Cosmogenic ^{10}Be chronology of the last deglaciation of western Ireland, and implications for sensitivity of the Irish Ice Sheet to climate change

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ABSTRACT

Accelerator mass spectrometry (AMS) ^{14}C dates of fossiliferous marine mud identify a readvance of the Irish Ice Sheet from the north and central lowlands of Ireland into the northern Irish Sea Basin during the Killard Point Stadial at ca. 16.5 cal k.y. B.P., with subsequent deglaciation occurring by ca. 15.0–15.5 cal k.y. B.P. Killard Point Stadial moraines have been mapped elsewhere in Ireland but have previously remained undated. Here, we report sixteen ^{10}Be surface exposure dates that constrain the age of retreat of the Killard Point Stadial ice margin from western Ireland. Eight ^{10}Be dates from the Ox Mountains (13.9–18.1 ka) indicate that final deposition of the moraine occurred at 15.6 ± 0.5 ka (mean age, standard error). Eight ^{10}Be dates from Furnace Lough (14.1–17.3 ka, mean age of 15.6 ± 0.4 ka) are statistically indistinguishable from the Ox Mountain samples, suggesting that the moraines were deposited during the same glacial event. Given the agreement between the two age groups, and their common association with a regionally significant moraine system, we combine them to derive a mean age of 15.6 ± 0.3 ka (15.6 ± 1.0 ka with external uncertainty). This age is in excellent agreement with the timing of deglaciation from the Irish Sea Basin (at or older than 15.3 ± 0.2 cal k.y. B.P.) and suggests the onset of near-contemporaneous retreat of the Irish Ice Sheet from its maximum Killard Point Stadial limit. A reconstruction of the ice surface indicates that the Irish Ice Sheet reached a maximum surface elevation of ~500 m over the central Irish Lowlands during the Killard Point Stadial, suggesting a high sensitivity of the ice sheet to small changes in climate.

Keywords: glacial geology, geochronology, cosmogenic dating, Irish Ice Sheet.

INTRODUCTION

Recently developed ^{14}C chronologies for the British-Irish Ice Sheet have demonstrated that the ice sheet responded to abrupt climate change during the last glaciation and deglaciation (McCabe and Clark, 1998, 2003; McCabe et al., 2005). The best-dated ice-sheet fluctuation occurred during the last deglaciation, when the British-Irish Ice Sheet margin readvanced across the northern and central lowlands of Ireland and into the northern Irish Sea Basin during the Killard Point Stadial, ca. 16.5 cal k.y. B.P., followed by rapid retreat sometime after 15.3 ± 0.2 cal k.y. B.P. (all radiocarbon ages calibrated with Calib 5.0.2; http://calib.qub.ac.uk) (McCabe and Clark, 1998; McCabe et al., 2005, 2007). This terrestrial record indicates that the British-Irish Ice Sheet readvanced in association with Heinrich event 1, consistent with response of a small ice sheet in this climatically sensitive region to North Atlantic cooling.

Opportunities for developing such chronologies, however, are restricted to sites where fossil-bearing marine sediments occur in association with glaciogenic sediments. Additional dating is thus required to determine whether terrestrial sectors of the British-Irish Ice Sheet fluctuated during the last deglaciation and ascertain their relation to those dated by ^{14}C of marine fossils. Surface exposure dating using in situ cosmogenic nuclides that are produced in glacial boulders provides such an opportunity for developing chronologies of terrestrial ice margins with millennial-scale resolution (Gosse et al., 1995; Licciardi et al., 2001, 2004; Kerschner et al., 2006; Rinterknecht et al., 2006; Schaefer et al., 2006).

Bowen et al. (2002) reported the first cosmogenic nuclide ages from glacial erratics associated with the British-Irish Ice Sheet. Their ^{36}Cl chronology identified several glacial events with ages closely corresponding to the ages of Heinrich events. Thus, similar to the ^{14}C chronology,

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these data indicate a close relationship between Heinrich events and related effects of North Atlantic climate on the behavior of the British-Irish Ice Sheet. Additional ^{36}Cl ages with external uncertainty (including the uncertainty of the ^{36}Cl production rate used) are needed, however, in order to further evaluate this relationship.

Elsewhere in Ireland, regionally extensive moraine systems have been identified that record a readvance or stillstand of the British-Irish Ice Sheet margin following the Last Glacial Maximum advance onto the continental shelf (Fig. 1). Based on their common association with contiguous and coeval bed form patterns that mark the most recent ice flow across the northern Irish Lowlands, McCabe et al. (1998) proposed that these moraines are correlative and correspond to readvance of the Irish sector of the British-Irish Ice Sheet during the Killard Point Stadial. Here, we test this hypothesis by using in situ ^{10}Be to date glacial erratics from a regionally significant moraine system located in the west of Ireland (Fig. 1). The results of this study provide an improved deglacial chronology of the western margin of the British-Irish Ice Sheet in Ireland, as well as help address ice-sheet sensitivity to millennial-scale climate changes in the North Atlantic during the last glaciation.

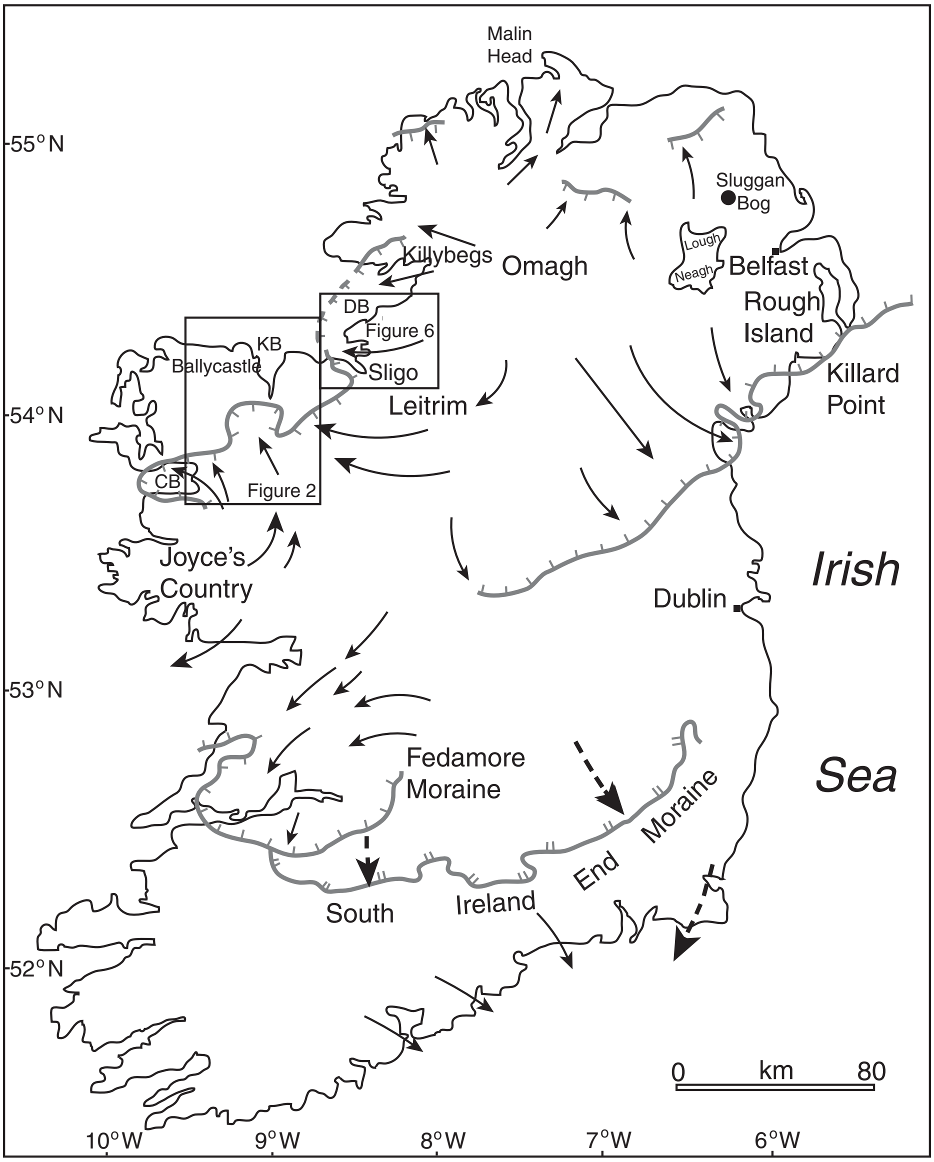

Figure 1. Map of Ireland showing general ice-flow patterns and ice limits associated with the Killard Point Stadial (lines with single hachures) and earlier events (from McCabe et al., 1998). Areas of this study in western Ireland are outlined, corresponding to areas shown in Figures 2 and 6. CB—Clew Bay; KB—Killala Bay; DB—Donegal Bay.

FIELD MAPPING

Previous Work

Although the general distribution of drumlins and hummocky topography in western Ireland has been identified from over 100 yr of mapping, the interpretation of the glacial history of the region has differed. Charlesworth (1928) mapped striae, erratics, moraines, and marginal drainage features, from which he interpreted two sources of ice advancing into the region: one from the Joyce's Country to the south and one from Leitrim to the northeast (Fig. 1). He interpreted deglacial features as indicating a "staged ice retreat." Charlesworth argued that a moraine marking an ice margin that ran south from Ballycastle to Clew Bay (Figs. 1 and 2) recorded a particularly significant event during deglaciation.

Synge (1968) mapped a moraine in a similar position as Charlesworth's moraine, but he interpreted it as marking the western limit of the last (Midlandian) glaciation rather than a recessional feature. This moraine corresponds with the termination of the well-known drumlin field that extends into Clew Bay (Fig. 2).

McCabe et al. (1998) proposed that a moraine system identified across western Ireland was correlative with the Killard Point Stadial (Fig. 1). This moraine system extends northward into the lowlands to the east of the Nephin Beg Range, and then across an extensive area of hummocky topography between Crossmolina and Ballina (Fig. 2). As mapped, this moraine thus represents a less-extensive ice limit than that represented by Synge's (1968) moraine. McCabe et al. (1998) then traced this moraine system from the Ballina area across the eastern flanks of the Ox Mountains, beyond which it then loops back to the west, extending into Donegal Bay near Sligo (Fig. 1). The ice limit is then inferred to come back onshore on the northern side of Donegal Bay in association with a moraine east of Killybegs (Fig. 1).

McCabe et al. (1998) argued that regional northward ice-flow patterns that extend to Crossmolina are contemporaneous with those of the Killard Point Stadial identified across the north Irish Lowlands (Fig. 1). That the moraine system represented a readvance rather than a recessional phase of the ice sheet was based largely on geomorphic evidence of crosscutting relations of ice-flow indicators related to the younger moraine system. For example, McCabe et al. (1998) argued that such crosscutting relations in Clew Bay suggest that the moraine ridges on the northern side of the bay at ~100 m above sea level (asl) represent the Killard Point Stadial limit. Calibrated accelerator

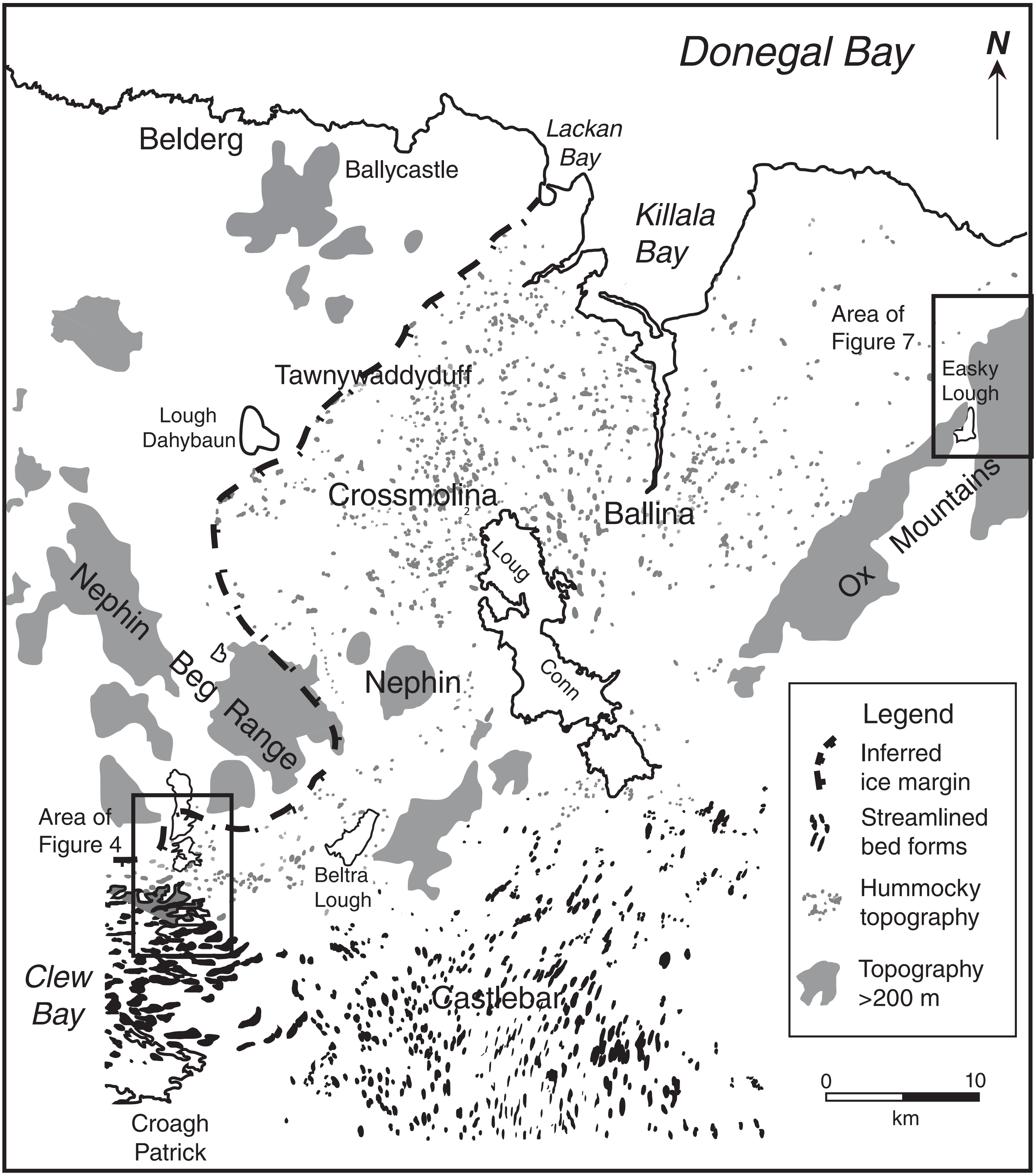

Figure 2. Glacial landforms in western Ireland between Clew Bay and Killala Bay associated with most recent ice limit shown by heavy dashed line corresponding to the Tawnywaddyduff moraine. See Figure 1 for location of area.

mass spectrometry (AMS) radiocarbon ages of 19.0–19.6 cal k.y. B.P. on in situ foraminifera from raised glaciomarine deposits at Belderg, on the south side of Donegal Bay (McCabe et al., 2005) (Fig. 1), are distal to this moraine system and are associated with retreat of the Last Glacial Maximum ice margin from the continental shelf. These dates are from sites to the west of Synge's (1968) moraine, indicating that Last Glacial Maximum ice was more extensive than inferred by Synge.

The Tawnywaddyduff Moraine

Our mapping strategy involved using historic maps, aerial photographs, satellite images, and field mapping to document the extent of the most recent ice advance into the region. In particular, mapping focused on evaluating the regional context of the moraine segments from which we sampled boulders for cosmogenic nuclide dating. Accordingly, we identified a regional moraine system that can be traced nearly continuously from the north side of Clew Bay to northeast of the Ox Mountains. We name this the Tawnywaddyduff moraine after the area near Tawnywaddyduff (Fig. 2) where the moraine is particularly well expressed. On the north side of Clew Bay and on the flanks of Nephin (Fig. 2), the Tawnywaddyduff moraine corresponds to the moraine mapped by Synge (1968) and McCabe et al. (1998). Otherwise, the location of the Tawnywaddyduff moraine is new, and we were unable to confirm the existence of previously mapped ice limits in the region.

On the north side of Clew Bay, where our first set of samples comes from, the moraine system is characterized by multiple linear, semiparallel boulder-studded till ridges that abut against the southern flank of the Nephin Beg Range (Figs. 3A and 4). The upper elevation of the moraine in this area is ~100 m asl. Immediately south (up-ice) of the moraine, at the head of Clew Bay, McCabe et al. (1998) noted that east-west linear till ridges with concave embayments that face northwest indicate an earlier westward ice flow followed by a younger event that flowed to the northwest, overprinting the existing drumlinized topography and corresponding to the ice advance that deposited the moraine in this area (Figs. 2 and 4).

To the east of Clew Bay, the moraine continues northeast along the flank of the Nephin Beg Range toward Beltra Lough, but there becomes characterized by hummocky topography underlain by ice-contact glaciofluvial sediments (Figs. 2, 3B, and 3C). At Beltra Lough, the moraine then turns north, following the eastern flank of the Nephin Beg Range and recording the advance of ice into the Lough Conn lowlands. Drumlins immediately south of the moraine indicate that associated northerly ice flow began to diverge near Castlebar, in response to the topographic influence of the Nephin Beg Range, into a northwesterly flow component directed into Clew Bay and a northeasterly flow component directed toward Lough Conn (Fig. 2).

The northward continuation of the Tawnywaddyduff moraine northwest of the Lough Conn lowlands marks the western margin of a broad ice lobe that extended north into Killala Bay. The western margin of the lobe was initially in contact with the Nephin Beg Range but then became confined to the lowlands, extending north to Lough Dahybaun and then northeast through Tawnywaddyduff and into Donegal Bay by way of Lackan Bay (Fig. 2). Here, our mapped moraine is ~10 km proximal to (east of) the Ballycastle-Mulraney moraine mapped in this region by Synge (1968). We were unable to confirm the existence of this part of Synge's moraine, although he shows it as a dashed line, indicating that it might simply be inferred.

The ice margin in these lowlands is delimited by the conspicuous boundary between the widespread hummocky topography to the east of the Tawnywaddyduff moraine and relatively flat topography to the west of the moraine (Fig. 2); marked linear morainal ridges with steep proximal slopes further demarcate this boundary (Fig. 3D). The elevation of the ice limit between the Nephin Beg Range to Lackan Bay, 30 km to the northeast, ranges from 50 to 80 m, suggesting relatively thin ice that was strongly controlled by topography. Most of the sediments in the region of hummocky topography to the east of the ice limit are ice-contact, stratified sand and gravel, but exposures in the moraine where it first extends beyond the Nephin Beg Range reveal massive, compact diamicton with abundant large angular clasts (Fig. 3E).

Geophysical surveys of the floor of Donegal Bay reveal a prominent set of moraines that extend from ~15 km offshore of Malin Beg south to where a projection of their continuation would place them as coming onshore in Lackan Bay (Fig. 5), or where our mapping places the northern end of the Tawnywaddyduff moraine (Fig. 2). Accordingly, our mapping indicates that the associated ice margin extended north into Donegal Bay and that much of the bay was filled with ice.

We have also identified segments of the Tawnywaddyduff moraine that, rather than demarcating the maximum spatial extent of the ice margin, instead record the upper limit of the ice surface where the topography projected above the ice surface as nunataks. These relations are critical to establishing the regional surface elevation of the former ice surface, from which we can reconstruct former ice-surface gradients and ice thicknesses. One such moraine segment is the kame terrace described by Synge (1968) that crosses the western flank of Nephin (Fig. 2) at an elevation of 300 m asl (Fig. 3F). Synge (1968) also identified a similar glacial limit at an elevation of ~290 m on the flank of Croagh Patrick, just south of Clew Bay (Fig. 2). These ice limits occur at the distal end of coeval ice-slow lines across the adjacent lowlands that emanate from discrete dispersal centers (Fig. 1).

We also mapped moraine segments in the coastal mountains between Ballina and Ballyshannon that further record the former ice-surface elevation at 300 m (Figs. 2 and 6). In the Ox Mountains, a right-lateral moraine descends down the north slope of the Easky Lough valley (Fig. 3G) from an elevation of ~300 m. A prominent moraine that backfills a cirque valley between Truskmore and Benbulbin Mountains occurs at an elevation of ~300 m (Figs. 3H and 6). Finally, a well-defined right-lateral moraine occurs at ~300 m on the northwestern flank of the massif immediately west of Lough Melvin (Fig. 6). As discussed further next, our dating results from these moraine segments in the Ox Mountains support our interpretation that the moraines record the upper limit of the last ice advance, rather than a recessional phase associated with thicker ice.

SAMPLING

We sampled eight boulders from each of two areas where we mapped moraine systems in the west of Ireland: the Ox Mountains 5–10 km south of Sligo Bay, and adjacent to Furnace Lough, directly north of Clew Bay (Figs. 1 and 2). The sampling strategy accounted for the following parameters: height of each boulder, latitude, longitude, and altitude recorded by global positioning system, and horizon geometry (i.e., the proportion of the sky above the horizon of the sample site shielded by the surrounding topography). The presence of lichen, surface weathering, and any evidence of striation(s) and glacial polish were noted.

We sampled the largest boulders available, with heights above the ground ranging from 0.7 m to 2.5 m (average of 1.35 m). Large boulders were chosen on stable surfaces to minimize possible postdepositional movements. Samples were taken from gneissic or granitic boulders in the Ox Mountains, and quartz-pebble conglomerate or arkose boulders in the Furnace Lough area. We sampled from the flat (dip <10°), uppermost portion of each boulder with a hammer and chisel.

The Furnace Lough area is located directly north of Clew Bay (Figs. 2 and 4). The study area

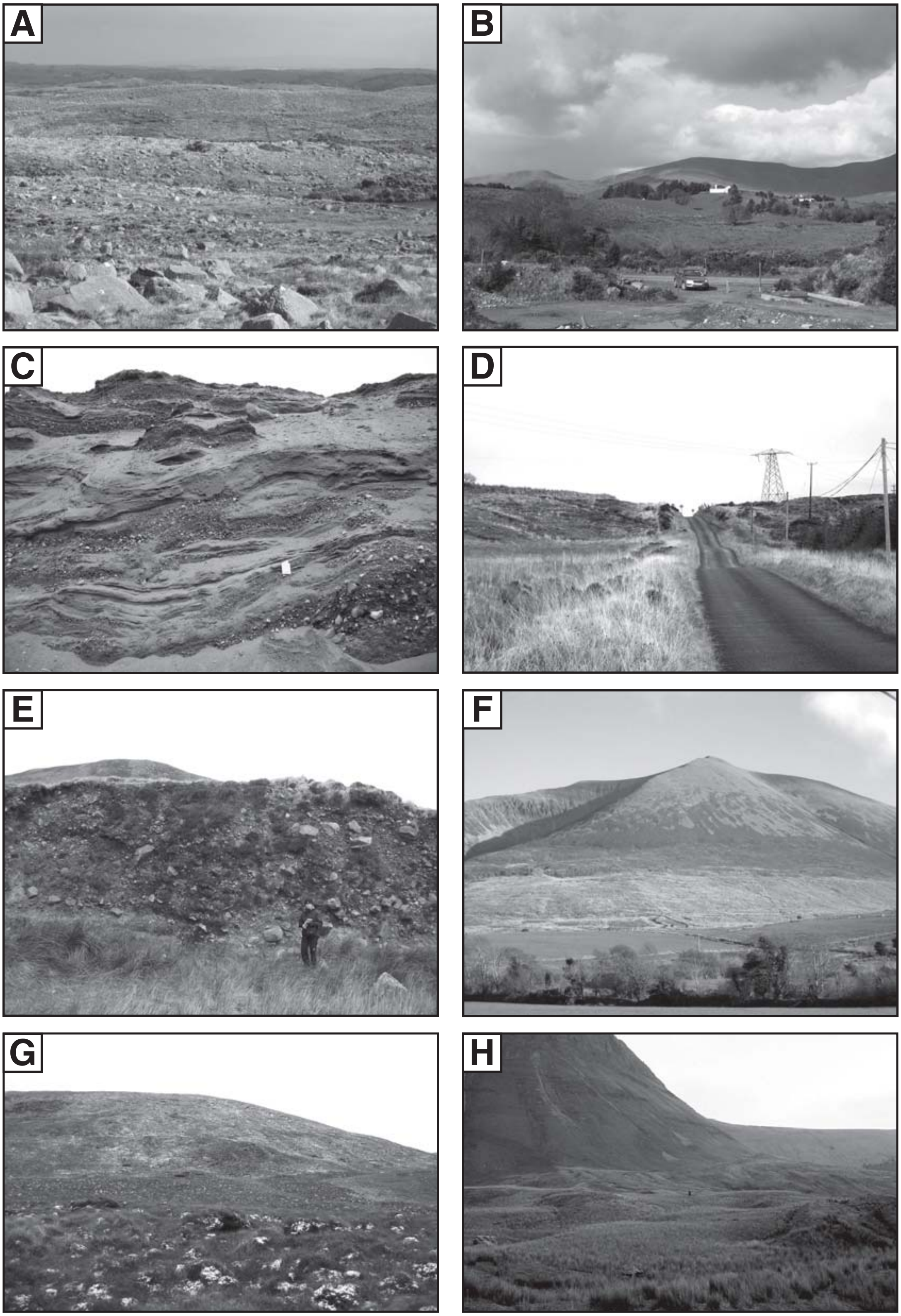

Figure 3. Photographs illustrating field relations of moraines and samples for cosmogenic dating. (A) Hummocky, bouldery moraine in the Furnace Lough area. (B) Hummocky moraine near Beltra Lough. (C) Internal sedimentary architecture of hummocks near Beltra Lough. Field notebook for scale (10 cm long). (D) Proximal slope of Tawnywaddyduff moraine in its type area. (E) Coarse, poorly sorted gravels comprising Tawnywaddyduff moraine where it abuts the eastern slopes of the Nephin Beg Range. (F) Well-defined lateral moraine crossing the western slopes of Nephin. (G) Right-lateral moraine descending slope from right to left, immediately to the east of Easky Lough. (H) Well-defined moraines backfilling valley between Benbulbin and Truskmore, south of Donegal Bay (see Fig. 6).

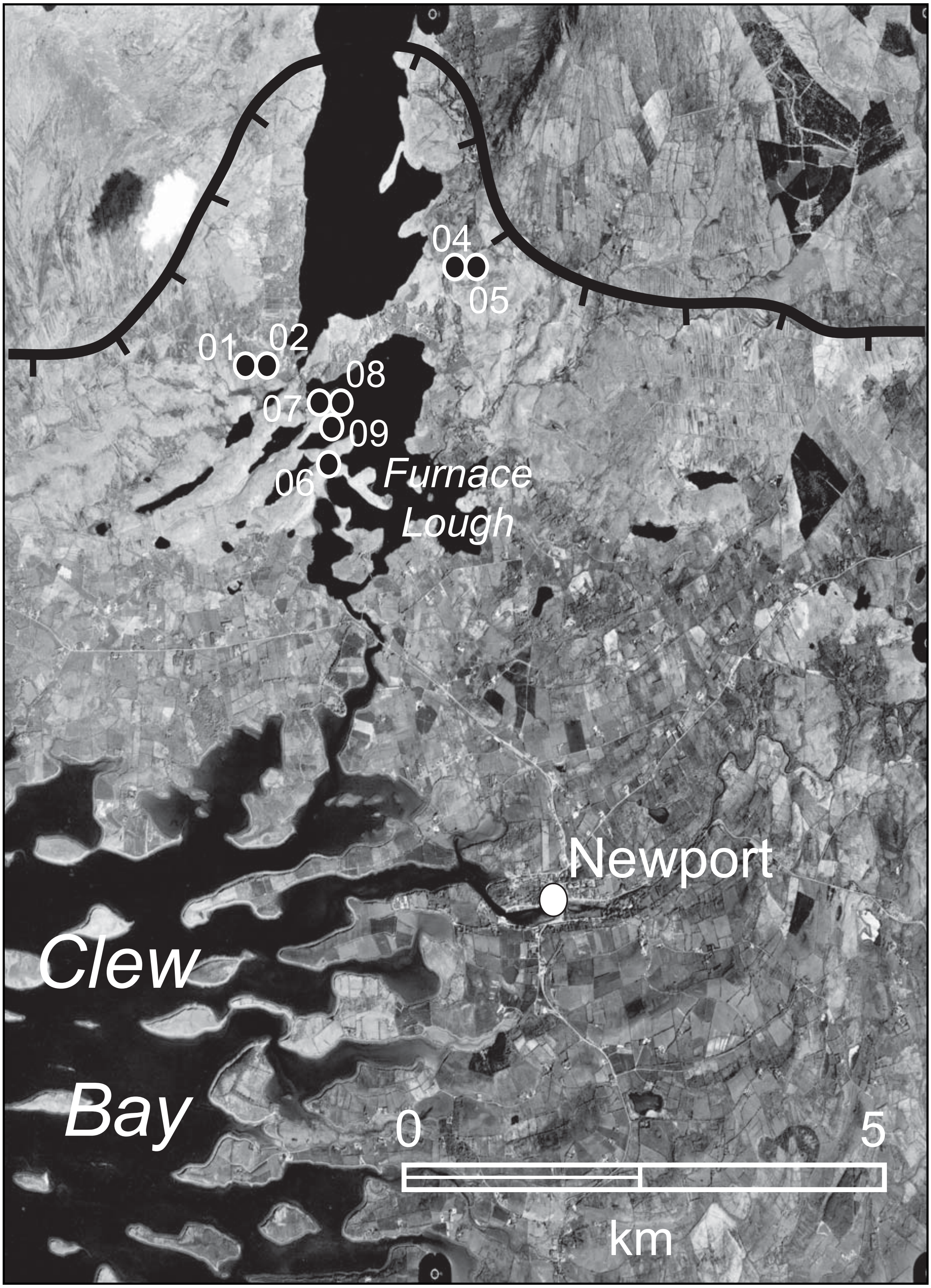

Figure 4. Location of samples for cosmogenic dating in the Furnace Lough area (see Table 1 for geographic coordinates).

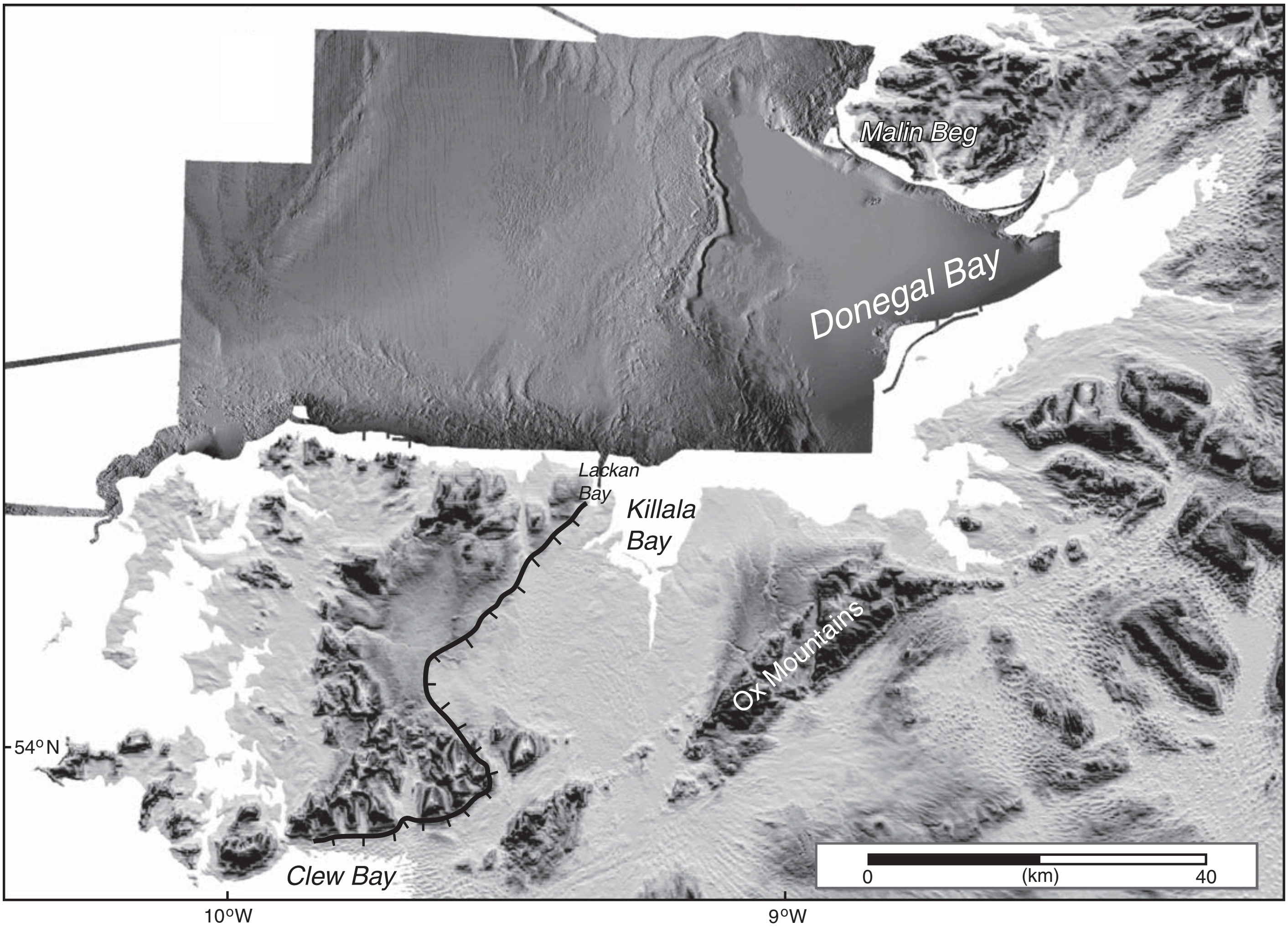

Figure 5. Digital elevation model of topography in western Ireland and bathymetry on the floor of Donegal Bay. Multibeam bathymetry data were acquired in 2002 onboard the R/V *Celtic Voyager* under the Irish National Seabed Survey (data collected in 2002 by Geological Survey of Ireland). The Tawnywaddyduff moraine is shown by line with single hachures.

is underlain by Dalradian metasediments, which consist of Carboniferous shales, sandstone, and quartz-pebble conglomerate (McCabe et al., 1986). All our samples were from the crests of the bouldery ridges that make up the Tawnywaddyduff moraine in this area (Fig. 4). Samples FL-04–04 and FL-04–05 were from arkosic boulders near the maximum limit of the moraine at an altitude of 75 m. Samples FL-04–01 (arkose) and FL-04–02 (quartz pebbles) were from a moraine segment at an altitude of 39 m, and samples FL-04–06, FL-04–07, FL-04–08, and FL-04–09 were from quartz-pebble conglomeratic boulders on a moraine segment at an altitude of 14–22 m.

The Ox Mountains are 5–10 km south of Sligo Bay. These mountains form a mostly bog-covered upland cut by valleys lakes including Easky Lough. Underlying rocks are Cambrian-Ordovician schists with thick quartz veins and Silurian quartz-rich granodiorites (MacDermott et al., 1996). Regional ice flow during the Last Glacial Maximum was toward the northwest, and the ice margin advanced seaward onto the continental shelf. Our samples are associated with the stratigraphically younger Tawnywaddyduff moraine system found in the main valleys of the Ox Mountains, indicating that uplands remained ice free as nunataks at the time of formation of this moraine. Large (1–3 m) striated granite erratics and gneissic erratics with quartz veins are common on segments of this moraine system, providing excellent material for ^{10}Be dating.

Samples came from three locations in the Ox Mountains (Fig. 7). Four samples (Ox-03–01, Ox-03–02, Ox-03–03, Ox-03–05) are associated with hummocky moraine in the Ladies Brae area. Three samples (Ox-03–06, Ox-03–07, Ox-03–09) are associated with a right-lateral moraine that descends from ~300 m to ~200 m on the north side of Easky Lough (Fig. 3G). One sample (Ox-03–10) was from the north-facing side of the Ox Mountains at 290 m, between Easky Lough and Ladies Brae.

METHODOLOGY

We extracted beryllium from quartz following a modified version of the procedure given in Kohl and Nishiizumi (1992) and Schnabel et al. (2007). The chemical preparation was performed in the Natural Environment Research Council (NERC) Cosmogenic Isotope Analysis Facility (CIAF) at the Scottish Universities Environmental Research Centre (SUERC). The grain-size fraction between 0.25 and 0.71 mm was selected. Grains containing minerals of

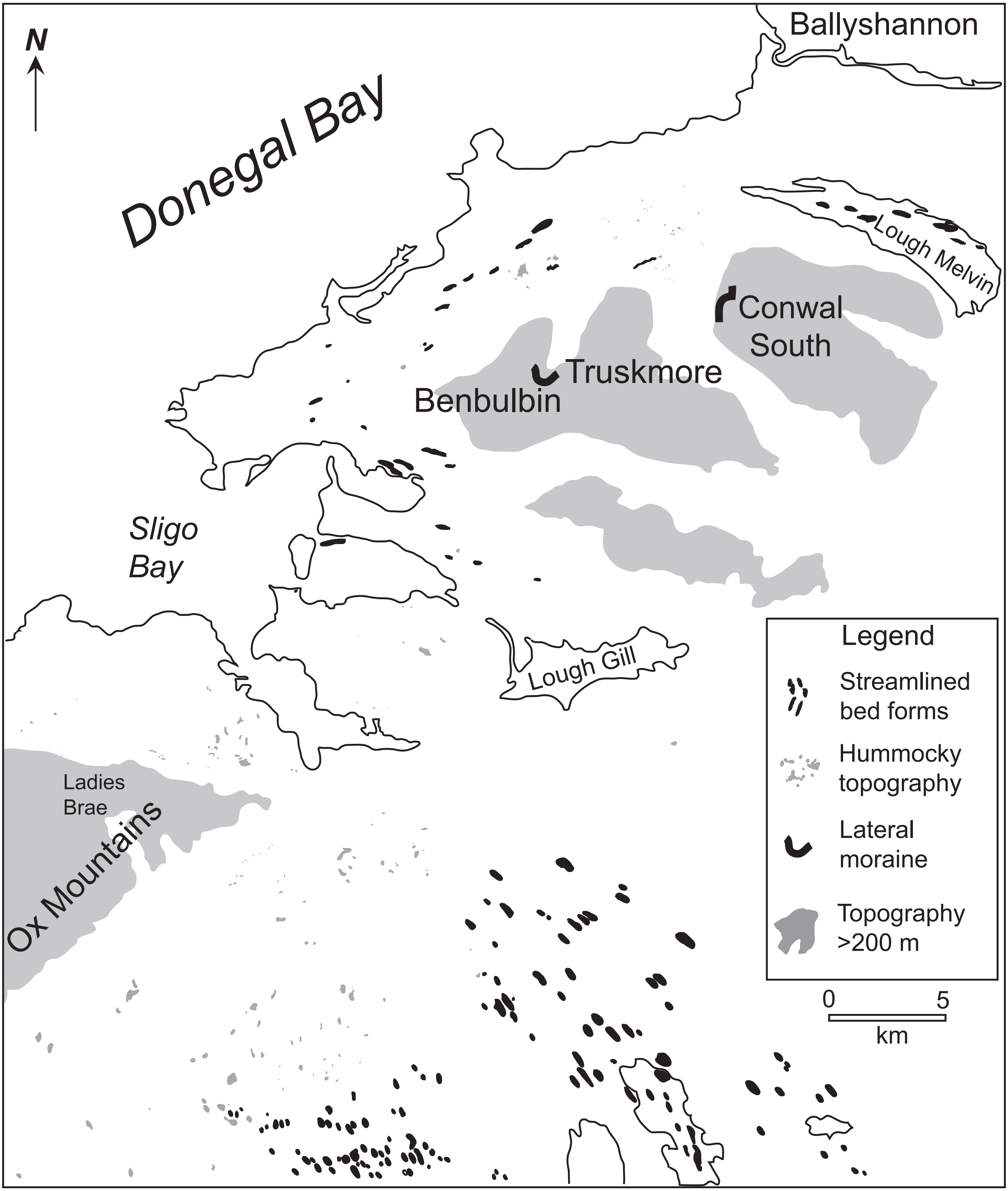

Figure 6. Glacial landforms in western Ireland between the Ox Mountains and Lough Melvin. Prominent lateral moraines marking upper limit of last ice advance occur in valley between Benbulbin and Truskmore (see Fig. 3H) and on western slope of Cornwal South. See Figure 1 for location of area.

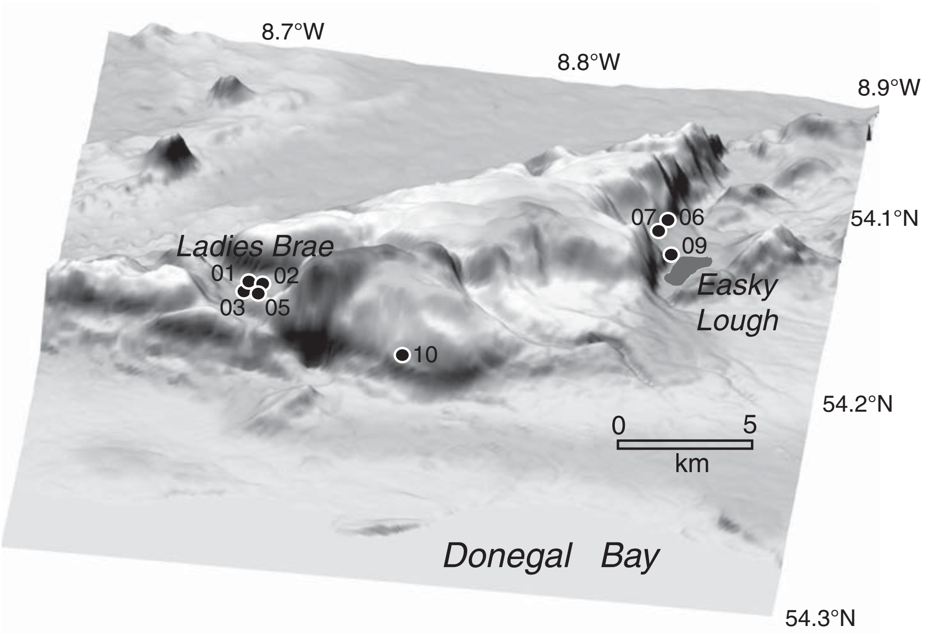

Figure 7. Location of samples for cosmogenic dating from the Ox Mountains. Perspective is oblique view looking south from Donegal Bay.

higher magnetic susceptibility were separated using a Frantz isodynamic magnetic mineral separator. The resulting quartz-rich fraction was leached repeatedly with solution of HF (2%). After three etching steps, minerals of specific gravity higher than quartz were removed using a mixture of LST solution (containing lithium heteropolytungstates) with water, so that quartz-rich grains floated on the heavy liquid. Etching was continued until satisfactory quartz purity was reached, as determined by atomic absorption spectrometry (AAS) measurement of the Al concentration. A 0.3 mg ^{9}Be spike was added to each sample of ~30 g purified quartz. Anion exchange and cation exchange chromatography, and selective precipitation techniques (including precipitation at pH 4 to remove Ti in the precipitate) were used to isolate and purify the beryllium, which was baked to BeO in a quartz crucible. A processing blank was prepared with every batch of eight samples using about the same ^{9}Be carrier mass as for the unknowns.

The ^{10}Be/^{9}Be ratios of the prepared BeO targets (mixed with Nb) were determined using the 5 MV Nippon Electric Company (NEC) accelerator mass spectrometer at SUERC (Freeman et al., 2004). The measurement is described in detail in Maden et al. (2007) and Schnabel et al. (2007). For normalization, we use 3.06×10^{-11} as the ^{10}Be/^{9}Be ratio for the National Institute of Standards and Technology (NIST) SRM4325 standard. This ratio agrees to within <0.5% with measurements from standard materials purchased from Kunihiko Nishiizumi (Nishiizumi, 2002). The nominal ratios used for primary and secondary standards disagree with the recalibration reported by Nishiizumi et al. (2007). However, the production rates used are consistent with the ratios used in this work. Consequently, only ^{10}Be concentrations reported here would be affected by implementing ^{10}Be/^{9}Be ratios from Nishiizumi et al. (2007), but not the exposure ages. Full chemistry blanks gave ^{10}Be/^{9}Be ratios of $(2–5) \times 10^{-15}$, whereas the samples were measured at $\sim 1 \times 10^{-13}$. All single exposure ages are reported with a 1σ uncertainty corresponding to the analytical uncertainties only (internal uncertainty). Analytical uncertainties associated with the AMS measurements vary from 5% to 7% in our samples. The standard deviations given include the standard deviation of the sample and of the standard measurement as well as the standard deviation associated with the blank correction and 3% for the chemical separation, which is dominated by the uncertainty of the Be concentration of the carrier solution used.

We calculated exposure ages using an assumed ^{10}Be production rate of 5.1 ± 0.3 atom g^{-1} yr^{-1} (before adjustments) at 1013.25 hPa air pressure at latitude >60° (Lal, 1991; Stone, 2000). This production rate was scaled for each site altitude and latitude using the correction factors based on air pressure (Stone, 2000). Where we compare our data to other chronologies, we also include the uncertainty in the ^{10}Be production rate (external uncertainty).

We corrected for surrounding topographic shielding and sample thickness according to Dunne et al. (1999). Cosmogenic ^{10}Be production decreases with depth, and the production rate must be scaled accordingly. We corrected the production rate for sample thickness using an exponential function (Lal, 1991), and using measured densities of the rock samples and an attenuation length of 155 g/cm^2. Given that all of our processed samples are from within 3 cm of the boulder surface, this amounts to a production rate correction of less than 3%.

Intermittent snow cover could reduce the production rate. However, because there is virtually no snow cover in western Ireland now, and because our samples come from the top of large boulders, we assume that any snow was blown off rapidly after storms and had a negligible effect on production. Thus, we do not correct for potential snow cover.

Samples are quartz veins from gneissic boulders in the Ox Mountains, and quartzites or quartz pebbles from conglomerates in the Furnace Lough area. Quartz veins and quartz pebbles projected up to 2 cm above the surrounding boulder surface, suggesting that some erosion of the surrounding boulder had occurred, but the preservation of polish and striations on the resistant quartz material suggests minimal or zero erosion of the material that was sampled. An erosion rate of 3 mm/k.y., for example, would increase our ages by 4%, but we did not correct our ages because of the evidence against erosion.

We report our final moraine ages as the mean of the sample population because the standard deviations of the sample-age populations are dominated by the geological uncertainties, as suggested by the standard deviation of the exposure ages being larger than the analytical uncertainty. We interpret the mean age and standard error of the mean exposure age as the final time of deposition of the moraine and thus as the time that the ice margin retreated from the moraine.

RESULTS

Our ^{10}Be ages from the Furnace Lough area range from 14.1 ± 0.7 to 17.3 ± 1.0 ka (Table 1; Fig. 8). A Shapiro-Wilk test indicates that we cannot reject the assumption that the data from this moraine have a normal distribution (W = 0.9675, p = 0.878). The mean age and standard error of the eight ^{10}Be samples from the Furnace Lough area is 15.6 ± 0.4 ka (Table 1).

Our ^{10}Be ages from the Ox Mountains range from 13.9 ± 0.9 to 18.1 ± 0.8 ka (Table 1; Fig. 9). A Shapiro-Wilk test indicates that we cannot reject the assumption that the data from this moraine have a normal distribution (W = 0.9037, p = 0.3117). Although we cannot statistically reject our two oldest samples (Ox-03–09 and Ox-03–10) from this population, we note

TABLE 1. ACCELERATOR MASS SPECTROMETRY MEASURED CONCENTRATIONS OF ^{10}Be AND ACCELERATOR AGES OF BOULDERS SAMPLED ON THE TAWNYWADDYDUFF MORAINE

Sample ID	Lab ID	Boulder height (m)	Sample thickness (cm)	Altitude (m)	Latitude (°N)	Longitude (°E)	[^{10}Be] (atoms/g)	Scaling factor	^{10}Be age (yr)	1σ (yr)[#]
OX-03-01*	b476SUERC	1.7	2.00	205	54.1874	8.7085	9.32E+04	1.22	15,000	870
OX-03-02*	b224SUERC	1.2	2.70	205	54.1874	8.7085	8.57E+04	1.21	13,910	930
OX-03-03*	b225SUERC	1.9	4.10	205	54.1874	8.7085	8.81E+04	1.19	14,530	810
OX-03-05*	b477SUERC	2.0	2.00	205	54.1874	8.7085	9.29E+04	1.22	14,960	790
OX-03-06*	b219SUERC	2.5	1.80	285	54.1388	8.836	9.93E+04	1.32	14,790	1100
OX-03-07*	b478SUERC	1.1	2.00	269	54.1404	8.836	1.10E+05	1.30	16,600	930
OX-03-09*	b270SUERC	2.5	1.90	199	54.1478	8.8448	1.12E+05	1.22	18,150	820
OX-03-10*	b472SUERC	2.7	2.00	290	54.2129	8.7633	1.14E+05	1.33	16,960	1060
FL-04-01[†]	b586SUERC	1.2	2.00	39	53.9209	9.5886	8.26E+04	1.04	15,690	1030
FL-04-02[†]	b587SUERC	0.9	2.00	39	53.9209	9.5886	8.01E+04	1.04	15,220	1010
FL-04-04[†]	b382SUERC	1.3	3.70	72	53.9259	9.558	7.57E+04	1.04	14,270	850
FL-04-05[†]	b383SUERC	1.0	5.20	74	53.9277	9.5595	9.31E+04	1.06	17,280	1020
FL-04-06[†]	b588SUERC	1.0	2.00	22	53.9151	9.5807	8.46E+04	1.02	16,370	1070
FL-04-07[§]	b274SUERC	1.2	2.86	14	53.9187	9.5791	8.20E+04	1.00	16,110	720
FL-04-08[§]	b473SUERC	0.7	2.00	14	53.9187	9.5791	7.90E+04	1.01	15,410	890
FL-04-09[§]	b271SUERC	0.8	3.57	14	53.9187	9.5791	7.16E+04	1.00	14,110	710

*Gneiss lithology.
[†]Arkose lithology.
[§]Quartz-pebble conglomerate lithology.
[#]Analytical uncertainty only.

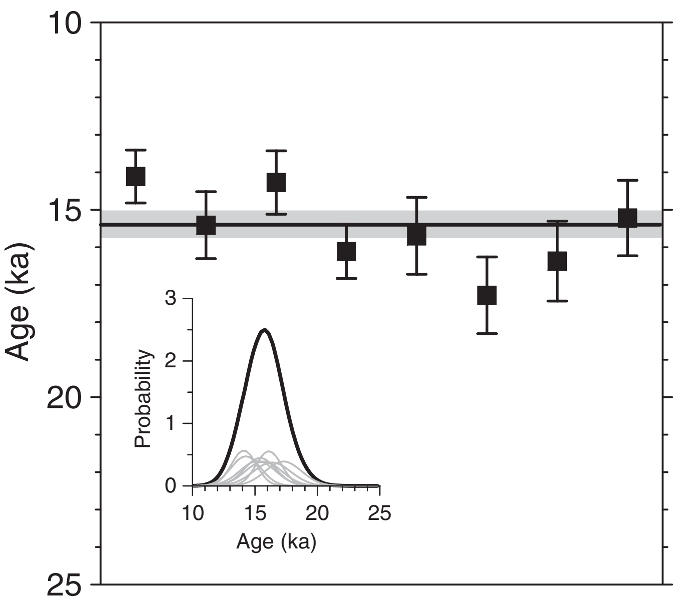

Figure 8. ^{10}Be exposure ages from Furnace Lough. Error bars correspond to 1σ analytical uncertainty only. The black horizontal line identifies the mean age of the data set. Shaded gray band corresponds to 1σ uncertainty (the standard error of the full data set). Inset graph shows the probability curves for the individual exposure ages (light-gray curves). Probability values are normalized so that each probability distribution is equal to 1. The sum of the probabilities is shown as the thick black curve.

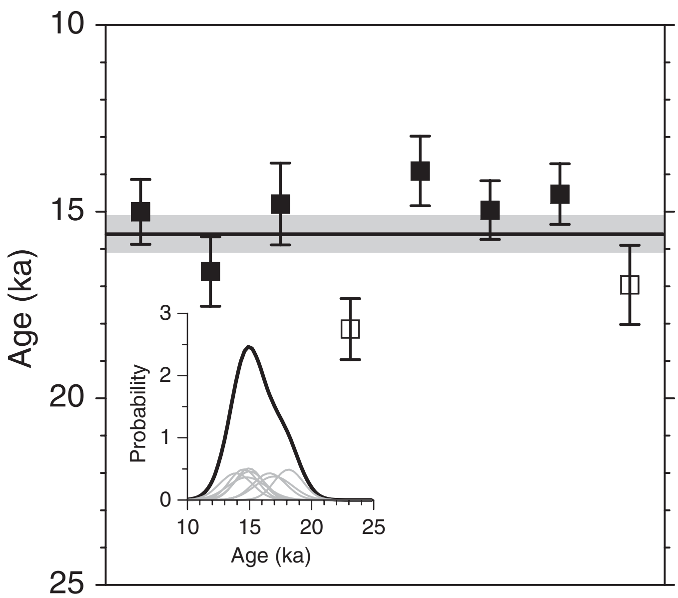

Figure 9. ^{10}Be exposure ages from the Ox Mountains. Error bars correspond to 1σ analytical uncertainty only. The black horizontal line identifies the mean age of the data set. Shaded gray band corresponds to 1σ uncertainty (the standard error of the full data set). Inset graph shows the probability curves for the individual exposure ages (light-gray curves). Probability values are normalized so that each probability distribution is equal to 1. The sum of the probabilities is shown as the thick black curve.

that their field relations may suggest that they may be associated with an older event. Sample Ox-03–09 was from the valley floor immediately (~200 m) south of the distal end of a right-lateral moraine that descends toward the west down the adjacent valley wall, and thus may have been deposited immediately beyond that ice margin by an earlier, more extensive event. Sample Ox-03–10 occurred on the northern flanks of the Ox Mountains at an elevation of 290 m. Although this is the same elevation as the upper limit of the Tawnywaddyduff moraine elsewhere where the corresponding ice surface intersected coastal mountains bordering Donegal Bay (Fig. 6), the location of the sample on the distal side of the Ox Mountains suggests that it may be associated with a higher ice surface than the ice margin that deposited the Tawnywaddyduff moraine. In both cases, however, these samples may just as readily be associated with the Tawnywaddyduff moraine, and their older ages relative to the remainder of the population may simply reflect some inheritance of ^{10}Be from previous exposure. Additional samples will be required to distinguish between these two hypotheses.

The mean age and standard error of the eight ^{10}Be samples from the Ox Mountains is 15.6 ± 0.5 ka (Table 1). On the other hand, if the two older samples discussed previously date from an earlier event, the mean age and standard error of the six younger ^{10}Be samples is 15.0 ± 0.4 ka, and that of the two older samples is 17.5 ± 0.6 ka.

A *t*-test indicates that the mean ages of the complete populations from the two areas are equal (two-tailed p value = 0.9318). Because there are no statistical differences, we combined the two populations to calculate a mean age and standard error of 15.6 ± 0.3 ka (n = 16) for onset of retreat of the western British-Irish Ice Sheet margin from the Tawnywaddyduff moraine in western Ireland. If the two older samples from the Ox Mountains do date from an earlier event, the mean age of the remaining 14 samples is 15.3 ± 0.3 ka, or statistically indistinguishable from the complete sample population.

DISCUSSION

Field mapping in the west of Ireland has identified a regional moraine system, herein named the Tawnywaddyduff moraine, which extends from the north side of Clew Bay northeast to Donegal Bay (Fig. 2). A prominent moraine system on the floor of Donegal Bay is correlative to the Tawnywaddyduff moraine, indicating that the actual maximum extent of the readvance was several tens of kilometers to the west of the coastal mountains, and that coastal lowlands bordering the bay were covered by ice (Fig. 5). Regional ice flow indicators suggest that ice flow into Donegal Bay during this event originated from two major ice-dispersal centers: one to the south in Joyce's Country and one to the east in the Omagh region (Fig. 1) (McCabe et al., 1998).

The age of this moraine, as constrained by our ^{10}Be ages from Furnace Lough (15.6 ± 1.0 ka, including production-rate uncertainty), is significantly younger than the age of glaciomarine sediments at Belderg distal to the Tawnywaddyduff moraine (ca. 19.5 cal k.y. B.P.), which constrains the time of deglaciation from the Last Glacial Maximum (McCabe et al., 2005). Combined with evidence for the crosscutting of ice-flow indicators associated with these two events (McCabe et al., 1998), these relations suggest that the Tawnywaddyduff moraine marks the limit of a post–Last Glacial Maximum readvance of the British-Irish Ice Sheet.

The agreement between our ^{10}Be ages from moraines in the Ox Mountains and Furnace Lough area support our field mapping, which shows that these segments are correlative and are part of the regional Tawnywaddyduff moraine system. This result is significant because it demonstrates that the moraine system identifies not only the maximum extent of the ice readvance (Tawnywaddyduff moraine) but also the upper limit of the ice surface in the mountains, where land areas at higher elevations remained ice free as nunataks. In particular, if the moraine segments in the Ox Mountains were recessional from thicker ice that covered more of the uplands, their ages should be younger than the Tawnywaddyduff moraine, marking the maximum spatial extent of this event. Accordingly, moraines in the coastal mountains provide a key constraint on the surface elevation of the former ice sheet. In particular, we find that the regionally consistent moraine elevation identifies a 300 m contour on the former ice surface where it encountered the mountains bordering Donegal Bay.

The moraine along the western edge of the Lough Conn lowlands demarcates a broad, low-gradient extension that advanced through the lowlands separating the Ox Mountains to the east and the Nephin Beg Range to the west. The geomorphological and sedimentological characteristics of the moraine in the Lough Conn lowlands indicate that the ice in the lowlands experienced widespread stagnation following advance to its maximum position. Hummocky topography also occurs immediately proximal to the Tawnywaddyduff moraine, where it marks the upper ice limit in mountainous terrain. In general, this hummocky topography is confined to a relatively narrow belt (1–10 km wide) that parallels the maximum limit of the ice margin (Fig. 2) but is otherwise largely absent up-ice toward the former center of the ice sheet (McCabe et al., 1998). This relatively narrow distribution of hummocky topography confined to the maximum ice-sheet limit may suggest an initial period of relatively slow retreat and corresponding local ice-margin stagnation, followed by rapid ice-margin retreat associated with the final deglaciation of the Irish Ice Sheet. Nevertheless, the absence of any substantial ice-contact deposits across central Ireland is surprising and perhaps reflects the fact that most debris was transported in a subglacial deforming bed and overlying ice was relatively debris free.

WIDER IMPLICATIONS

In the Irish Sea Basin, calibrated AMS ^{14}C ages from marine muds overridden by Killard Point ice indicate that the ice margin advanced sometime after 16.9 ± 0.2 cal k.y. B.P. (McCabe et al., 2005), whereas calibrated AMS ^{14}C ages from the type site of the Killard Point Stadial (Fig. 1) constrain the ice margin to have been at its maximum extent ca. 16.5 cal k.y. B.P. (McCabe and Clark, 1998) (Fig. 10). Calibrated AMS ^{14}C ages from Rough Island (Fig. 1) provide limiting ages on initial retreat of Killard Point Stadial ice from the Irish Sea Basin by ca. 15.5 cal k.y. B.P. (McCabe et al., 2007) (Fig. 10).

Insofar as our ^{10}Be ages date the time of retreat from the Tawnywaddyduff moraine, their mean age of 15.6 ± 0.3 ka (15.6 ± 1.0 ka with external uncertainty) is in excellent agreement with the limiting ages for ice retreat from Rough Island (Fig. 10), thus supporting the argument by McCabe et al. (1998), based on regional ice-flow patterns, that moraines in the west of Ireland are correlative to the Killard Point Stadial as dated in the Irish Sea Basin. Accordingly, these ages indicate the onset of near-contemporaneous retreat of the Irish Ice Sheet from its Killard Point Stadial limit at this time. We have thus constrained the Killard Point Stadial as being an ~1200-yr-long fluctuation of the Irish Ice Sheet margin. Calibrated basal ^{14}C ages from peat deposits at Sluggan Bog, north of Belfast (Fig. 1), provide limiting ages for complete deglaciation of the Irish Ice Sheet at or before ca. 14.4 ± 0.2 cal k.y. B.P. (Lowe et al., 2004) (Fig. 10).

Given the possibility that the two older ages from the Ox Mountains may record an older event, we note that they may be associated with the Clogher Head Stadial identified in the Irish Sea Basin (McCabe et al., 2007). The mean age of the remaining samples from the Ox Mountains and the samples from Furnace Lough is not significantly different if we include these two older samples (Fig. 10), and so our conclusions

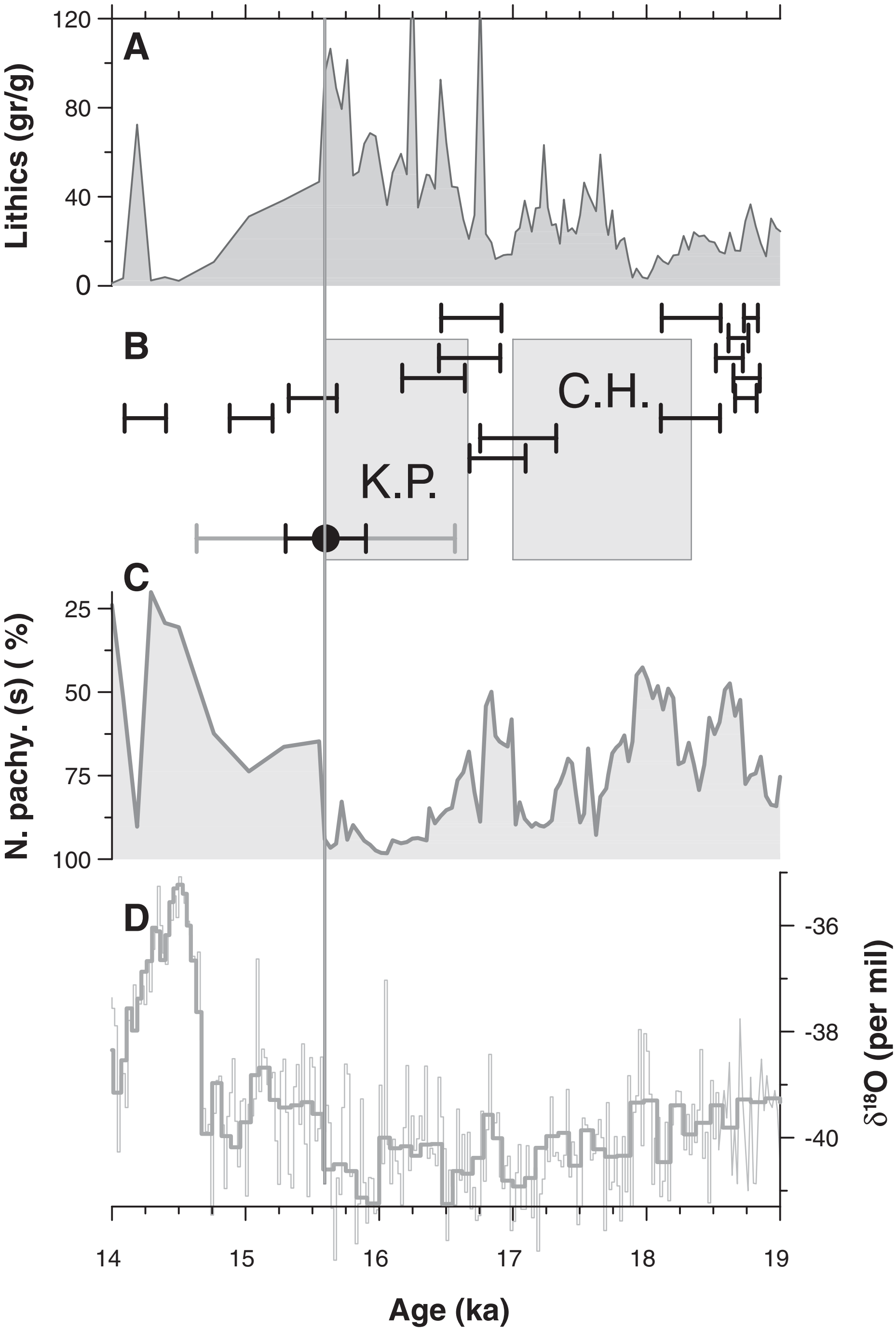

Figure 10. (A) Record of ice-rafted debris from marine core DAPC2 from the northeast Atlantic, ~200 km west of northwest Scotland (Knutz et al., 2007). (B) Our mean ^{10}Be age from combined Furnace Lough and Ox Mountain data set (black solid dot; 1σ internal error shown by black line; 1σ error, including production-rate error, shown by gray line) and calibrated radiocarbon dates (Stuiver et al., 2005) that constrain ages of the Killard Point (K.P.) and Clogher Head (C.H.) stadials during last deglaciation (duration indicated by gray boxes) (Lowe et al., 2004; McCabe and Clark, 1998, 2003; McCabe et al., 2005, 2007). Specifically, these ages identify interstadials represented by raised marine deposits that underlie and overlie glacial sediments associated with the two stadials. (C) Percent *Neoglobquadrina pachyderma* from marine core DAPC2. (D) The δ^{18}O record from the Greenland Ice Sheet Project 2 (GISP2) ice core (Grootes et al., 1993; Stuiver and Grootes, 2000). N. pachy—*Neoglobquadrina pachyderma*.

regarding the age of deglaciation from the Tawnywaddyduff moraine are not affected.

In Figure 11, we show the approximate extent of the Irish Ice Sheet in central and western Ireland during the Killard Point Stadial as reconstructed by McCabe et al. (1998) but modified in the west of Ireland as suggested by our new mapping. Regional ice-flow indicators associated with the Killard Point Stadial suggest that ice-dispersal centers existed over the north-central Irish Lowlands and in Joyce's Country (Synge, 1969; McCabe et al., 1998).

To further evaluate the geometry of the Irish Ice Sheet during the Killard Point Stadial, we reconstructed two east-west ice-surface profiles, one in Donegal Bay and the other to the south of the bay (Fig. 11). Both profiles are based on the 300 m surface elevation from the moraines in the coastal mountains and the position of the terminal Tawnywaddyduff moraine to the west. Neither profile is corrected for isostatic depression nor do they account for the underlying topography. Basal ice shear stress (τ) was calculated from $\tau = \rho_i gh \sin \alpha$, where ρ_i = 910 kg m^{-3}, g = gravitational constant (9.8 m s^{-2}), h = ice thickness, and α = ice-surface slope.

The first profile (A–A′, Fig. 11) identifies a low ice-surface slope, with a corresponding basal shear stress of 14.9 kPa. Such a low shear stress must reflect significant basal motion through some combination of sliding and bed deformation associated with marine sediments in Donegal Bay. The second profile (B–B′, Fig. 11) identifies a steeper ice-surface slope and attendant basal shear stress (22.3 kPa) than that in Donegal Bay. Nevertheless, these values remain well below those associated with ice moving entirely by internal ice deformation, suggesting substantial basal motion by sliding or sediment deformation.

We next extrapolated the profiles up-ice (east) of the coastal mountains in the direction suggested by regional ice-flow indicators. Because the up-ice extrapolation of the Donegal profile is no longer in Donegal Bay, we assign this segment the higher shear stress that we derived from the profile south of Donegal Bay. We extended both profiles up-ice until they reached an altitude of 500 m. Given that the direction of an ice-surface slope is perpendicular to the direction of ice flow, we drew the 500 m ice surface contour between the two profiles, and then extended this contour to the east, so that it remained perpendicular to regional ice flow (Fig. 11). Approximately 30 km west of Armagh, the 500 m contour lies directly up-ice of a regional flow line that originates at the ice margin near Ardee (Fig. 11). The corresponding ice profile along this flow line (C–C′, Fig. 11) has a basal shear stress of 31.4 kPa, or somewhat higher than the

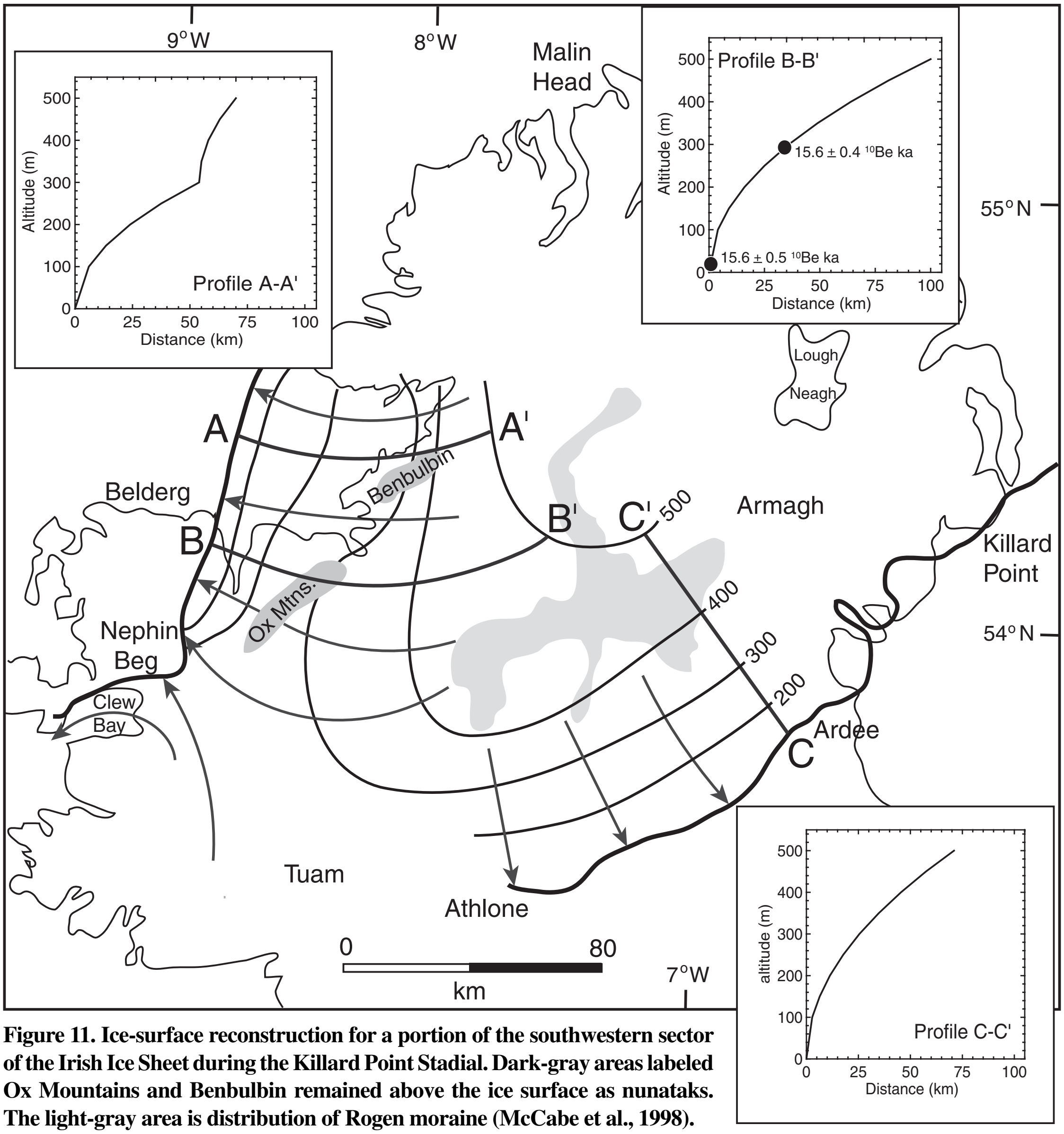

Figure 11. Ice-surface reconstruction for a portion of the southwestern sector of the Irish Ice Sheet during the Killard Point Stadial. Dark-gray areas labeled Ox Mountains and Benbulbin remained above the ice surface as nunataks. The light-gray area is distribution of Rogen moraine (McCabe et al., 1998).

profile south of Donegal Bay but still consistent with substantial basal motion by sliding or sediment deformation. Elevations along this profile were then used with those from the other two profiles to contour the surface of this sector of the Irish Ice Sheet, where contours remained perpendicular to ice flow (Fig. 11).

This reconstruction suggests that the Irish Ice Sheet during the Killard Point Stadial reached a maximum surface elevation of ~500 m over the northern Irish Lowlands (Fig. 11). Smaller satellite ice domes may have existed to the northwest over Donegal and to the southwest over Joyce's Country (Fig. 1), but because of their smaller size, they would also have been lower than the central ice dome over the northern Irish Lowlands.

McCabe et al. (1998) mapped the distribution of Rogen moraine across the central Irish Lowlands (Fig. 11), which they interpreted was formed during a pre–Killard Point event and then preserved beneath slow-moving ice during subsequent readvance during the Killard Point Stadial. We note that the Rogen moraine largely occurs beneath a broad, low-sloping shoulder of the reconstructed central Irish Lowlands ice dome associated with strongly divergent flow. The combination of position near the ice divide and divergent flow would result in extremely low basal ice velocities, which would promote conditions favorable for preservation of preexisting landforms, as inferred by McCabe et al. (1998).

The start of retreat of the Killard Point ice margin at 15.6 ± 1.0 k.y. B.P. in western Ireland and ≥15.5 cal k.y. B.P. in the Irish Sea Basin clearly indicates that deglaciation occurred during the Oldest Dryas stadial and ~1000 yr prior to the onset of the Bølling interstadial at 14.6 cal k.y. B.P. (Fig. 10). However, this deglaciation phase of the Irish Ice Sheet may have occurred in response to an earlier abrupt warming that is recorded in the Greenland Ice Sheet Project 2 (GISP2) $\delta^{18}O$ ice-core record and in a marine record from a core that was recovered 200 km east of the maximum extent of the British-Irish Ice Sheet margin off western Scotland (Knutz et al., 2007) (Fig. 10). This earlier abrupt warming is about one-quarter of the amplitude of the Bølling warming, which represented ~10 °C of atmospheric warming over Greenland (Severinghaus and Brook, 1999). If scaled accordingly, there may thus have been 2–3 °C of atmospheric warming at 15.5 ka. Assuming a lapse rate of 6 °C/1000 m and no change in precipitation, this warming would induce a 330–500 m rise of the equilibrium line altitude (ELA) of the Irish Ice Sheet.

The accumulation-area ratio (AAR) on modern valley glaciers, which represents the ratio of the accumulation ratio to the total area of the glacier, is well established at 0.65 ± 0.05, whereas on ice caps, it tends to be higher at 0.8 ± 0.05 (Andrews, 1975). For the reconstructed Killard Point ice surface, extending from the ice divide northwest to the ice margin in Donegal Bay, an AAR of 0.80 ± 0.05 would put the ELA 16–26 km up-ice from the ice margin, or at 200–250 m. Accordingly, in the absence of any change in precipitation, a 2–3 °C warming would induce the ELA to rise to an altitude higher than the reconstructed elevation of the ice divide, causing the entire ice sheet to be in the ablation area.

This result for sensitivity of the Irish Ice Sheet to a warming and attendant rise of the ELA is dependant on the reconstructed ice-surface gradient: a steeper (gentler) profile would make the ice sheet less (more) sensitive to minimal warming (Oerlemans, 2005). Our reconstructed profile seaward of the Ox Mountains to the ice margin in Donegal Bay is reasonably well constrained from moraines, whereas our estimate for the ice-divide elevation is based on the assumption that this profile can be extrapolated up ice at the same shear stress. An alternative constraint would involve the assumption that the shear stress for the flow line extending up-ice from the southern margin to the ice divide is 100 kPa, or a value that would be typical of a cold-based ice sheet (Paterson, 1994). Such a shear stress would result in the ice divide being ~100 m higher than we have reconstructed, with an attendant increase in the shear stress along the northern flow line to 35 kPa. In general, however, this does not significantly change the ice-surface profile and, thus, the sensitivity of the ice sheet to warming. We thus conclude that the relatively small size of the Killard Point ice sheet, combined with its low surface slope, resulted in the sensitivity of the ice sheet to relatively small changes in climate.

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