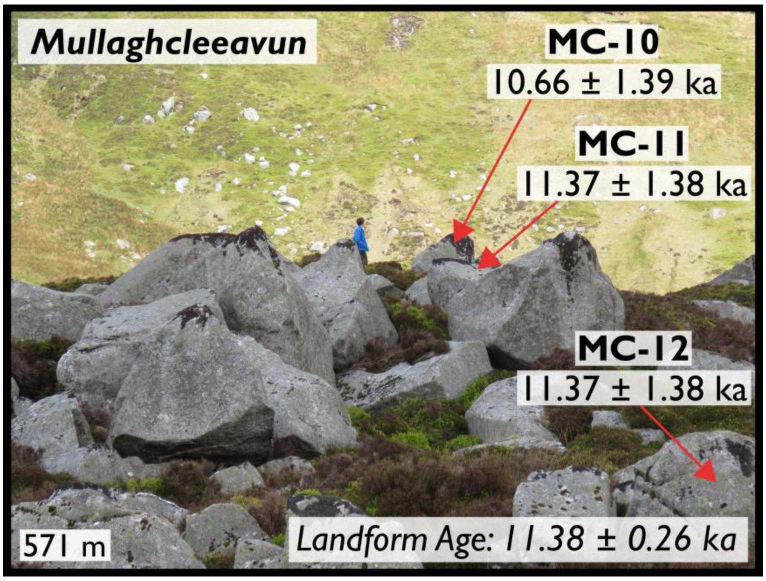

Figure 5. Matrix-poor, boulder-rich moraine at Mullaghcleevaun which likely stabilised rapidly after deglaciation.

Supplementary dataset

Supplementary dataset. Sample information for boulder and bedrock surfaces sampled in the Wicklow Mountains (n = 170). Reported R-values are the arithmetic mean of 30 measurements ± the standard error of the mean (SEM). Reported ages (ka) were calculated using the SHED-Earth online calculator (http://shed.earth; Tomkins et al., 2018a) based on the Loch Lomond Production Rate (LLPR; Fabel et al., 2012), the time-dependent Lm scaling (Lal, 1991; Stone; 2000) and assuming 0 mm ka^{-1} erosion. These numerical ages will be subjected to recalibration in light of future refinement of ^{10}Be production rates. Available here: https://www.researchgate.net/publication/325120644_Tomkins_JQS_Supplementary_Table

Tables

Table 1. Gaussian ages for cirque, valley and summit sites from the Wicklow Mountains.

Group	Site Name	Site Code	Site Elevation (m)	Site Latitude (°)	Deglaciation Age (ka)	±	Glacier Area (A_g km^2)
Cirque	Lough Nahanagan	LN	443	53.034	11.38	0.26	1.10
	Mullaghcleevaun	MC	571	53.089	11.40	0.13	0.61
	Kelly's Lough	KL	585	52.960	12.00	0.44	0.38
	Upper Lough Bray	ULB	466	53.178	12.31	0.51	0.35
Valley	Lough Brook	LB	393	53.070	15.41	0.30	7.63
	Carrawaystick Brook	CB	421	52.964	15.48	0.35	1.83
	Upper Glendasan	UGD	332	53.030	16.21	0.60	4.96
	Glenmacnass Waterfall	GW	268	53.062	16.46	0.58	12.46
Summit	Carrigshouk	CG	571	53.086	16.64	0.82	-

Table 2. Snow contributing areas (A_c), glacier areas (A_g) and ELAs for cirque moraines.

Site Code	Deglaciation Age (ka)	±	Snow Contributing Area (A_c km^2) [a]	Glacier Area (A_g km^2)	A_c / A_g Ratio	ELA [b]
LN	11.38	0.26	1.07	1.10	0.97	513
MC	11.40	0.13	0.99	0.61	1.62	561
KL	12.00	0.44	0.42	0.38	1.10	648
ULB	12.31	0.51	0.12	0.35	0.34	519

[a] Area within the glacier drainage basin within the 210 – 300° quadrant + all other slopes which overlook the glacier (Gradients > 25°), [b] AABR = 1.9 ± 0.81 (Rea, 2009)