Exploring controls of the early and stepped deglaciation on the western margin of the British Irish Ice Sheet

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ABSTRACT: New optically stimulated luminescence dating and Bayesian models integrating all legacy and BRITICE-CHRONO geochronology facilitated exploration of the controls on the deglaciation of two former sectors of the British-Irish Ice Sheet, the Donegal Bay (DBIS) and Malin Sea ice-streams (MSIS). Shelf-edge glaciation occurred ~27 ka, before the global Last Glacial Maximum, and shelf-wide retreat began 26–26.5 ka at a rate of ~18.7–20.7 m a⁻¹. MSIS grounding zone wedges and DBIS recessional moraines show episodic retreat punctuated by prolonged stillstands. By ~23–22 ka the outer shelf (~25 000 km²) was free of grounded ice. After this time, MSIS retreat was faster (~20 m a⁻¹ vs. ~2–6 m a⁻¹ of DBIS). Separation of Irish and Scottish ice sources occurred ~20–19.5 ka, leaving an autonomous Donegal ice dome. Inner Malin shelf deglaciation followed the submarine troughs reaching the Hebridean coast ~19 ka. DBIS retreat formed the extensive complex of moraines in outer Donegal Bay at 20.5–19 ka. DBIS retreated on land by ~17–16 ka. Isolated ice caps in Scotland and Ireland persisted until ~14.5 ka. Early retreat of this marine-terminating margin is best explained by local ice loading increasing water depths and promoting calving ice losses rather than by changes in global temperatures. Topographical controls governed the differences between the ice-stream retreat from mid-shelf to the coast. Copyright © 2021 The Authors Journal of Quaternary Science Published by John Wiley & Sons Ltd.

KEYWORDS: deglaciation; Donegal; ice streams; Malin Sea; retreat rate

Introduction

The assessment of the rate and style of ice sheet retreat closely relates to many globally important scientific and socio-economic questions (IPCC, 2013). Constraining the pace of ice-sheet retreat for both past and present ice sheets can improve our understanding of how large ice masses respond to local and global, internal and external forcing, such as glaciological, climatic and oceanographic changes. Once insights gained from such knowledge are incorporated into ice sheet models, they can improve the predictions on how modern ice sheets will evolve with the current changing climate, ocean temperature and sea level (Rignot et al., 2010; Joughin et al., 2014). The behaviour of ice streams is of interest as they are a major regulator of the mass balance of ice sheets (Stokes and Clark, 1999, 2001; Payne et al., 2004; Roberts et al., 2010; Stokes, 2018). At ice stream termini, the reduction or loss of buttressing ice shelves can lead to thinning of upstream-based ice and the acceleration of ice flow and this behaviour has been recorded in modern ice streams in Greenland and West Antarctica (Krabill et al., 2000, 2004; Rignot et al., 2004a; Pritchard et al., 2009; Sonntag et al., 2012). Additionally, ice streams react more readily than other parts of the ice margin to any perturbation in ocean circulation, atmospheric temperature and sea-ice distribution because of both thermal (melting) and mechanical (floatation and calving) stressors that occur along the margins of marine-terminating ice sheets (e.g. Hulbe et al., 2004; Payne et al., 2004; Scambos et al., 2004; Shepherd et al., 2004; Joughin et al., 2012).

While modern ice streams are being extensively studied, the temporal resolution of such studies is limited. Numerical–glaciological, isostatic and palaeoclimatic models all require
empirical constraints on past ice-sheet extent and dynamics either to direct their formulation or for the testing of model outputs (Hughes et al., 2016). Such information over centuries and millennia can only come from palaeo-analogues, where a complete record of deglaciation may be better visible and quantifiable (Svendsen et al., 2004; Bradwell et al., 2008; Chiverrell et al., 2013; Hughes et al., 2016). The last British Irish Ice Sheet (BIIS) has been proposed as a potential analogue for sensitive areas of modern ice sheets (Clark et al., 2012). The BIIS, at several times in the past, had an abundance of marine-terminating ice, which would have been sensitive to both climatic and oceanic forcing, and was drained by radiating ice streams, which were probably critical to BIIS dynamics and overall mass balance during retreat (e.g. Boulton, 1990; Rignot et al., 2004b; Boulton and Hagdorn, 2006; Sole et al., 2008; Hubbard et al., 2009; Pritchard et al., 2009).

The Malin Sea includes the continental shelf to the west of Scotland, often referred to as the Malin Shelf, and the portion of the continental shelf northwest of Ireland that includes Donegal Bay (Fig. 1). The Malin Sea received flows from two large convergent ice masses derived from the Hebridean Islands, mainland Scotland, the North Channel and the north of Ireland. Ice radiating from the mountains of Donegal in northwest Ireland formed an independent centre of ice dispersal that not only fed ice towards the north, but also west and southwest into Donegal Bay (Fig. 1 inset). Ice in Donegal Bay was also fed from the lowland ice domes through Counties Mayo and Sligo (Greenwood and Clark, 2009a, 2009b). The former ice masses occupying the Malin Shelf and Donegal Bay meet the fundamental geomorphological criteria for ice streams (Stokes and Clark, 1999, 2001; Stokes, 2018) in the form of convergent flows and ubiquitous elongated bedforms (Dunlop et al., 2010; Stokes and Clark, 2001; Stokes, 2018) in the form of convergent flows and ubiquitous elongated bedforms (Dunlop et al., 2010; Stokes and Clark, 2001; Stokes, 2018).
Finlayson et al., 2014; Dove et al., 2015). The ice on the Malin Shelf has attracted a variety of names, including Barra Fan Ice Stream (Scourse et al., 2009; Dunlop et al., 2010; Callard et al., 2018), Hebrides Ice Stream (Dove et al., 2015; Small et al., 2017a) and Malin Sea Ice Stream (Wilson et al., 2019). In addition, sectors of the ice mass have also been referred to separately as other names including the North Channel Ice Stream (Finlayson et al., 2010, 2014; Hughes et al., 2014).

Here, this marine-terminating ice stream is termed the Malin Sea Ice Stream (MSIS); it drained between 5 and 10% of the BIS and fed the southernmost glaciogenic fan on the European continental margin, as well as largest sedimentary depocentre of the BIS, the Donegal-Barra Fan (DBF – Fig. 1) (Knutz et al., 2001; Howe et al., 2012; Dove et al., 2015). Deep-water cores suggest that this portion of the ice sheet responded quickly to millennial-scale climate oscillations suggesting a strong link between climate cycles and glaciological processes (Knutz et al., 2001; Scourse et al., 2009; Hibbert et al., 2010). Its sensitivity to climatic and oceanographic changes is also captured by numerical advances in offshore geomorphological mapping, through the mapping of landforms in representations of former ice limits (et al., 2010; Sacchetti et al., 2019). Reconstructions of the BIS have relied heavily on onshore geochronology, including radiocarbon, optically stimulated cosmogenic nuclide (TCN) ages (Table 4) (Schiele, 2017; Small et al., 2017a, 2019). Their stratigraphic and landform contexts, have been reported in a series of publications (Hubbard et al., 2009; Patton et al., 2013a, 2013b, 2016, 2017). The other marine-terminating ice stream in the southern portion of the Malin Sea was fed by ice flowing through Donegal Bay, and has surprisingly never been named and is referred to here as the Donegal Bay Ice Stream (DBIS). Less is known about the glaciological integrity and incorporate the age information within Bayesian chronosequence models for the differing ice streams or glaciers (e.g. Chiverrell et al., 2013). All published and new deglaciation ages are presented here for the DBIS and MSIS, with all ages subject to a triage system assessed according to quality criteria (Small et al., 2017b), and here only ages deemed of good quality (green and amber) are included. A few pre-Last Glacial Maximum (LGM) ages (flagged as problematic in a quality assessment of the value of legacy ages for constraining deglaciation) (Small et al., 2017b) were nonetheless used as indicators of previous ice-free conditions (Fig. 1; see Table 6) (Colhoun et al., 1972; Jardine et al., 1988; Bos et al., 2004). The new deglaciation ages include offshore radiocarbon (14C) ages (Table 3) (Callard et al., 2018; O Cofaigh et al., 2019) and onshore terrestrial cosmogenic nuclide (TCN) ages (Table 4) (Schiele, 2017; Small et al., 2017a; Wilson et al., 2019) (See Fig. 1 for all locations). In addition, 15 new optically stimulated luminescence (OSL) ages have been obtained and are presented here for the first time (Fig. 1; Table 1). The OSL sites were selected targeting spatial gaps in the retreat sequences and to reassess sites yielding conflicting ages in the existing deglacial chronology for the region.

Details on the methods used to process all TCN and 14C samples are reported in the relevant publications (see Tables 3 and 4 for references). The original 14C measurements have been calibrated using OxCal 4.2, with for marine-derived samples the Marine-13 calibration curve and applying a marine reservoir correction of 0 years (Bronk Ramsey, 2009a; Reimer et al., 2013). They are reported to two decimal places as cal ka BP. 14C ages were calibrated aresh using a consistent marine C reservoir during the Bayesian modelling. Only 14C ages representing latest glacial and deglaciation ages are included in this paper (i.e. not younger ones). Cosmogenic ages include 10Be and 36Cl exposure ages and, in the text, all TCN ages are rounded to the nearest 0.1 ka and shown with the ±1 sigma external uncertainty, unless otherwise stated. To be consistent across all BRITICE-CHRONO publications, 10Be ages presented here have been calculated using the calculator formerly known as the CRONUS-Earth calculator (Developmental version; Wrapper script 2.3, Main calculator 2.1, constants 2.2.1, muons 1.1; Balco et al., 2008). 10Be ages are calibrated using the Loch Lomond local production rate (LLPR; Fabel et al., 2012; Small and Fabel, 2015) which is linked to direct independent age control provided by limiting radiocarbon ages (MacLeod et al., 2011). All other Scottish calibration sites rely on an assumed Younger Dryas deglaciation age (Borchers et al., 2016; Martrero et al., 2016), assumed tephra age within a varve chronology (Small and Fabel, 2015), and a contested radiocarbon chronology (Lowe et al., 2019;
Radioactivity and dose rate data for luminescence samples.

Beta dose rate $- 1)$

| Site              | Sample   | Depth (m) | Water (%) | U (p.p.m.)* | Th (p.p.m.)* | K (%)* | Rb (p.p.m.)* |
|-------------------|----------|-----------|-----------|-------------|-------------|--------|--------------|
|                   |          |           |           |             |             |        |              |
| T6BROC01          | 2021     | 7         | 1.51±0.08 | 3.0±0.3    | 1.1±0.12   | 3.2±0.12| 0.07±0.04    |
| T6BROC02          | 5        | 5         | 1.03±0.07 | 3.4±0.19   | 1.3±0.12   | 3.0±0.04| 0.15±0.01    |
| T6GULR01          | 2.3      | 27        | 1.16±0.11 | 7.2±0.25   | 2.1±0.10   | 4.9±0.47| 0.38±0.02    |
| T6LNAC01          | 5        | 27        | 1.66±0.07 | 4.6±0.46   | 2.1±0.06   | 4.2±0.23| 0.44±0.01    |
| T7FAWN01          | 5        | 27        | 0.92±0.08 | 3.0±0.3    | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7FAWN02          | 1.1      | 18        | 0.82±0.06 | 3.0±0.3    | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7FAWN03          | 1.0      | 18        | 0.82±0.06 | 3.0±0.3    | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7FAWN04          | 1.1      | 18        | 0.82±0.06 | 3.0±0.3    | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7GLEN01          | 12       | 27        | 1.16±0.11 | 3.0±0.3    | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7GLEN02          | 2        | 12        | 0.54±0.02 | 2.0±0.10   | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7GLEN03          | 1.1      | 18        | 0.54±0.02 | 2.0±0.10   | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |
| T7GLEN04          | 1.0      | 18        | 0.54±0.02 | 2.0±0.10   | 2.0±0.10   | 3.5±0.26| 0.26±0.01    |

*The analytical chemistry laboratory did not provide uncertainties on individual U, Th, K or Rb concentrations. Based on replicate analyses, uncertainties of 10% were assumed for U, Th and Rb, and 5% for K, and these uncertainties were propagated through the dose rate calculations.

OSL dating

Samples for OSL dating were collected from eight sites across the two transects targeting glacialfluvial and deltacal outwash sands and gravels. All sites were selected based on their ice-proximal context and the potential to constrain the timing for well-defined ice margins (Fig. 1; Table 2). OSL dating is underpinned by the principle that exposure to sunlight zeros or bleaches an OSL signal that develops within mineral grains (typically quartz or K-feldspar). The OSL signal increases with the duration of burial in sediments as the materials are exposed to natural radiation, increasing the charge stored within quartz or feldspar. Here we use small aliquots (SA; ~20 grains) of sand-sized quartz grains separated from sediments to measure the OSL signal (Murray and Wintle, 2000; Duller, 2008). For samples that have been bleached heterogeneously, the measurement of multiple replicates (typically here ~50) can identify those grains exposed to sunlight most recently, which are referred to as a well-bleached population. With heterogeneous bleaching, statistical models are required to determine an accurate age, e.g., the Minimum Age Model (MAM) (Galbraith et al., 1999) or the internal–external uncertainty (IEU) model (Thomsen et al., 2007).

At all sites opaque tubes were hammered into sedimentary sections to prevent exposure to sunlight during sampling. The external gamma dose rates were determined using in situ gamma spectrometry, with external beta dose rates calculated from U, Th, K and Rb concentrations determined by inductively coupled plasma mass spectrometry (ICP-MS). The sample preparation and analysis methods used were identical to existing studies (Evans et al., 2017; Bateman et al., 2016). Appropriate conversion factors (Guérin et al., 2011, 2012) were applied to calculate the final total dose rate (Table 1) including grain size. The sites sampled all have water tables that are presently artificially low, owing to either coastal erosion or aggregate extraction. Maximum pore spaces in 180–250-µm sand are in the range between rhombohedral (26%) and random (40%) packing, which for moderately sorted rounded to sub-rounded sands equates to saturated water contents of around 30%. In terms of palaeomorristic attenuation, contents of 23±5% were used for shallow and drier samples and 27±5% for deeper saturated samples.

OSL analyses were performed on the 180–250-µm size fraction (Table 2) and using aliquots each containing ~20
grains. The low proportion of quartz grains emitting an OSL signal for these samples suggests the OSL signal was dominated by few grains, as has been the case elsewhere (e.g. Evans et al., 2017). All measured $D_o$ distributions were asymmetrically distributed and displayed a high over-dispersion (OD) (Table 2) confirming heterogeneous bleaching before burial. The $D_e$ values used for age calculation (Table 2) target the well-bleached component of these heterogeneously bleached $D_o$ distributions and were identified by applying age models. Final $D_e$ values for age calculation were calculated using as appropriate either the MAM (Galbraith applied successfully to glacial sediments elsewhere in the BiIS multiple given doses) for each site. Such an approach has been used to land in Ireland and western Scotland (Ó Cofaigh et al., 2012, 2019; Peters et al., 2015, 2016; Small et al., 2017a; Callard et al., 2018; Wilson et al., 2019). The Bayesian analysis of the MSIS was not straightforward because of the interaction between the MSIS draining the main ice sheet divides and more local ‘Irish’ ice that fed laterally into the ice stream. The recently mapped features in the Malin Sea provided the framework for the context of ice movement (Callard et al., 2018).

The Bayesian modelling was coded using OxCal 4.3 (Bronk Ramsey and Lee, 2013) and applied uniform phase sequence models that were punctuated by boundaries located at well-defined ice limits. Markov chain Monte Carlo (MCMC) sampling was used to build distributions of possible solutions, thereby generating modelling probabilities termed posterior density estimates for all measured ages and boundary limits. The probabilities are the product of the Prior model and the likelihood or measured age probabilities measured for each sample. Each sequence was divided into retreat zones that were coded as a Phase, defined as groups containing age information for sites sharing relationships with the adjacent zones. In the Bayesian analysis though TCN ages at some locations were consistent within a site and could be averaged using a reduced chi-square statistic ($\chi^2/\nu$) (Blevington et al., 2003), here the ages were included individually but grouped within a Phase in the Prior model. Phases were delimited by a series of Boundary commands that generated modelled age probability distributions for major ice limits. Both Sequence models were run to assess outliers in time using a scaling of $10^{-6}$–$10^{6}$ years and Student’s $t$-distributions to describe the outlier distribution (Bronk Ramsey, 2009b).

The Bayesian Prior models cover the ice marginal retreat from maximum limits near the continental shelf breaks draining DBIS and MSIS through a series of well-defined ice margin configurations identified on the seafloor and stepping back on to land in Ireland and western Scotland (Ó Cofaigh et al., 2012, 2019; Peters et al., 2015, 2016; Small et al., 2017a; Callard et al., 2018; Wilson et al., 2019).

### Table 2. Luminescence equivalent dose and age data.

| Site          | Sample | Labcode | Analysis | Grain size ($\mu$m) | DR OD (%) | Total analysed* | n (%) | OD (%) | Age model† | $\sigma$-b | $D_e$ (Gy) | Age (ka) |
|---------------|--------|---------|----------|--------------------|-----------|-----------------|-------|---------|------------|-----------|------------|----------|
| Brockhill     | T6BROCO1 | Shfd15171 | SA 180–250 | 12 | 102 | 60 | 60 | IEU | 0.307 | 59.43 ± 4.37 | 44.4 ± 4.1 | |
| Quarry        | T6BROCO2 | Shfd15013 | SA 212–250 | 96 | 48 | 65 | IEU | 0.307 | 58.97 ± 4.66 | 39.1 ± 3.8 | |
| Glenulra      | T6GULR01 | Shfd15172 | SA 180–250 | 76 | 53 | 45 | IEU | 0.307 | 55.68 ± 2.71 | 25.2 ± 1.9 | |
| Quarry        | T6GULR02 | Shfd15012 | SA 180–250 | 144 | 51 | 65 | IEU | 0.307 | 49.3 ± 2.71 | 24.1 ± 1.9 | |
| Lough         | T6LNAC01 | Shfd15173 | SA 180–250 | 92 | 36 | 40 | IEU | 0.307 | 243 ± 13 | 109 ± 8.4 | |
| Nacung        | T6LNAC02 | Shfd15104 | SA 180–250 | 48 | 25 | 71 | IEU | 0.307 | 63.6 ± 6.32 | 132 ± 11 | |
| Althowney Bay | T7ALTBO2 | Shfd15166 | SA 180–250 | 97 | 35 | 56 | MAM | 0.20 | 54.0 ± 8.16 | 30.4 ± 4.9 | |
| Carey         | T7CARV01 | Shfd15169 | SA 212–250 | 80 | 43 | 65 | MAM | 0.10 | 42.04 ± 2.41 | 22.6 ± 2.4 | |
| Valley        | T7CARV02 | Shfd15018 | SA 212–250 | 6 | 160 | 40 | 37 | MAM | 0.10 | 41.36 ± 2.61 | 22.1 ± 2.4 | |
| Castlereoe    | T7CAST01 | Shfd15167 | SA 212–250 | 72 | 57 | 42 | MAM | 0.10 | 31.91 ± 2.67 | 48.1 ± 4.8 | |
| Fawmmore      | T7FAMWO2 | Shfd15015 | SA 212–250 | 70 | 45 | 58 | MAM | 0.10 | 24.29 ± 2.02 | 38.3 ± 3.8 | |
| Gleshesk      | T7GLEN01 | Shfd15017 | SA 212–250 | 78 | 34 | 65 | MAM | 0.20 | 41.34 ± 5.28 | 25.8 ± 4.2 | |
| Valley        | T7GLEN02 | Shfd15170 | SA 212–250 | 103 | 45 | 70 | MAM | 0.20 | 46.43 ± 7.08 | 27.1 ± 3.7 | |

*Total analysed is the number of small aliquots or single grains measured for a sample, while the column headed ‘n’ is the number of small aliquots of single grains accepted for $D_e$ modelling.
†The age model used, either the Minimum Age Model (MAM) or the internal–external uncertainty (IEU) model.
‡Where the IEU model was used, the first parameter ‘a’ is given in this column. The second parameter ‘b’ is 1.5 for all samples. For samples analysed using the MAM, the value given here is that for sigma b.
Table 3. Previously published BRITICE-CHRONO $^{14}$C ages included in this paper.

| Transect | Cruise-core | Code          | Sample ID      | Latitude  | Longitude | Depth (m) | Sample type         | Stratigraphical context                                                                 | Conventional radiocarbon age (years BP) | ±1σ (radiocarbon yrs BP) | Reference          |
|----------|-------------|---------------|----------------|-----------|-----------|-----------|---------------------|---------------------------------------------------------------------------------------|----------------------------------------|------------------------|-----------------------|
| T6       | CE08-018VC  | UCIAMS-133552 | CE_08-018_CC   | 54.98     | -9.92     | 122       | Mixed forams        | Core catcher                                                                                       | 20 170                                 | 90                     | Ó Cofaigh et al. (2019) |
| T6       | JC106-92VC  | Beta432793    | T6-92VC-258cm  | 54.405517 | -9.1768   | 75        | Foraminifera        | mixed benthic                                                                                      | 16 250                                 | 60                     | Schiele (2017)         |
| T6       | JC106-97VC  | Beta432794    | T6-97VC-468cm  | 54.454783 | -9.17203  | 75        | Foraminifera        | mixed benthic                                                                                      | 16 350                                 | 60                     | Schiele (2017)         |
| T6       | JC106-099VC | UCIAMS-164429 | T6-099VC-474   | 54.60363  | -9.33564  | 99        | Foraminifera        |                                                                                                     | 17 180                                 | 80                     | Ó Cofaigh et al. (2019) |
| T6       | JC106-101VC | UCIAMS-164431 | T6-101VC-548-551 | 54.61307 | -9.42068  | 100       | Foraminifera        |                                                                                                     | 20 110                                 | 120                    | Ó Cofaigh et al. (2019) |
| T6       | JC106-102VC | UCIAMS-164437 | T6-102VC-247   | 54.62345  | -9.5189   | 90.5      | Foraminifera        |                                                                                                     | 21 000                                 | 110                    | Ó Cofaigh et al. (2019) |
| T6       | JC106-103VC | SUERC-63558   | T6-103VC-145   | 54.64063  | -9.59722  | 100       | Foraminifera        |                                                                                                     | 22 521                                 | 70                     | Ó Cofaigh et al. (2019) |
| T6       | JC106-112VC | SUERC-63584   | T6-112VC-51    | 54.84513  | -10.18137 | 125       | Shell fragment      |                                                                                                     | 22 582                                 | 67                     | Ó Cofaigh et al. (2019) |
| T6       | JC106-112VC | SUERC-63585   | T6-112VC-59.5  | 54.84513  | -10.18137 | 125       | Shell fragment      |                                                                                                     | 22 572                                 | 71                     | Ó Cofaigh et al. (2019) |
| T7       | JC106-125VC | SUERC-72873   | T7JC106-125VC115 | 55.73367167 | -9.251471389 | 91    | Shell fragment      |                                                                                                     | 22 813                                 | 61                     | Callard et al. (2018)  |
| T7       | JC106-125VC | SUERC-72874   | T7JC106-125VC17 | 55.73367167 | -9.251471389 | 91    | Shell fragment      |                                                                                                     | 22 906                                 | 62                     | Callard et al. (2018)  |
| T7       | JC106-146VC | UCIAMS-176382 | T7-146VC-223   | 56.47296  | -8.70696  | 150       | Foraminifera, mixed assemblage |                                                                                                     | 20 200                                 | 80                     | Callard et al. (2018)  |
|          |             |               |                |           |           |           |                     |                                                                                                     |                                       |                        | (Continued)           |
| Transect | Cruise-core | Code       | Sample ID | Sample depth (cm) | Stratigraphical context                                                                 | Conventional radiocarbon age (years BP) | ±1σ (radiocarbon yrs BP) | Reference |
|----------|-------------|------------|-----------|-------------------|----------------------------------------------------------------------------------------|-----------------------------------------|--------------------------|------------|
| T7       | JC106-146VC | UCIAMS-176383 | T7-146VC-369 | 369–372          | IRD-rich mud, not overconsolidated; Glaciomarine/ice proximal; alternating laminated silt and clay with IRD-rich mud, not overconsolidated | 22 030 | 100 | Callard et al. (2018) |
| T7       | JC106-146VC | UCIAMS-164440 | T7-146VC-389 | 389–392          | Glaciomarine/ice proximal; alternating laminated silt and clay with IRD-rich mud, not overconsolidated | 20 730 | 100 | Callard et al. (2018) |
| T7       | JC106-149VC | SUERC-59509 | T7-149VC-421 | 421               | Soft diamict, possibly glaciomarine/ice proximal – IRD                                   | 17 155 | 47 | Callard et al. (2018) |
| T7       | JC106-151VC | UCIAMS-179841 | T7-151VC-389 | 389–394          | Glaciomarine/ice proximal in a stiffer mud unit with some IRD                           | 19 690 | 90 | Callard et al. (2018) |
| T7       | JC106-153VC | UCIAMS-164432 | T7-153VC-277 | 277–279          | Glaciomarine/ice proximal in a stiffer mud unit with some IRD                           | 19 210 | 110 | Callard et al. (2018) |
| T7       | JC106-154VC | UCIAMS-164433 | T7-154VC-211 | 211–214          | Glaciomarine/ice distal/proximal                                                      | 18 670 | 90 | Callard et al. (2018) |
Table 4. All published BRITICE-CHRONO TCN ages.

| Transect | Code  | Location         | Region          | Latitude | Longitude | Elev. (m) | Sample      | Lithology       | \(^{10}\)Be age (ka)* | CRONUScalc v2.0 \(^{10}\)Be age (ka)* | CRONUScalc v2.0 \(^{36}\)Cl age (ka)* | Reference                  |
|----------|-------|------------------|-----------------|----------|-----------|-----------|-------------|-----------------|------------------------|--------------------------|---------------------------|-------------------------------|
| T6       | T6BS01| Eglish Valley    | Blue Stack      | 54.7228  | –8.1132   | 149       | Boulder     | Conglomerate    | 13.1 ± 0.9 (0.7)    | 13.1 ± 1.3 (0.7)         |                           | Wilson et al. (2019)          |
| T6       | T6BS02| Eglish Valley    | Blue Stack      | 54.7225  | –8.114    | 148       | Boulder     | Sandstone       | 14.9 ± 0.9 (0.7)    | 14.9 ± 1.4 (0.7)         |                           | Wilson et al. (2019)          |
| T6       | T6BS03| Eglish Valley    | Blue Stack      | 54.7231  | –8.1165   | 150       | Boulder     | Conglomerate    | 14.4 ± 0.8 (0.4)    | 14.4 ± 1.2 (0.4)        |                           | Wilson et al. (2019)          |
| T6       | T6BS04| Eglish Valley    | Blue Stack      | 54.7225  | –8.115    | 163       | Boulder     | Conglomerate    | 13.1 ± 0.9 (0.7)    | 13.1 ± 1.3 (0.7)         |                           | Wilson et al. (2019)          |
| T6       | T6GCS02| Glencolumkille | SW coast        | 54.7079  | –8.7589   | 36        | Boulder     | Schist (Qtz vein) | 17.3 ± 1.1 (0.7)    | 17.3 ± 1.6 (0.7)        |                           | Wilson et al. (2019)          |
| T6       | T6GCS03| Glencolumkille | SW coast        | 54.7076  | –8.7589   | 34        | Boulder     | Schist (Qtz vein) | 16.5 ± 1.0 (0.7)    | 16.5 ± 1.5 (0.7)        |                           | Wilson et al. (2019)          |
| T6       | T6PG01| Poisoned Glen   | Donegal         | 55.0150  | –8.10675  | 73        | Boulder     | Granite         | 17.2 ± 1.1 (0.8)    | 17.2 ± 1.6 (0.8)        |                           | Wilson et al. (2019)          |
| T6       | T6PG04| Poisoned Glen   | Donegal         | 55.01495 | –8.10572  | 75        | Boulder     | Granite         | 16.2 ± 1.0 (0.8)    | 16.2 ± 1.5 (0.8)        |                           | Wilson et al. (2019)          |
| T6       | T6PG05| Poisoned Glen   | Donegal         | 55.01498 | –8.10572  | 75        | Boulder     | Granite         | 13.0 ± 0.9 (0.6)    | 13.0 ± 1.2 (0.6)        |                           | Wilson et al. (2019)          |
| T6       | T6ROS01| Rosguill        | Donegal         | 55.2269  | –7.84304  | 65        | Boulder     | Granite         | 18.7 ± 1.0 (0.6)    | 18.7 ± 1.6 (0.6)        |                           | Wilson et al. (2019)          |
| T6       | T6ROS02| Rosguill        | Donegal         | 55.2252  | –7.84062  | 105       | Boulder     | Granite         | 21.4 ± 1.4 (1.0)    | 21.0 ± 2.0 (1.0)        |                           | Wilson et al. (2019)          |
| T6       | T6ROS04| Rosguill        | Donegal         | 55.22412| –7.84055  | 105       | Boulder     | Granite         | 18.9 ± 1.0 (0.6)    | 18.9 ± 1.6 (0.6)        |                           | Wilson et al. (2019)          |
| T6       | T6REN01| Ben Bulben      | Sligo           | 54.36215 | –8.4939   | 204       | Boulder     | Sandstone      | 13.0 ± 0.7 (0.5)    | 13.0 ± 1.1 (0.5)        | Schiele (2017)             |
| T6       | T6REN02| Ben Bulben      | Sligo           | 54.361967| –8.49217  | 198       | Boulder     | Sandstone      | 14.3 ± 0.8 (0.5)    | 14.3 ± 1.2 (0.5)        | Schiele (2017)             |
| T6       | T6REN03| Ben Bulben      | Sligo           | 54.363433| –8.49533  | 203       | Boulder     | Sandstone      | 14.3 ± 0.8 (0.5)    | 14.3 ± 1.2 (0.5)        | Schiele (2017)             |
| T6       | T6REN04| Ben Bulben      | Sligo           | 54.363617| –8.494717| 200       | Boulder     | Sandstone      | 15.7 ± 0.9 (0.5)    | 15.7 ± 1.4 (0.5)        | Schiele (2017)             |
| T6       | T6KC01 | Kikar            | Donegal         | 54.6187  | –8.6096  | 50        | Boulder     | Dolerite       | 18.2 ± 1.7 (0.7)    |                           | Wilson et al. (2019)       |
| T6       | T6KC02 | Kikar            | Donegal         | 54.6187  | –8.6096  | 50        | Boulder     | Dolerite       | 37.5 ± 6.3 (2.1)    |                           | Wilson et al. (2019)       |
| T6       | T6KC03 | Kikar            | Donegal         | 54.6187  | –8.6096  | 50        | Boulder     | Dolerite       | 42.8 ± 6.1 (2.0)    |                           | Wilson et al. (2019)       |
| T6       | T6KC04 | Kikar            | Donegal         | 54.6189  | –8.609  | 45        | Boulder     | Dolerite       | 37.4 ± 5.4 (1.7)    |                           | Wilson et al. (2019)       |
| T6/T7    | T7MH02| Malin Head       | N coast         | 55.38112 | –7.37255  | 65        | Bedrock     | Quartzite     | 23.2 ± 1.4 (0.9)    | 23.2 ± 2.1 (0.9)        |                           | Wilson et al. (2019)       |
| T6/T7    | T7MH03| Malin Head       | N coast         | 55.38156 | –7.37716  | 30        | Bedrock     | Quartz vein (Qtz vein) | 20.7 ± 1.1 (0.7)    | 20.7 ± 1.8 (0.7)        |                           | Wilson et al. (2019)       |

(Continued)
| Transect | Code | Location | Region | Latitude | Longitude | Elev. (m) | Sample | Lithology | \(^{10}\text{Be age (ka)}^*\) | CRONUScalc v2.0 \(^{10}\text{Be age (ka)}\) | CRONUScalc v2.0 \(^{36}\text{Cl age (ka)}\) | Reference |
|----------|------|----------|--------|----------|-----------|----------|--------|-----------|----------------|----------------|----------------|----------|
| T6/T7    | T7MH04 | Malin Head | N coast Donegal | 55.38033 | -7.37458 | 55 | Bedrock | Quartzite | 25.5 ± 2.1 (1.8) | 25.8 ± 2.8 (1.8) |  | Wilson et al. (2019) |
| T7       | T7CAR02 | Caman Mor | Tiree | 56.45521 | -6.92344 | 136 | Boulder | Lewisian gneiss | 21.1 ± 1.2 (0.7) | 21.1 ± 1.8 (0.7) |  | Small et al. (2017a) |
| T7       | T7CAR05 | Caman Mor | Tiree | 56.45464 | -6.92271 | 133 | Boulder | Lewisian gneiss | 20.2 ± 1.1 (0.6) | 20.2 ± 1.7 (0.6) |  | Small et al. (2017a) |
| T7       | T7CAR07 | Caman Mor | Tiree | 56.45236 | -6.91878 | 111 | Boulder | Lewisian gneiss | 20.8 ± 1.1 (0.7) | 20.9 ± 1.8 (0.7) |  | Small et al. (2017a) |
| T7       | T7MIN02 | Mingulay | Mingulay | 56.82096 | -7.63059 | 223 | Boulder | Lewisian gneiss | 18.7 ± 1.0 (0.5) | 18.7 ± 1.6 (0.5) |  | Small et al. (2017a) |
| T7       | T7MIN03 | Mingulay | Mingulay | 56.82096 | -7.63059 | 223 | Bedrock | Quartzite | 21.6 ± 1.1 (0.6) | 21.6 ± 1.8 (0.6) |  | Small et al. (2017a) |
| T7       | T7MIN04 | Mingulay | Mingulay | 56.81998 | -7.63172 | 196 | Boulder | Lewisian gneiss | 17.4 ± 0.9 (0.5) | 17.4 ± 1.5 (0.5) |  | Small et al. (2017a) |
| T7       | T7MIN06 | Mingulay | Mingulay | 56.81521 | -7.63793 | 52 | Boulder | Lewisian gneiss | 19.2 ± 1.0 (0.5) | 19.2 ± 1.6 (0.5) |  | Small et al. (2017a) |
| T7       | T7MIN07 | Mingulay | Mingulay | 56.81521 | -7.63793 | 52 | Bedrock | Quartzite | 20.9 ± 1.1 (0.6) | 20.9 ± 1.8 (0.6) |  | Small et al. (2017a) |
| T7       | T7SGU02 | North Barra | North Barra | 57.05256 | -7.44933 | 65 | Boulder | Lewisian gneiss | 17.4 ± 1.0 (0.6) | 17.4 ± 1.5 (0.6) |  | Small et al. (2017a) |
| T7       | T7SGU03 | North Barra | North Barra | 57.05273 | -7.44955 | 69 | Boulder | Lewisian gneiss | 19.8 ± 1.1 (0.6) | 19.8 ± 1.7 (0.6) |  | Small et al. (2017a) |
| T7       | T7SGU04 | North Barra | North Barra | 57.05349 | -7.45106 | 78 | Boulder | Lewisian gneiss Schist | 17.0 ± 0.9 (0.5) | 17.0 ± 1.5 (0.6) |  | Small et al. (2017a) |
| T7       | T7TMC01 | Torr Mor a’Chonairst | Ross of Mull | 56.28791 | -6.34428 | 42 | Boulder | Granite | 17.3 ± 0.9 (0.5) | 17.3 ± 1.5 (0.5) |  | Small et al. (2017a) |
| T7       | T7TMC05 | Torr Mor a’Chonairst | Ross of Mull | 56.28716 | -6.34287 | 57 | Boulder | Granite | 17.8 ± 0.9 (0.5) | 17.8 ± 1.5 (0.5) |  | Small et al. (2017a) |
| T7       | T7TMC06 | Torr Mor a’Chonairst | Ross of Mull | 56.28617 | -6.3411 | 46 | Boulder | Granite | 17.9 ± 1.0 (0.6) | 18.0 ± 1.5 (0.6) |  | Small et al. (2017a) |
| T7       | S1     | Scriob na Caillich | Jura | 55.9176 | -6.0509 | 106 | Boulder | Quartzite | 17.6 ± 1.2 (0.8) | 17.5 ± 1.6 (0.8) |  | Small et al. (2017a) |
| T7       | S2     | Scriob na Caillich | Jura | 55.9172 | -6.0512 | 106 | Boulder | Quartzite | 16.5 ± 1.1 (0.8) | 16.4 ± 1.5 (0.8) |  | Small et al. (2017a) |
| T7       | S3     | Scriob na Caillich | Jura | 55.9176 | -6.0522 | 92 | Boulder | Quartzite | 15.0 ± 1.1 (0.8) | 14.9 ± 1.4 (0.8) |  | Small et al. (2017a) |

*Calculated with calculator formerly known as the CRONUS-Earth calculator (Developmental version: Wrapper script 2.3, Main calculator 2.1, constants 2.2.1, muons 1.1; Balco et al., 2008) with LM scaling method, Loch Lomond reference production rate (LLPR) (see text), 1 mm ka\(^{-1}\) erosion rate, and one sigma external uncertainty (internal in parentheses).

†Calculated with CRONUScalc v2.0 (Marrero et al., 2016) with LM scaling method, default global reference production rate, 1 mm ka\(^{-1}\) erosion rate, and one sigma external uncertainty (internal in parentheses).
Table 5. All legacy and other TCN ages published after the beginning of the BRITICE-CHRONO project and included in the Bayesian age modelling.

| Site          | Sample   | $^{10}$Be (ka)* | CRONUScalc v2.0 $^{10}$Be (ka)* | CRONUScalc v2.0 $^{36}$Cl (ka)* | Material and context                                      | Reference         |
|---------------|----------|-----------------|---------------------------------|---------------------------------|----------------------------------------------------------|-------------------|
| Donegal       | Malin Head |                 |                                 |                                 |                                                          |                   |
| Bloody Foreland | BF-01    | 21.2 ± 1.1 (1.0) | 21.0 ± 2.0 (1.1)                |                                 | Glacially smoothed quartzite bedrock                      | Bowen et al. (2002) |
|               | BF-02    | 18.5 ± 0.9 (0.8) | 18.6 ± 1.7 (0.9)                |                                 |                                                          | Ballantyne et al. (2007) |
| Bloody Foreland | BF-04-01  | 17.9 ± 1.7 (1.6) | 18.0 ± 2.3 (1.6)                |                                 | Glacially transported granite boulder                     |                   |
|               | BF-04-03  | 33.5 ± 2.7 (2.6) | 34.0 ± 4.0 (2.9)                |                                 |                                                          |                   |
|               | BF-04-04  | 21.8 ± 1.6 (1.5) | 22.0 ± 2.4 (1.7)                |                                 | Glacially transported granite boulder                     |                   |
|               | BF-04-05  | 21.2 ± 1.7 (1.6) | 21.4 ± 2.5 (1.6)                |                                 |                                                          |                   |
|               | BF-04-06  | 21.2 ± 1.9 (1.9) | 21.4 ± 2.7 (2.1)                |                                 | Glacially transported granite boulder                     | Clark et al. (2009c) |
|               | BF-04-08  | 23.6 ± 2.0 (1.9) | 23.8 ± 2.8 (2.1)                |                                 |                                                          |                   |
|               | BF-04-09  | 21.7 ± 2.1 (2.0) | 21.9 ± 2.8 (2.2)                |                                 | Glacially transported granite boulder                     |                   |
|               | BF-04-10  | 22.1 ± 2.0 (2.0) | 22.3 ± 2.8 (2.2)                |                                 |                                                          |                   |
| Average       |          | 21.6 ± 0.7      | 21.7 ± 1.8                      |                                 | Glacially transported granite boulder                     |                   |
| Aran Island    | ARAN01   | 21.8 ± 0.9 (0.7) | 21.6 ± 1.9 (0.7)                |                                 |                                                          | Cullen (2012)     |
|               | ARAN02   | 21.5 ± 0.9 (0.7) | 21.3 ± 1.8 (0.7)                |                                 |                                                          |                   |
| Average       |          | 21.7 ± 0.8      | 21.5 ± 1.8                      |                                 |                                                          |                   |
| Glencolumbkille | MAL-03  | 17.8 ± 0.6 (0.5) | 17.9 ± 1.5 (0.5)                |                                 | Vein quartz in glacially transported schist boulder       | Ballantyne et al. (2007) |
|               | MAL-05   | 19.6 ± 0.7 (0.5) | 19.8 ± 1.7 (0.6)                |                                 |                                                          |                   |
| Errigal col    | ERGL-COL-01 | 17.6 ± 0.8 (0.6) | 17.4 ± 1.5 (0.6)                |                                 | Glacially plucked quartzite bedrock                      |                   |
|               | ERGL-COL-02 | 18.2 ± 0.7 (0.6) | 18.0 ± 1.5 (0.6)                |                                 |                                                          |                   |
|               | ERGL-COL-04 | 18.1 ± 0.8 (0.6) | 17.9 ± 1.6 (0.6)                |                                 | Glacially plucked quartzite bedrock                      |                   |
| Average       |          | 18.0 ± 0.6      | 17.8 ± 1.4                      |                                 |                                                          |                   |
| Sleeve League | SL-02    | 17.1 ± 0.8 (0.7) | 16.9 ± 1.5 (0.7)                |                                 | Quartzite boulder from rockslope-failure debris          | Ballantyne et al. (2013) |
|               | SL-03    | 17.8 ± 1.0 (0.9) | 17.6 ± 1.7 (0.9)                |                                 |                                                          |                   |
|               | SL-04    | 17.1 ± 1.0 (0.9) | 16.9 ± 1.6 (0.9)                |                                 | Quartzite boulder from rockslope-failure debris          |                   |
| Average       |          | 17.3 ± 0.6      | 17.1 ± 1.5                      |                                 |                                                          |                   |
| North Mayo     | OX-03-01 | 16.9 ± 1.4 (1.4) | 17.0 ± 2.0 (1.5)                |                                 | Vein quartz in glacially transported gneissic boulder     | Clark et al. (2009a) |
|               | OX-03-02 | 15.7 ± 1.5 (1.4) | 16.0 ± 2.0 (1.6)                |                                 |                                                          |                   |
|               | OX-03-03 | 16.4 ± 1.3 (1.3) | 16.4 ± 1.9 (1.4)                |                                 | Vein quartz in glacially transported gneissic boulder     |                   |
|               | OX-03-05 | 16.9 ± 1.3 (1.2) | 16.9 ± 1.9 (1.4)                |                                 |                                                          |                   |
|               | OX-03-06 | 17.0 ± 1.7 (1.7) | 17.0 ± 2.3 (1.8)                |                                 | Vein quartz in glacially transported gneissic boulder     |                   |
| Average       |          | 16.6 ± 0.6      | 16.7 ± 1.5                      |                                 |                                                          |                   |
|               | OX-03-07 | 19.1 ± 1.6 (1.5) | 19.1 ± 2.3 (1.7)                |                                 | Vein quartz in glacially transported gneissic boulder     |                   |
|               | OX-03-09 | 20.9 ± 1.5 (1.4) | 21.1 ± 2.3 (1.6)                |                                 |                                                          |                   |
|               | OX-03-10 | 20.5 ± 1.9 (1.8) | 20.7 ± 2.6 (2.0)                |                                 | Vein quartz in glacially transported gneissic boulder     |                   |
| Average       |          | 20.2 ± 1.1      | 20.3 ± 1.9                      |                                 |                                                          |                   |

(Continued)
| Site                        | Sample | Material and context                      | Reference                                      |
|----------------------------|--------|-------------------------------------------|-----------------------------------------------|
| Hebrides and Scottish mainland | D1     | Glacially transported granite boulder     | Finlayson et al. (2014)                       |
|                            | D2     | Glacially transported granite boulder     | Stone and Ballantyne (2006)                   |
|                            | BM2    | Strongly re-moulded granite boulder       | Ballantyne et al. (2014)                      |
| South Uist: Beinn Mhor col | BM2    | Quartzite boulder                         |                                                |
|                            | SNC02  | Quartzite boulder                         |                                                |
|                            | SNC06  | Quartzite boulder                         |                                                |
|                            | SNC07  | Quartzite boulder                         |                                                |

Calculated with CRONUScalc v2.0 (Marrero, 2016) with LM scaling method, default global reference production rate, and one sigma external uncertainty (internal in parentheses).

† Calculated with CRONUScalc v2.0 (Marrero, 2016) with LM scaling method, default global reference production rate, and one sigma external uncertainty (internal in parentheses).

Probabilities for subsequent iterations until the overall model agreement was >60%. Samples handled as outliers (p = 1; 100%) are detailed in later sections.

Results and interpretations

This section presents the new OSL age assessments from land areas adjacent to the DBIS and MSIS (Figs. 2–11; Tables 1 and 2). In addition, we summarize 71 radiocarbon ages from offshore glacial and glaciomarine sediments previously presented in Schiele (2017), Callard et al. (2018) and Ó Cofoigh et al. (2019); and 41 TCN ages already included in Schiele (2017), Small et al. (2017a) and Wilson et al. (2019); and the legacy ages published previously (Small et al., 2017b) that have been included in the Bayesian age modelling. The ages presented here may differ slightly to original published ages owing to differences in exposure-age calculations and statistical treatments (e.g. evaluation using the LLPR; Fabel et al., 2012; Small and Fabel, 2015).

Offshore geomorphology and dating

The geology of the Malin Shelf is characterized by a series of northeast-trending troughs and basins, and basement blocks. Home to the former MSIS, these over-deepened troughs and basins interlink from the Sea of Hebrides to the mid-shelf and were probably major flow paths for ice streaming across the Malin Shelf from the Scottish Highlands and Ireland during past glacial periods (Davies et al., 1984; Dobson and Whittington, 1992). Two basins, the Malin Deep and the trough of the Sea of Hebrides, are separated by the Stanton Bank, a bedrock high at the centre of the inner Malin Shelf (Dobson and Whittington, 1992). For the former DBIS, the shelf offshore NW Ireland in the southern part of the Malin Sea has a smoother profile with a gentle gradient from the mouth of Donegal Bay to the shelf edge, with Donegal Bay having the characteristics of an over-deepened basin like those further to the north (Fig. 1).

For the former MSIS, the geomorphological evidence shows the presence of a compound ridge close to the shelf edge comprising a series of moraines and grounding-zone wedges (GZW’s) mapped from 55°30′N to 56°30′N (Dunlop et al., 2010; Callard et al., 2018). Further to the north, a series of morainal banks with a similar N-S orientation have been broadly mapped from seismic data down to 150 m water depth and are likely to be the continuation of the same ice margin and to be related to the extension of the Outer Hebrides Ice Cap on the Scottish continental shelf (Bradwell et al., 2021). Moraines of different orientations are observed at the boundary between the DBIS (T6) and MSIS (T7) on the Malin Shelf (trending, respectively NW–SE and NE–SW; Figs. 1 (inset), 13 and 14) and it was suggested that they mark the retreat of the two ice threshold advocated by Bronk Ramsey (2009a). Thereby outliers were given a probability scaling of p < 0.2, p < 0.5, p < 0.75 and p = 1 (100%) on a scale of increasing outlier severity. Dating bottlenecks in the Prior models were handled by increasing iteratively the outlier probability for all ages in selected Phases until the model produced overall agreement, which then calculates model agreement indices for all individual ages. Outlier ages were identified statistically, and then scrutinized for reasons that might explain the outlier behaviour either in the Prior model (e.g. the sample context) or in the measurement data (e.g. nuclide inheritance). Ages were not excluded arbitrarily but identified statistically and then weighted p = 1. Cycles of the Bayesian modelling then continued decreasing and increasing other less severe outlier probabilities for subsequent iterations until the overall model agreement was >60%. Samples handled as outliers (p = 1; 100%) are detailed in later sections.
streams in the direction of the inner Malin Shelf to the north and northern Donegal to the south (Benetti et al., 2010; Dunlop et al., 2010; Ó Cofaigh et al., 2012). In the inner part of the Malin Shelf a series of smaller recessional moraines and GZWAs step back eastwards and become increasingly abundant on the inner shelf, with De Geer moraines in the shallower waters of the sea-lochs and sounds, marking the pattern and direction of retreating ice (Dunlop et al., 2010; Small et al., 2016). It has been suggested that, because of the presence of the over deepened troughs, retreat from the shelf back towards the Inner Hebrides was probably rapid (Dove et al., 2015), although previously estimated ice sheet retreat suggests that this process was slow (Clark et al., 2012). For the DBIS, a set of arcuate, nested moraines extend across the entire continental shelf from within Donegal Bay to the shelf edge up to a distance of 90–120 km from the coastline (Benetti et al., 2010; Dunlop et al., 2010; Ó Cofaigh et al., 2012), and they are indicative of grounded ice and a stepped glacial retreat across the shelf.

The dating of these glacial and glacially derived landforms and sediments provides key datasets to support a more refined chronological reconstruction of the behaviour of the two ice streams during the last glaciation (Arosio et al., 2018a; Callard et al., 2018; Ó Cofaigh et al., 2019; Tarlati et al., 2020). Constraining a maximum extent of the BIIS across the Malin Sea has not been straightforward due to the presence of intense iceberg turbation at the shelf edge in correspondence with the margin of the MSIS at the shelf edge. However, the youngest radiocarbon ages obtained from shell fragments in subglacial diamict constrain shelf edge glaciation to after 26.3 ka BP for both the MSIS and the DBIS (JC106-125VC and JC106-112VC in Table 3) (Callard et al., 2018; Ó Cofaigh et al., 2019). Retreat from the shelf edge has been dated using mixed foraminifera assemblages in glaciomarine sediments between 26.3 and 23–24 ka BP and extensive iceberg scouring at the shelf edge across the entire margin of the Malin Sea indicate that it happened initially through intense calving. All foraminiferal and sedimentological data suggest that glaciomarine conditions prevailed during retreat. By 21 ka BP (i.e. global LGM; Clark et al., 2009c; Hughes et al., 2013; Hughes and Gibbard, 2015), most of the Malin Sea was free of grounded ice with glaciomarine conditions recorded offshore Tiree (IC106-149VC; Table 3) (Callard et al., 2018) and a morainic complex of a similar age at the mouth of Donegal Bay (JC106-92VC and JC106-97VC; Table 3) (Ó Cofaigh et al., 2019). Sedimentological evidence from the DBF suggests some marine extension of the BIIS until ~16.5 ka BP that allowed glaciomarine sediment deposition on the fan, with discrete episodes of calving recorded as peaks in ice-rafted debris (Tarlati et al., 2020).

**OSL geochronology**

On the BRITICE-CHRONO project, the timing and pace of deglaciation in other sectors of the BIIS has been in part secured by OSL dating of proglacial and ice-marginal sediments (Evans et al., 2017; Smedley et al., 2017a, 2017b; Bateman et al., 2018; Chiverrell et al., 2018, 2020; Small et al., 2018). Here, we...
report 16 new OSL ages, sampled between 2014 and 2016, from glacigenic sediments at eight terrestrial sites, three associated with DBIS and five from the north of Ireland constraining the MSIS (Tables 1 and 2). Exposures were logged using field sketches, vertical lithofacies logs and photo‐montages following standard procedures (Evans and Benn, 2004; Thomas et al., 2004). Other characteristics recorded included textural classifications, sorting and grain size, palaeocurrents or till fabric indicators, sedimentary structures, nature of contacts and the lithofacies.

OSL sites from the MSIS

OSL samples were collected from natural and quarried sediment exposures extending from in the west Altwinny Bay and Fawnmore (Co. Donegal) progressing west to east to Castleroe, Glenshesk Valley and Carey Valley in Co. Antrim, Northern Ireland (Fig. 1).

Altwinny Bay (55.1432 N, 8.2929 W)

A continuous coastal section is exposed at Altwinny Bay (Cullen, 2012), which is composed of sands, gravels and diamictons (Fig. 2). The sequence, from stratigraphically oldest to youngest, comprises basal laminated gravels, sands and fines that interdigitate with largely massive gravels which are atop a weathered and granite bedrock that has been mobilized glacially. The massive gravels are interpreted as the product of ice‐marginal debris flows, with the more stratified gravel, sand and mud interbeds suggestive of deposition into a water body. Above this, there is a massive diamict containing evidence of deformation including sandy hydro‐fractures injected from above. This in turn is capped by two over‐consolidated matrix‐supported diamictics displaying a strong clast orientation to the south and boulder pavements that suggest a subglacial origin. In the centre of the exposures these subglacial tills are capped by planar cross‐stratified sands,
which have flow directions to the south, probably reflecting outwash deposition with ice margin retreat. These sands appear to have been tilted and deformed, suggesting proximity to and overriding by ice following deposition. The exposures are capped by a further series of matrix-supported diamictons and finally a clast-supported massive gravel with some stratification that is associated either with later re-advance of ice and/or deposition as flow diamiccts during ice retreat.

Cullen (2012) interpreted the sequence to record ice-marginal and glaciomarine debris flows from efflux jets draining ice from inland Donegal. That interpretation conflicts with the exposures observed in 2014, which show growth of the units in a southerly direction, flow directions to the southwest in the outwash sands, and lithologies of erratic clasts in the diamiccts that are all consistent with an ice mass sourced from the Malin Sea rather than inland Donegal. Two OSL samples were collected T7ALTB01 (not measured) and T7ALTB02. T7ALTB01 was taken from a unit of horizontally stratified sand that forms the oldest waterlain deposits identified stratigraphically within the section (Fig. 2B). T7ALTB02 was taken from the youngest water-lain deposit in the sequence, which was composed of deformed (tilted) planar cross-stratified sands (Fig. 2F). These two samples were the most westerly onshore materials collected for the MSIS.

The asymmetrical $D_0$ distribution (Fig. 3A) derived for T7ALTB02 (Shid15166) suggests heterogeneous bleaching before burial, and that a small proportion of the grains characterizes the minimum dose population. The OSL age determined for T7ALTB02, 30.4 ± 4.9 ka, is considered slightly old relative to the dating of shelf-break glaciation at 26.3 ka BP. That said, the stratigraphic position buried by >8 m of diamiccts shows over-ride by ice and the 30.4 ± 4.9 ka age could constrain the expansion of the MSIS to the coast of NW Donegal. Alternatively, this age, slightly old in the sequence, reflects potentially poor bleaching of the OSL signal, which would not be surprising given the relatively short sediment transport distances implicit in an ice contact setting.

Fawnmore (55.1536 N, 8.0329 W)
Located ~10.5 km east of Altwinny Bay, Fawnmore is a sand and gravel pit that has excavated an ice-marginal terrace at ~30 m I.O.D., and has potential to record the step back of the MSIS eastward along the north coastline of Co. Donegal. Two sections were examined in 2014 (Fig. 4). Section 1, although degraded, was composed of sand, gravel and fine-grained units dipping to the southeast. Original observations at Fawnmore (McCabe, 1995) suggest ice retreat to the south, but the southward delta progradation is more consistent with an ice source to the north, a view supported further by the presence of erratic clasts (e.g. basalt).
sourced up-ice within the MSIS. Consequently, this deposit is interpreted as a delta deposited within a lake dammed by the left-lateral margin of the MSIS to the north. Three OSL samples were collected: T7FAWN01 sampling rippled fine to medium sand with fine laminations from Section 1, and from Section 2 horizontally stratified fine–medium sands (T7FAWN02) and fining-upward couplets of rippled to horizontally laminated fine to medium sand (T7FAWN03). T7FAWN02 and T7FAWN03 were priorities for OSL because these were taken from better exposed sediments that indicate deposition as ice proximal delta foresets. Both samples yielded broad $D_0$ distributions (Fig. 3D) suggestive of heterogeneous bleaching before burial, and thus a small proportion of the grains probably characterizes the minimum dose population. The OSL ages determined for T7FAWN02 (Shfd15015) of 25.8 ± 4.2 ka and T7FAWN03 (Shfd15168) of 27.1 ± 3.7 ka are slightly old relative to the geochronology for adjacent zones. These ages show wide distributions reflective of the poor bleaching of the OSL signal, not unexpected given the relatively short sediment transport distances associated with a small ice proximal delta topset.

The Armoy moraine

The Armoy moraine is a major glacigenic landform in the north of Ireland and forms a series of interlinked ridges hummocks and kettle-holes that extend discontinuously for 50 km between Articlave and Ballycastle [Figs. 1 (inset) and 13] (Knight, 2004, 2008a, 2008b). It is generally agreed that the moraine, given the orientation of its arcuate morphology, marks advance of ice from southwestern Scotland into Northern Ireland, but the timing is not well constrained. The
samples collected at Castleroe, Glenshesk Valley and Carey Valley are all distributed along the length of, or immediately down ice from, the moraine. The objective was to constrain the timing of this ice incursion into the north of Ireland.

Castleroe (55.0987 N, 6.6363 W): Within the outwash sands and gravels immediately down ice from the Armoy moraine, 4 km southeast of Coleraine and west of the River Bann, a small dormant sand and gravel pit is set within an undulating bench of glacigenic sediments (Knight, 2004). The sections, when visited in September 2014, showed a fragmentary sequence of what are probably high-energy outwash sands and gravels beneath a massive diamicton containing occasional gravel layers. The sequence is then capped by a unit of clay–silt glaciolacustrine rhythmites containing occasional drop-stones (Fig. 5). Three samples were taken from the middle to lower part of the sequence within the outwash sands and gravels, with T7CAST01 highest in the sequence sampling horizontally stratified coarse sand. Towards the base of a >10-m-thick sequence samples of planar cross-stratified sands with fine laminations (T7CAST02) and rippled and planar cross-set fine to coarse sands with fine laminations (T7CAST03) were taken (Fig. 5). All three samples targeted appropriate lithofacies for OSL dating within the lower and middle part of the sequence, but unfortunately sand-rich facies did not feature within the uppermost glaciolacustrine unit.

Both samples yielded broad $D_a$ distributions with T7CAST02 asymmetrical (Fig. 3C) suggestive of heterogeneous bleaching before burial, and thus a small proportion of the grains characterizes the minimum dose population. The OSL age determined for T7CAST01 (Shfd15015) is too old at 48.1 ± 4.8 ka and pre-dates a younger sample that was taken from lower in the sequence. T7CAST01 sampled a thin sand unit within high-energy gravel outwash laid down potentially in deep channels of back-bar gravel fore-sets, which may have limited the potential for re-setting of the OSL signal. The $D_a$ distribution for T7CAST02 is slightly better behaved with a younger population of aliquots and yielded an age of 38.3 ± 3.8 ka. Chronologically 38.3 ± 3.8 ka pre-dates the MSIS advance to the shelf break (Callard et al., 2018), but the stratigraphical location of these samples beneath 6-m-thick diamicts and evidence for deformation of the outwash sediments samples is intriguing. Taken at face value the T7CAST02 (Shfd15168) age of 38.3 ± 3.8 ka may relate to an earlier advance of the ice sheet during the build up towards the LGM. These older glaciofluvial/deltaic sediments at Castleroe were then incorporated within the Armoy Moraine, with ice advance adding the diamict and the uppermost proglacial glaciolacustrine muds as a lake formed between the MSIS and inland ‘Irish’ ice. The alternative hypothesis is that the Castleroe outwash units are younger and relate to the most recent deglaciation, but where the OSL signals have not been reset for these samples.

Glenshesk Valley (55.1447 N, 6.2211 W): East of the Armoy Moraine and 7 km south from Ballycastle, Glenshesk is one of a series of valleys where water ponded when dammed by ice to the west at Armoy and in the north towards the coast at Ballycastle. Within the Glenshesk valley a set of broad and relatively flat drift surfaces occur, which are stratigraphically above glacigenic features (drumlinized till) associated with Irish ice (Knight, 2008a). These drift surfaces are believed to be associated with water flow and damming between Irish and Scottish ice and could only be deposited when Scottish-sourced ice formed margins at the Armoy moraine (Knight, 2004). A small gravel pit within one of these surfaces reveals it to be composed of distal glaciofluvial sands and gravels, with the uppermost near-surface sequence showing sands capped by planar cross-stratified gravels. Two samples (T7GLEN01 and T7GLEN02) were collected from units of rippled medium sands (Fig. 6), both yielding asymmetrical $D_a$ distributions (Fig. 3E) suggestive of heterogeneous bleaching and a small minimum dose population. The $D_a$ distribution for T7GLEN02 (Shfd1570) is better behaved and yields a younger age of 23.6 ± 3.4 ka, with T7GLEN01 (Shfd15017) 30.4 ± 4.2 ka probably too old. Both samples were lain down in similar environments and so the between-sample differences in signal resetting probably simply reflect the heterogeneity of bleaching in these environments.

Figure 6. (A) Photomontage of the main section at Glenshesk Valley. Lithofacies codes are the same as Fig. 5 (see Evans and Benn, 2004). The labelled boxes show the locations covered by the photographs in B and C. Close-up photographs of the units from which (B) T7GLEN01 and (C) T7GLEN02 were sampled. The circles highlight sample positions. [Color figure can be viewed at wileyonlinelibrary.com]
Carey Valley (55.1918 N, 6.1555 W): Further east still, the Carey Valley is ~6 km east of Ballycastle and ~2 km inland of the present coastline to the north. Situated down-ice from the most eastern end of the Armoy moraine, the valley contains a set of terraced flat-topped surfaces that form a deltaic sequence, which has been subsequently incised. This sequence, which is described by McCabe and Eyles (1988), is composed of two lower diamictons separated by gravelly debris flows. The upper sequence comprises a classic Gilbert-type delta sequence of horizontal fine-grained silty bottom-set units, gently dipping gravel and sand fore-set beds, and planar massive top-set gravels (Fig. 7). These sediments were interpreted to reflect deposition into an open marine setting to the north (McCabe and Eyles, 1988). Given the high elevation of the deposit surface at 113 m above O.D. this seems unlikely and instead we suggest it was deposited within a lake dammed by Scottish ice at the Armoy Moraine near the coast. Two OSL samples were taken from the deltaic sequence, targeting horizontally stratified fine to medium sands in the bottom-set units (T7CARV01) and an upper sample (T7CARV02) from planar cross-stratified sand in the delta fore-sets. Both samples yielded broad and slightly asymmetrical $D_k$ distributions (Fig. 3B) suggestive of heterogeneous bleaching, but contain small minimum dose populations producing similar ages of 22.6 ± 2.4 ka (T7CARV01: Shfd15169) and 22.1 ± 2.4 ka (T7CARV02: Shfd15018).

Taken as a group these sites constraining the Armoy Moraine highlight the challenges of dating heterogeneously bleached materials, but the cluster of three OSL ages ranging from 23.6 ± 3.4 to 22.1 ± 2.4 ka from Glenshesk and Carey Valley are in broad agreement. The OSL ages from Castleroe are interesting, but suboptimal in terms of their stratigraphical position, and the youngest of the age measurements may instead constrain the build-up of regional ice to ~38.3 ± 3.8 ka. Those sediments were perhaps deposited, and then later ridden over by, MSIS ice and incorporated into the Armoy Moraine. Alternatively, given ~38.3 ± 3.8 ka predates evidence for ice-free conditions in western Scotland.
(Jardine et al., 1988), potentially the OSL signals were not reset completely for those samples during the last depositional cycle.

OSL dating sites in the Donegal Mountains and flanking the DBIS

Lough Nacung (55.0405 N, 8.2132 W)
Located in the Donegal Mountains, the Gweedore Valley contains the Clady River which drains these uplands westwards to the coast near Bunbeg (Fig. 1). Immediately downstream of Lough Nacung and south of the Clady River, a large sand and gravel pit has been excavated into dome-shaped low valley-side hillocks at elevations of 93 m (Cullen, 2012). The setting is within the mountain interior of Donegal and glacigenic landforms therein are more likely to relate to the Donegal Ice Dome, though the exit to the valley reaches the coast between the DBIS (T6) and MSIS (T7). The exposures, visited in September 2014, comprised a Gilbert-type deltaic sequence of massive basal gravels, capped by steeply dipping sand and gravel fore-sets, and capped by planar gravel top-sets (Fig. 8). The sequence has been interpreted by Cullen (2012) as subaqueous fan sediments capped by an ice-distal deltaic sequence. The apparent dip direction of the deltaic fore-sets suggests delta progradation towards the northwest. Deltaic sedimentation was probably within a lake dammed to the north and west by coalesced MSIS and DBIS ice masses and fed by ice sourced to the east in Poisoned Glen, Donegal Mountains. Two samples were collected from units of rippled medium sands (T6LNAC01) and rippled medium to coarse sands (T6LNAC02) located towards the top of the fore-sets. These samples potentially constrain sedimentation within a lake that could only have existed while ice was present to the northwest. Both samples yielded broad $D_e$ distributions (Fig. 9C) suggestive of heterogeneous bleaching and contain a small minimum dose population. They produced ages of $109.4 \pm 8.4$ ka (T6LNAC01: Shfd15173) and $132.0 \pm 10.5$ ka (T6LNAC02: Shfd15014) that though similar do not overlap within uncertainties.

The ages for the Lough Nacung delta are substantially too old relative to the LGM shelf-break maxima for the MSIS and DBIS (Peters et al., 2015, 2016; Callard et al., 2018). The lack of evidence for overriding by ice in the sequence in the form of deformation and disruption of the Gilbert-type delta poses questions about the Donegal ice dome. It seems implausible having ice margins at the shelf break and ice-free enclaves in the Donegal Mountains, and so a more likely explanation is poor resetting of the OSL signal in these uppermost fore-set sands of this ice-proximal delta. TCN ages for three glacially transported granite boulders at Poisoned Glen ~8 km up ice from the delta produced a mean age of $16.9 \pm 0.7$ ka and indicate that the Derryveagh Mountains (north Donegal) were largely deglaciated by ~18–17 ka (Wilson et al., 2019). The relatively short sediment transport distances implicit in this ice-proximal delta lend further support to the poor resetting of the OSL signal.

Glenulra (54.3023 N, 9.4330 W)
Located near the coast on the southern flanks of Donegal Bay, the exposures at Glenulra are a small aggregate pit and natural river-cut exposures in part incised probably by glacial meltwater. The exposures show a sequence cut into an ice contact delta with a surface at 80 m I.O.D. (Hallissy, 1911; Hinch, 1913; McCabe et al., 2007a; Ballantyne and Ó Cofaigh, 2017). The sediments at Glenulra Quarry and Farm are an important site for the evolution of the Irish Ice Sheet, though the glaciological interpretation of the sequence and the chronology is equivocal (McCabe et al., 2007a; Ballantyne and Ó Cofaigh, 2017). McCabe et al. (2007a) described a sequence of basal high-density gravelly flows, ~16 m of...
bedded muddy fine-grained units and sands, overlain by 5 m of dipping gravelly delta fore-sets prograding northwards onshore to offshore and capped by planar massive gravel delta top-sets (Fig. 10). Marine fauna occur throughout, and have been 14C dated by analysing a mixture of reworked Arctica islandica shells from the basal gravels and gravel delta top-set, and in situ monospecific Elphidium clavatum from muds interpreted as glaciomarine in origin (McCabe et al., 2007a).

Reconstructions of regional ice flows affecting the Glenulra area show ice generated in the mountains in the southern part of Co. Mayo extended north to Donegal Bay (Synge, 1963, 1965; Greenwood and Clark, 2009a, 2009b). Offshore in Donegal Bay, mapping of submarine landforms affirm the extension of ice northwards from land offshore including a late-stage set of moraines extending from Killala Bay 20 km east of Glenulra (Fig. 14) (Ó Cofaigh et al., 2012), but moraines with geometries reflecting ice extending westwards from the Irish Midlands to the continental shelf break dominate and suggest that the DBIS came close to or impinged on the north coast of Mayo (Ó Cofaigh et al., 2012). McCabe et al. (2007a) interpreted the Glenulra 14C ages as reflecting high relative sea levels from 26 to 45 ka, perhaps discontinuously, but implying substantial isostatic depression. That would require the proximity and some persistence of a thick ice sheet for a significant period before the LGM. Fifteen 14C ages have been obtained for the sequence, with the ages ranging from 21.1 ± 0.2 to 39.5 ± 0.5 14C ka BP. The 11 ages for reworked Arctica islandica shells can only provide maximal constraint on the sequence and the ages may pre-date reworking by millennia.

Ballantyne and Ó Cofaigh (2017) summarize an alternative view that ice cover in Ireland was limited before 32 ka supported by 14C dating of organic and faunal remains from various sites. Were the Arctica islandica shells found at Glenulra reworked from the sea floor in Donegal Bay, those 14C ages imply ice-free conditions in

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those waters before any build-up of land-based ice and advance to shelf-break glaciation 27.8–27.6 ka (Ballantyne and Ó Co-faigh, 2017). The four $^{14}$C ages for monospecific *Elphidium clavatum* from Glenulra form a tighter cluster spanning 23.7 ± 0.1 to 21.1 ± 0.2 $^{14}$C ka BP and include the youngest $^{14}$C age in the sequence. The ages for these foraminifera, if in situ, suggest also significant isostatic depression and proximity to a thick ice sheet 27.8–25.3 cal ka BP (McCabe et al., 2007a). Given the timing for
shelf-break glaciation presented here, the Glenulra 14C ages suggest either (i) the site was not run over by ice during the LGM advance, requiring an implausibly thin DBIS; (ii) there was preservation of the Glenulra deposits under the ice sheet; or (iii) that all the 14C ages are from reworked marine fauna and only provide minimal ages for the deposits (Ballantyne and Ó Cofaigh, 2017). The third scenario potentially still requires high relative sea levels (80 m OD) after 25.3 cal ka BP during deglaciation assuming the deposits are glaciomarine (Ballantyne and Ó Cofaigh, 2017), although a niche glaciolacustrine setting is an alternative hypothesis forming between the DBIS and local ice and thereby receiving reworked glaciomarine fauna. Regional striae patterns on the north Mayo coast (Smith et al., 2008) point to the deflection of ice feeding the DBIS via Bunatrahir and Killala Bays towards the west and northwest.

To address some of these palaeoenvironmental and geochronological uncertainties, this key site was revisited to apply OSL dating to the uppermost deltaic sediments. In November 2014, the upper Glenulra Quarry (54.3023 N, 9.4330 W) sequence displayed the uppermost 3 m comprising a thin diamicton beneath gently dipping sand and gravel fore-sets that were in turn capped by a planar geometry gravelly delta top-set (Fig. 10). The exposures were restricted due to talus and the patchy nature of aggregate extraction, but the dip to the uppermost fore-sets appeared to vary from a W to SW, which differs from McCabe et al. (2007a) who recorded a northerly dip to the fore-sets. A summary conclusion might be that the sediment efflux direction was variable, which supported in Geological Survey Ireland mapping showing a north-flowing down-valley meltwater input, but also coast-parallel west-flowing meltwater channels feeding towards the Glenulra delta (Meehan, 2013). A DBIS origin to the sediment efflux provides a mechanism for the reworking of marine fauna. Two samples (T6GULRK1, T6GULR02) were collected for OSL dating from rippled medium to fine sands with fine laminations. These sampled units are located from the top of the sequence within the gravelly topsets. Both samples constrain potentially sedimentation within either a small ice-marginal lake or proglacial glaciomarine delta flanking Donegal Bay, with two OSL ages that overlap within uncertainties at 25.2 ± 1.9 ka (T6GULRK1: Shfd15172) and 24.1 ± 1.9 ka (T6GULR02: Shfd15012). Both samples yielded asymmetrical Dk distributions (Fig. 9B) suggestive of heterogeneous bleaching, contain a small minimum dose population and are probably maximal ages for the delta. The youngest of these, 24.1 ± 1.9 ka, slightly post-dates though overlaps within uncertainties the youngest of the Glenulra 14C ages at 25.4 ± 0.3 cal ka BP. Regardless, all the chronology from Glenulra is old relative to the DBIS retreat sequence, and we favour an interpretation that the fauna is constrained by the unzipping of ice retreating inland into Co. Mayo. Both samples yielded asymmetrical Dk distributions (Fig. 9A) suggestive of heterogeneous bleaching, contained a small minimum dose population and produced ages of 44.4 ± 4.1 ka (T6BROC01: Shfd15171) and paired small aliquot and single grain (SG) measurements for the second sample of 39.1 ± 3.8 ka (T6BROC02: Shfd15013) and 45.8 ± 8.2 ka (T6BROC02: Shfd15013-1). There is no real evidence for subsequent overriding by ice, and so the most likely explanation is poor resetting of the OSL signal given the relatively short sediment transport distances implicit in this ice proximal delta.

### Synthesis of onshore ages

The Bayesian age modelling uses the new geochronological data obtained during the BRITICE-CHRONO project (Table 4) and already published data (Schiele, 2017; Small et al., 2017a; Wilson et al., 2019), alongside clusters of previously published geochronological information at several onshore locations in Scotland and Ireland (Tables 5 and 6). These are predominately TCN ages (Table 5) but include some radiocarbon ages from various organic material recovered in mostly glaciomarine sediments in coastal proximal settings (Table 6).

Legacy TCN research from before BRITICE-CHRONO includes the Bloody Foreland moraine and other sites in the Donegal and Ox mountains (Ballantyne et al., 2007; Clark et al., 2009a, 2009b; Ballantyne and Ó Cofaigh, 2017). In Hebridean ice feeders towards the MSIS, other ages come from Arran, South Uist (Stone and Ballantyne, 2006; Finlayson et al., 2014; Small et al., 2016; Ballantyne and Small, 2019). BRITICE-CHRONO conducted a programme of sampling at 12 suitable locations distributed across the two transects aiming to fill in gaps in the existing datasets or resolve issues with the previous dating (Fig. 1 for locations). Small et al. (2017a) presented 17 10Be exposure ages from glacial bedrock and bedrock at sites across western Scotland within the area drained by the MSIS. These TCN ages include measurements on Tiree, Mull, Jura, Mingulay and Barra. Wilson et al. (2019) presented 20 new 10Be and 26Cl surface exposure ages from six sites in Donegal, including Malin Head, Rosguill, and Poisoned Glen in northern Donegal and Glencolumbkille, Kilcar and Blue Stacks Mountains in southern Donegal, and Schiele (2017) worked on four 10Be samples from Ben Bulben in Co. Sligo. Some TCN samples at the boundary between the two transects have been used for ensuing Bayesian modelling in both transects (Table 7).

Overall, all these ages provide evidence of the timing of the DBIS first landfall across the Malin Sea and ensuing retreat further inland ultimately towards isolated mountain glaciers. Sites around the coastline of Donegal (including Malin Head, Bloody Foreland, Aran Island, Belderg Pier and Fiddauntaw, nanoneen; Tables 5 and 6) indicate that the ice margin around 20.5 ka was at the Donegal and north Mayo coasts. In Scotland, ice landfall occurred first at Tiree at around the same time (20.6 ± 1.2 ka) and slightly later in Mingulay (18.9 ± 1.0 ka) on the Outer Hebrides. The TCN ages at Malin Head were used alongside 13C chronology from Corvish to suggest an early separation of Scottish-sourced ice and Donegal-sourced ice by ~20.7 ka (Wilson et al., 2019). This
Table 6. Previously published legacy $^{14}$C ages.

| Site                     | Code        | Sample type | Stratigraphical context                                                                 | Conventional radiocarbon age (years BP) | $^{1^\sigma}$ (radiocarbon yrs BP) | Quality | Reference                     |
|--------------------------|-------------|-------------|----------------------------------------------------------------------------------------|----------------------------------------|-----------------------------------|---------|-------------------------------|
| **Transect 6**           |             |             |                                                                                       |                                        |                                   |         |                               |
| Derryvree                | BRM-166     | TOC         | Moss-rich mud overlaid by proglacial sands and till                                     | 30 500                                 | 1100                              | Red     | Colhoun et al. (1972)         |
| Belderg Pier, Co. Mayo   | SSR-2713    | Mollusc     | Laminated muds and sands and diamictons interpreted as glaciomarine sediments over glacially striated rock surface | 16 940                                 | 120                                | Green   | McCabe et al. (1986)          |
|                          | AA56703     | Foraminifera|                                                                                        | 16 627                                 | 83                                 | Green   | McCabe et al. (2005)          |
|                          | AA56704     | Foraminifera|                                                                                        | 16 830                                 | 130                                |         |                               |
|                          | AA56706     | Mollusc     |                                                                                        | 16 389                                 | 74                                 |         |                               |
|                          | AA56707     | Mollusc     |                                                                                        | 16 328                                 | 67                                 |         |                               |
| Fiddauntown, Co. Mayo    | SSR-2714    | Mollusc     |                                                                                        | 17 370                                 | 100                                | Yellow  | McCabe et al. (1986)          |
| Kesh Corran Caves, Co. Sligo | OxA-3693   | Bone (red deer) | Faunal remains in a very thin series of earth and clay strata above ‘sterile’ deposit | 13 622.5                               | 136.5                              | Yellow  | Woodman et al. (1997)         |
|                          | OxA-3706    | Bone (brown bear) |                                                                                   | 13 776.5                               | 105.5                              |         |                               |
|                          | OxA-3708    | Bone (wolf) |                                                                                        | 13 030                                 | 118                                |         |                               |
|                          | OxA-5736    | Bone (hare) |                                                                                        | 14 029.5                               | 210.5                              |         |                               |
| **Transect 7**           |             |             |                                                                                       |                                        |                                   |         |                               |
| S sourlie                | SRR3023     | Antler of *Rangifer tarandus* | Fluvialite sediments between two glacial diamictons                                      | 29 900                                 | 430                                | Red     | Jardine et al. (1988)         |
|                         | SRR3146     | Plant debris |                                                                                        | 29 290                                 | 350                                |         |                               |
|                         | AA45968     | Foraminifera | Glaciomarine sediments                                                                 | 16 120                                 | 160                                | Green   | McCabe and Clark (2003)       |
|                         | AA45967     | Foraminifera |                                                                                        | 15 490                                 | 150                                |         |                               |
|                         | AA45966     | Foraminifera |                                                                                        | 16 460                                 | 430                                |         |                               |
|                         | AA33831     | Foraminifera |                                                                                        | 15 425                                 | 95                                 |         |                               |
| West of Islay            | SUERCI 3122 | Shell       | Glaciomarine sediments                                                                 | 13 103                                 | 40                                 | Green   | Peacock (2008)                |
| Loch Indaal              | SUERCI 3123 | Shell       | Glaciomarine sediments                                                                 | 13 054                                 | 39                                 | Green   | Peacock (2008)                |
|                          | SUERCI 3124 | Shell       |                                                                                        | 13 120                                 | 39                                 |         |                               |
| Loch Sunart              | UL2853      | Mollusc (*Pecten maximus*) | Mud with occasional dropstones and pecten in life position interpreted as glacial diamic and sediments indicating glaciomarine and fully marine conditions | 14 020                                 | 210                                | Green   | Baltzer et al. (2010)         |
| Lochgilphead             | OxA-1697    | Shell       | Glaciomarine sediments                                                                 | 14 488                                 | 303                                | Yellow  | Hedges et al. (1989)          |
|                          | OxA-1698    | Shell       |                                                                                        | 14 848                                 | 302                                |         |                               |
Table 7. The modelled boundary limits for the MSIS and DBIS. All boundary ages are expressed as ±1 sigma. Ages marked with an asterisk are identified as outliers that did not influence the modelled outputs.

| Model structure | Age information | Modelled age | Boundary age |
|------------------|-----------------|--------------|--------------|
| Boundary Base | Ice-free Scotland | 31.5 ± 1.2 | 31.5 ± 1.2 |
| Phase Zone 1 | Donegall Bay Ice Stream (T6) | 23.1 ± 0.2 | 23.1 ± 0.2 |
| Phase Zone 2 | Sourlie SRR3146 | 26.8 ± 0.2 | 26.8 ± 0.2 |
| Phase Zone 3 | Phase Zone 1 Derryve | 25.3 ± 0.2 | 25.3 ± 0.2 |
| Phase Zone 4 | Phase Zone 3 | 24.9 ± 0.2 | 24.9 ± 0.2 |
| Boundary BL2 | 23.5 ± 0.3 | 23.5 ± 0.3 |
| Phase Zone 5 | Phase Zone 4 T6ROS01 Rosguill | 19.7 ± 0.2 | 19.7 ± 0.2 |
| Phase Zone 6 | Phase Zone 5 T6ROS01 Rosguill | 19.2 ± 0.2 | 19.2 ± 0.2 |

Note: All boundary ages are expressed as ±1 sigma. Ages marked with an asterisk are identified as outliers that did not influence the modelled outputs.
| Model structure     | Age information | Modelled age | Boundary age | Model structure     | Age information | Modelled age | Boundary age |
|---------------------|-----------------|--------------|--------------|---------------------|-----------------|--------------|--------------|
| Boundary BL6        | T6ROS03 Rosguill | 18.9 ± 1.0  | 19.7 ± 0.3   | MAL05               | 17.8 ± 0.6  | 17.8 ± 0.5 |
| Phase Zone 7        | T7MIN02 Mingulay | 18.7 ± 1.0  | 19.3 ± 0.3   | T6KC-01             | 37.4 ± 5.4  | 17.9 ± 0.7 |
|                     | T7MIN03 Mingulay | 21.1 ± 1.1  | 19.3 ± 0.3   | T6KC-02             | 37.5 ± 6.3  | *           |
|                     | T7MIN04 Mingulay | 17.4 ± 0.9  | 19.2 ± 0.3   | T6KC-03             | 42.8 ± 6.1  | *           |
|                     | T7MIN06 Mingulay | 19.2 ± 1.0  | 19.3 ± 0.3   | T6KC-04             | 37.4 ± 5.4  | *           |
|                     | T7MIN07 Mingulay | 20.9 ± 1.1  | 19.3 ± 0.3   | ERGL-Col-01         | 17.6 ± 0.8  | 17.7 ± 0.6 |
| Boundary BL7        |                  |              | 19 ± 0.3     | ERGL-Col-02         | 18.2 ± 0.7  | 18.1 ± 0.5 |
| Phase Zone 8        | AA45968 Corvish  | 16.1 ± 0.2  | 18.8 ± 0.3   | ERGL-Col-04         | 18.1 ± 0.8  | 18.0 ± 0.6 |
|                     | AA45967 Corvish  | 14.5 ± 0.2  | 18.4 ± 0.2   | SL-02                | 17.1 ± 0.8  | 17.5 ± 0.6 |
|                     | AA45966 Corvish  | 16.5 ± 0.4  | 18.8 ± 0.3   | SL-03                | 17.8 ± 1.0  | 17.8 ± 0.6 |
|                     | AA33831 Corvish  | 15.4 ± 0.1  | 18.3 ± 0.2   | SL-04                | 17.1 ± 1.0  | 17.6 ± 0.6 |
| Boundary BL8        |                  |              | 18.1 ± 0.7   | T6PG-01              | 17.2 ± 1.1  | 17.7 ± 0.6 |
| Phase Zone 9        | S1 Jura          | 17.6 ± 1.2  | 17.1 ± 0.9   | T6PG-04              | 16.2 ± 1.0  | 17.3 ± 0.6 |
|                     | S2 Jura          | 16.5 ± 1.1  | 16.5 ± 1     | T6PG-05              | 13.0 ± 0.9  | *           |
|                     | S3 Jura          | 15.0 ± 1.1  | 15 ± 1       | Boundary BL5         | 16.8 ± 0.5  | *           |
|                     | SNC-06 Jura      | 16.8 ± 1.1  | 16.7 ± 1     | Phase Zone 6         | OX-03-01    | 16.9 ± 1.4  | 16.2 ± 0.6 |
|                     | SNC-07 Jura      | 16.8 ± 1    | 16.7 ± 0.9   | OX-03-02             | 15.7 ± 1.5  | 16.1 ± 0.6 |
|                     | SNC-02 Jura      | 14.0 ± 1.7  | *            | OX-03-03             | 16.4 ± 1.3  | 16.1 ± 0.6 |
|                     | SNC-03 Jura      | 12.3 ± 1.4  | *            | OX-03-05             | 16.9 ± 1.3  | 16.2 ± 0.6 |
|                     | T7SGU02 North Barra | 17.4 ± 1.0 | 17.1 ± 0.9   | OX-03-06             | 17.0 ± 1.7  | 16.2 ± 0.6 |
|                     | T7SGU03 North Barra | 19.8 ± 1.1 | *            | Boundary BL6         | 15.3 ± 0.6  |              |
|                     | T7SGU04 North Barra | 17.0 ± 0.9 | 16.9 ± 0.9   | Phase Zone 7         | T6BS-04     | 14.4 ± 0.8  | 14.5 ± 0.5 |
|                     | T7TMCO1 Torr Mor a’Chonairst | 17.3 ± 0.9 | 17.1 ± 0.9   | T6BS-01              | 13.1 ± 0.9  | *           |
|                     | T7TMCO5 Torr Mor a’Chonairst | 17.8 ± 0.9 | 17.3 ± 0.9   | T6BS-02              | 15.4 ± 1.0  | *           |
|                     | T7TMCO6 Torr Mor a’Chonairst | 17.9 ± 1.0 | 17.3 ± 0.9   | T6BS-03              | 14.9 ± 0.9  | 14.6 ± 0.5 |
|                     | Arran D1         | 16.1 ± 1.0  | 16.2 ± 0.9   | T6BEN01              | 13.0 ± 0.7  | *           |
|                     | Arran D2         | 16.9 ± 1.0  | 16.8 ± 0.9   | T6BEN02              | 14.3 ± 0.8  | 14.5 ± 0.5 |
|                     | SUBRC13122 W Islay | 13.1 ± 0.04 | 15.2 ± 1.6   | T6BEN03              | 14.3 ± 0.8  | 14.5 ± 0.5 |
|                     | SUBRC13123 Loch Indaal | 13.1 ± 0.04 | 15.1 ± 1.2   | T6BEN04              | 15.7 ± 0.9  | 15.7 ± 0.9 |
|                     | SUBRC13124 Loch Indaal | 13.1 ± 0.04 | 15.2 ± 1.4   | Boundary Ice free Midlands | 13.9 ± 0.4 |              |
|                     | UL2853 Baltzer   | 14.0 ± 0.2  | 17 ± 0.9     | Phase Zone 7         | OXA-13706   | 13.8 ± 0.1  | 13.4 ± 0.3 |
|                     |                  | 14.9 ± 1.5  |              | OXA-13693             | 13.6 ± 0.14 | *           |
|                     |                  | 14.9 ± 1.5  | 14.7 ± 1.6   | OXA-13708             | 13.0 ± 0.1  | *           |
|                     |                  | 14.5 ± 0.3  | 14.6 ± 1.5   | OXA-13736             | 14.0 ± 0.2  | 13.6 ± 0.3 |
| Boundary BL10-end   |                  | 14.3 ± 1.8  | Boundary End of sequence | 13.1 ± 0.9 |
Figure 12. Bayesian chronosequence age-model output of dating constraints using OxCal 4.3. (A) The MSIS (T7) and (B) DBIS (T6). The model structure shown uses OxCal brackets (left) and keywords that define the relative order of events (Bronk Ramsey, 2009a). Each original distribution (hollow) represents the relative probability of each age estimate with posterior density estimate (solid) generated by the modelling. Shown are 14C ages (black), OSL ages (orange), cosmogenic nuclide ages (blue) and modelled boundary ages (red). Outliers are denoted by ‘?’ and their probably (P) of being an outlier indicated by low values <5 (95% confidence). Model agreement indices for individual ages show their fit to the model with >60% the widely used threshold for ‘good’ fit (Bronk Ramsey, 2009b). [Color figure can be viewed at wileyonlinelibrary.com]
### Key

- **Measured age probability (hollow)**
- **Modelled age probability (filled)**

**Cosmogenic nuclide age**

**Radiocarbon age**

**OSL age**

**Modelled boundary age**

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**FIGURE 12** Continued
implied that by this time a marine embayment extended eastward along the north coast of Donegal, separating ice flowing north and northeast from the Donegal Ice Centre from the retreating MSIS. The northern mountains of Donegal (Poisoned Glen and Errigal Col) were largely deglaciated by ~18–17 ka (Wilson et al., 2019). By 17.5–16.5 ka the ice margin straddled the fjos, islands and peninsulas of the western seaboard of Scotland, and the Outer Hebrides Ice Cap had shrunk to expose most of the southern Outer Hebridean islands (Small et al., 2017a). In north Co. Mayo, the five younger $^{10}$Be exposure ages from glacially transported boulders within the moraine system on the northern slopes of the Ox Mountains (Table 5) indicate that ice persisted in much or all Donegal Bay and covered southern Donegal as late as 17 ka. By ~15.0 ka the Donegal Ice Centre had shrunk to a small ice cap or ice field of very limited extent on the Blue Stack Mountains (Wilson et al., 2019).

### Bayesian models

Bayesian age modelling of all the dating control for both transects has calculated the timing for the advance and retreat of the DBIS and the MSIS (Figs. 12–14, 16; Table 7). Additional coastal and inland sites with organic remains dated to before the LGM and after deglaciation in both Scotland and Ireland were used to identify ice-free conditions before and after the last glacial advance and are discussed in the next section in the context of the Bayesian models (Table 6). Ultimately both Bayesian analyses produced conformable age models with an overall agreement index of 188% for the DBIS and 119% for the MSIS, both exceeding the >60% threshold advocated by Bronk Ramsey (2009a). Iterative cycles of the Bayesian modelling varying the outlier probabilities led to the identification of the outlier ages shown on Fig. 12. Italics from now on denote the posterior density estimates or modelled ages derived from the Bayesian modelling to distinguish them from the unmodelled individual ages obtained for directly dated samples.

### Malin Sea Ice Stream

Basal constraint on the retreat model for the MSIS is provided by radiocarbon ages obtained for faunal remains and organic deposits in western Scotland denoting ice-free conditions before the advances to LGM limits (Jardine et al., 1988; Bos et al., 2004; Brown et al., 2007). At Sourlie on the Ayrshire coast (Fig. 13) in the inner feeder zone of the MSIS (Finlayson et al., 2010, 2014), organic packets of sediment in cold-stage fluviatile sediments between two glacial diamictoids yielded antler of Rangifer tarandus with the collagen extract dated to 29 900 ± 420 yr (SRR-3023) and plant debris dated to 29 290 ± 350 yr (SRR-3146) (Jardine et al., 1988; Bos et al., 2004). Support for ice-free conditions in the hinterland of the MSIS is provided further east in central Scotland by equivalent organic-rich sediments at Balglass Burn, north of Glasgow (Brown et al., 2007) spanning 39.8–32.8 ka yr. The Bayesian modelling (Fig. 12A) has produced modelled age probability distributions for ice dynamics in the MSIS sector. Organic sites in western Scotland show ice-free conditions around 34.4 ± 1.8 ka and provide maximum constraint on the build-up and extension of ice into the Malin Sea. In zone 1, on the outer shelf, the youngest $^{14}$C ages on shells reworked into over
consolidated diamicts (Callard et al., 2018) constrain shelf break glaciation to $27.9 \pm 2.2$ ka (BL0), before marine fauna in the softer overlying glacimarine diamict indicated rapid retreat to the zone 2 moraines by $26.3 \pm 0.3$ ka (BL1; Fig. 13).

Decline of ice in the more open Malin Sea proceeded with an ice margin $>120$ km wide retreating east reaching BL2 at $23.5 \pm 0.3$ ka, BL3 at $22 \pm 0.3$ ka and BL4 at $21.2 \pm 0.5$ ka (Fig. 13). Deglaciation of zone 2 vacated the Malin Deep (>150 m) and the outer portion of the Hebrides Trough (>150 m) to establish a series of GZW and the BL2 ice margin east and landward of Stanton Banks. Glacimarine sediments in front of BL2 yielded basal $^{14}$C ages ranging from $23.2 \pm 0.3$ to $22.1 \pm 0.3$ ka and denote ice-free conditions on the inner Malin shelf by $23.5 \pm 0.3$ ka (BL2). The constraint on BL3 is provided by TCN and OSL ages from northwest Donegal, with boulders on the Bloody Foreland and Malin Head peninsulas forming a coherent grouping. Two of the Bloody Foreland granite boulder ages were treated as outliers leaving seven consistent TCN ages. The OSL age from Altwinny Bay, notwithstanding the substantial uncertainty, is an outlier in this grouping, and the age of $30.4 \pm 4.9$ ka (T7ALTBO2) is intriguing given that the sand unit sampled was beneath thick diamict units, which reflects later over-riding by ice. It is feasible that the thin outwash pre-dates ice advance and may be better positioned in zone 1 of the Bayesian sequence model. The deltaic deposits at Fawnmore, though on the face of it a little old, are given the wide uncertainties conformable with the Bayesian model. Together, these ages constrain zone 3 ice margin retreat to BL3 by $22 \pm 0.3$ ka. BL4 is constrained by TCN ages from Tiree (Inner Hebrides) and OSL ages from outwash draining into lakes ponded by Scottish ice impinging on the lowlands of the north of Ireland broadly at the Armoy Moraine (Knight, 2004; Knight, 2008a). Evidence of ice-free conditions from a zone 4/5 marine core (149VC) is provided by a shell fragment in a soft diamicton dated to $20.2 \pm 0.2$ cal ka ($\pm$ Callard et al., 2018). These ages constrain BL4 at $21.2 \pm 0.5$ ka (Fig. 13).

BL5 at $20 \pm 0.3$ ka and BL6 at $19.5 \pm 0.3$ ka (Fig. 13) describe the MSIS dividing into increasingly separate lobes with the ice margin in the Sea of Hebrides entering the fjord landscape of western Scotland, and further south, Scottish ice extended across the North Channel impinging on the lowlands north of Ireland. The cluster of TCN ages from Rosguill document the retreat of ice margins from the outer headlands of the north of Ireland into the mountains of Donegal (Wilson et al., 2019), and across the Malin Sea, TCN measurements from Mingulay (southern Outer Hebrides) (Small et al., 2017a) are very similar in age. In the Bayesian model, the Rosguill and Mingulay clusters are conformable as a single grouping, though the overall model performance is better with Rosguill before Mingulay. The pragmatic interpretation is that the BL5 to BL6 limits were established between $20 \pm 0.3$ ka and BL6 at
FIGURE 15 Continued.
Figure 15. For (A) the Donegal Bay Ice Stream and (B) the Malin Sea Ice Stream, all plotted against age (ka), showing, (bottom) the boundary ages (circle and ±1 sigma whisker plots) and retreat zones of the respective Bayesian models and the rates of net axial ice margin retreat. (middle) Modelled palaeo-water depths (relative to present-day bathymetry) for the inner and outer shelf derived from a glacial isostatic adjustment (GIA) model (Bradley et al., 2011) updated to include the latest BRITICE-CHRONO ice sheet reconstruction and accounting for global ice sheet variations. (top) Mean and 95% ice bed elevations from the NEXTMap Elevation Data Suite (https://www.intermap.com/nextmap) and EMODnet bathymetry (www.emodnet-hydrography.eu/). (C) Ice rafted debris (IRD) flux records from marine cores within the Donegal-Barra Fan MD04-2822 (Hibbert et al., 2010) and MD05-2006 (Knutz et al., 2001, 2002) plotted against an updated age model (Waebroeck et al., 2019). Heinrich Events H2 and H1 are highlighted in grey (Bond et al., 1992). (D) Ocean-climate parameters showing (bottom) sea surface temperature (SST) records determined for the North Atlantic using SST (°C) calculated using planktonic foraminifera for core SO82-02 at 59°N, 31°W (red line) (Van Kreveld et al., 2000; Rasmussen et al., 2016) plotted using an updated age model (Waebroeck et al., 2019) and the MD01-2461 site from the Porcupine Seabight at 51.7°N, 12.9°W (blue line) (Peck et al., 2006, 2007). (middle) δ18O concentrations, Greenland Stadials (GS) and Interstadials (GI) from the GISP2 and GRIP Greenland ice cores (Rasmussen et al., 2016) plotted with modelled surface-air temperatures (black line) relative to present for land masses north of ~45°N (Berger and Loutre, 1991). (Color figure can be viewed at wileyonlinelibrary.com)

Figure 16. Overview of modelled isochrones and retreat rates across the two transects of the Malin Sea and Donegal Bay Ice Streams with relevant geomorphological context (from BRITICE Glacial Map v2.0; Clark et al., 2018 and references therein). Dashed lines indicate more tentative isochrone positions due to lack of geomorphological evidence at the required spatial resolution. Background bathymetry and topography from EMODnet data services (https://portal.emodnet-bathymetry.eu/services/). [Color figure can be viewed at wileyonlinelibrary.com]
19.5 ± 0.3 ka (Fig. 13). Boundary limits documenting the step back of increasingly separated ice lobes into the fjords of western Scotland and into the mountains of Donegal integrates evidence distributed across the Malin Sea. BL7 at 19 ± 0.3 ka is constrained between TCN ages in the southern Outer Hebrides (Cullen, 2012; Dove et al., 2015) and 14C dates from Corvish (Donegal) (McCabe and Clark, 2003). There is strong geographical spread to the age constraint on BL8 at 18.1 ± 0.7 ka and BL9 at 14.9 ± 1.5 ka (Fig. 13), and this is supported by an array of dated TCN sites on Mull, Jura, North Barra and Arran. The three ages treated as outliers within zone 9 were two TCN ages from Jura that Ballantine et al. (2014) had previously interpreted as too young owing to the probable burial of the boulders under a former cover of sediment and/or peat. Three more TCN ages obtained more recently from Jura included a further slightly young age (S3-Jura) (Small et al., 2017a) and was also handled as an outlier. Together, four of Jura TCN ages form a coherent set within the Bayesian model. Ultimate deglaciation of the western Scottish Highlands occurred by 14.3 ± 1.8 ka (BL10) (Fig. 13).

Donegal Bay Ice Stream

There are fewer locations in the hinterland of the DBIS that constrain ice-free conditions pre-dating Marine Isotope Stage 2 advances, though Colhoun et al. (1972) described organic freshwater sills and fine sands at Derryvree (Co. Fermanagh; Fig. 14) that nestled between two thick diamict sheets from a road-cut exposure of a drumlin (54.3031 N, 7.4411 W). The Derryvree cold stage organic deposits yielded an age of 30.5 ± 1.1 14C ka yr (Brim-166) and indicate ice-free conditions (Colhoun et al., 1972). Bayesian modelling indicates a maximum constraint on the build-up and extension of ice into Donegal Bay at 35.1 ± 3.2 ka (pre-LGM ice-free conditions; Figs. 12B and 14), in a similar age range to the western coastline of Scotland, further to the north (Jardine et al., 1988; Bos et al., 2004).

In zone 1 (Fig. 14), on the outer shelf, the youngest 14C ages on shells reworked into over-consolidated diamicts constrain shelf break glaciation to 26.6 ± 1.3 ka (BL0) and the establishment of the shelf break moraine (BL1) at 26.3 ± 0.1 ka (BL1). In zone 2, moving landwards, a series of 14C ages from glaciomarine muds constrain ice-free conditions in the outer Donegal Bay across a series of accreted sea floor moraines. These 14C ages with the more landwards zone 3 chronology constrain BL2 to 22.9 ± 0.7 ka. Zone 3 contains a series of nine 10Be ages from Bloody Foreland (Clark et al., 2009a; Wilson et al., 2019) and two from Aran Island (Cullen, 2012; Wilson et al., 2019) both in northwest Donegal. The location of these sites is marginal to both the DBIS and MSIS, and probably developed a suture between the two ice-masses with ice margin retreat. Two of the Bloody Foreland ages plot too young and were handled as outliers, with all the others forming a coherent grouping. These sites constrain deglaciation of the outer headlands and islands of northwest Donegal and correlate with BL3 ice margins in Donegal Bay to 20.5 ± 0.3 ka (Fig. 14). Zone 4 comprises dating of ice-free conditions moving further east into Donegal Bay and a series of marine fauna 14C-dated on the north coast of County Mayo. Our attempt to date the uppermost deltaic deposits by OSL logically suggests that they form part of this cluster but form a clear ‘too old’ outlier in the Bayesian model. The zone 4 chronology and bracketing ages in zone 5 constrain the BL4 limit to 19 ± 0.4 ka. Interestingly, the modelling combines together 14C ages from the Donegal Bay moraine complex and the Killala Bay moraines, thus suggesting that they are not statistically differentiated and therefore part of a single phase of the ice margin. Within this phase, it is possible that the Killala Bay moraines represent a rapid and short-lived advance of an ice tongue from the north Mayo coast due to debuddressing of northward-flowing ice caused by retreat of the DBIS.

Zones 5–7 record the stepping back of ice margins from Donegal Bay into the flanking mountain regions in counties Donegal, Mayo and Sligo. Zone 5 integrates dating information from typically the coastal fringe around the mountains of Donegal and includes six locations yielding 18 TCN ages. These form a coherent grouping in the Bayesian model, with three of four 26Cl ages from Kilcar too old and probably compromised by nuclide inheritance though the fourth age is consistent within that grouping. Elsewhere, one of three 10Be ages from Glencolumbkille, southwest Donegal (MAL-05; Ballantyne et al., 2007; Wilson et al., 2019). Together, these 13 ages form a conformable group and constrain retreat of ice margins on land into the mountains of Donegal by 16.8 ± 0.5 ka (BL5). In zone 6, eight 10Be ages came from the northern Ox Mountains, south of Donegal Bay, and were published originally by Clark et al. (2009b). Later authors have rationalized the division of the ages into two clusters regarding the five younger ages (mean 16.6 ± 0.6 ka) as better constraint on deglaciation, with the older cluster affected by nuclide inheritance (Ballantyne and Ó Coi-faigh, 2017; Wilson et al., 2019). These five ages form a conformable grouping and constrain retreat of ice margins further inland to the Ox Mountains and BL6 by 15.3 ± 0.6 ka. Deglaciation of zone 7 of the DBIS is constrained by TCN ages from Eglish Valley in the Blue Stack Mountains and Binn Ghulbain (Ben Bulben) in County Sligo. These TCN ages form a broadly conformable set, with two of the Eglish Valley ages and two of Binn Ghulbain ages handled as outliers. In total, four TCN from the two localities indicate that by 13.9 ± 0.4 ka (BL7) the mountains of the inner DBIS had deglaciated (Fig. 14).

Discussion

The seafloor geomorphology (Bradwell et al., 2008; Benetti et al., 2010; Dunlop et al., 2010; Ó Coi-faigh et al., 2012, 2019; Howe et al., 2012; Dove et al., 2015; Callard et al., 2018) and terrestrial landforms in western Scotland, the north of Ireland and around Donegal Bay suggest the presence of former ice streaming across both the Malin Shelf and Donegal Bay (McCabe, 2008; Greenwood and Clark, 2009a; Greenwood and Clark, 2009b; Finlayson et al., 2010, 2014; Clark et al., 2012). However, these two adjoining sectors of the former BIIS display clearly different characteristics and rates of retreat during the last glaciation and deglacial period (Figs. 13–16).

The MSIS had a wide ice margin (120 km; Fig. 13) that remained so as the ice retreated across the shelf. The shelf topography is characterized by pronounced areas of deeper water, with normal and adverse slopes corresponding to the major seabed troughs, including the Malin Deep and extensions of the Hebrides Trough (Fig. 16), both separated by the Stanton Banks bedrock high (Lewisian Gneiss) (Dobson and Whittington, 1992). The geomorphological features associated with ice margin retreat across this outer to mid-shelf topography are complex systems of GZW's, while moraines and much smaller GZW's are found mostly in the inner shelf and close to the coastline and are much smaller (Dunlop et al., 2010; Howe et al., 2012; Dove et al., 2015; Callard et al., 2018). Conversely, the DBIS was not as wide (ca 80 km), decreasing in width as the ice margin retreated landward and had a very gently normal-sloped bed (only the innermost part of the bay displays an adverse slope) and a distinct pattern of closely spaced recessional moraines across the shelf (Benetti et al., 2010).
et al., 2010; Ó Cofaigh et al., 2019). Some lateral moraines (Fig. 14) exist in a position that suggests the presence of a distinct small ice lobe extending northwards into the bay at some stage during deglaciation (Benetti et al., 2010; Ó Cofaigh et al., 2019).

From ice-free conditions in the hinterlands of the MSIS and DBIS—33 ka (Colhoun et al., 1972; jardine et al., 1988), glacial landforms and the presence of radiocarbon-dated subglacial diamicts at the shelf edge show that between 26 and 25.6 ka the BIIS had grown to its maximum extent with ice grounded to the shelf edge (Fig. 16). Evidence across the continental shelf of the western BIIS suggests that this ice margin extended also north and south of the Malin Sea, following predominantly the shelf edge at 140–150 m (current) water depth from northern Scotland to northern Porcupine Bank, with coalescing ice from Scotland and Ireland (Benetti et al., 2010; Schiele, 2017; Bradwell et al., 2021; Ó Cofaigh et al., 2021). This recognition that the BIIS extended to the edge of the Malin Shelf led Wilson et al. (2019) to suggest that the Donegal ice dome was of sufficient thickness to have buried all mountain summits. This hypothesis is supported by thermomechanical models of ice-sheet build-up and decay driven by proxy climate data (Hubbard et al., 2009) which predict thick cold-based ice over many summits. There is further support for these ice thicknesses elsewhere in Ireland (Ballantyne et al., 2011; Ballantyne and Stone, 2015; Ballantyne and Ó Cofaigh, 2017; Ballantyne and Small, 2019), and demonstrations that the last ice sheet overtopped all mountain summits in northwest Scotland (Fabel et al., 2012; Ballantyne and Small, 2019). This build-up of ice, from Greenland Stadial (GS) 4 into the beginning of GS-3, occurs relatively early within the context of the global LGM and pre-dated the maximum in global ice volume (Fig. 15). Variations in sea surface temperature across the North Atlantic Ocean and variations in air temperature before the global LGM may indicate that changes in ocean and atmospheric circulation patterns could have resulted in an increase in atmospheric moisture transport from the Equator to the Poles that is concomitant with a cooling at the northern latitudes favouring the accumulation of snow and ice (Khodri et al., 2001; Clark et al., 2009c; Hughes et al., 2013; Lambbeck et al., 2014; Hughes and Gibbard, 2015). In the Malin Sea, shelf edge glaciation appears to be relatively short-lived. By 26.5 ka the ice sheet had already started to retreat from the shelf edge and extensive iceberg scouring at the shelf edge across the entire margin of the Malin Sea indicates that it happened initially through intense calving. This is also before the global LGM and occurred during cold conditions of GS-3. It is possible, as suggested by Ó Cofaigh et al. (2019) and Callard et al. (2018), that this early retreat was related to the growth of the BIIS and driven by local ice loading increasing water depths and promoting calving ice loss rather than by any changes in oceanic and atmospheric temperatures. This early retreat coincides with the timing of Heinrich event 2 and the increased flux of BIIS-sourced ice-rafted debris (IRD) to the Donegal-Barra Fan at both MD04-2822 (Hibbert et al., 2010) and MD05-2006 (Knutz et al., 2001, 2002) (Fig. 15D).

After the maximum extension in Donegal Bay ~26.6 ka, the retreat and pullback of the DBIS margins across the outer shelf was occurring at a rate of ca. 20 m a⁻¹. Subsequently, we observe a clear pattern of episodic retreat and then stabilization of the ice margin each marked by a morainic ridge on the shelf; more than 25 such moraines can be counted across the Donegal Bay shelf and even more are visible in sub-surface geophysical data (Benetti et al., 2010; Ó Cofaigh et al., 2019). For the Malin Shelf, in contrast, an extensive GZW complex (zone 2; Fig. 13) is observed on the outer shelf for the entire width of the ice stream margin (Callard et al., 2018). This outer portion of the MSIS displays one of the lower rates in retreat for the MSIS (18.7 m a⁻¹; Fig. 15B), and this is smaller than the retreat rate for the corresponding zone of the DBIS (Fig. 15A). The timing of formation of the GZWs in this zone is consistent with the reconstructed 600–1500 years for the deposition of GZWs during ice stream retreat in Antarctica before the Holocene (palaeo-Pine Island ice stream) (Jakobsson et al., 2012). After the initial retreat from the shelf edge, there is a switch in the relative magnitude of retreat rates and in the MSIS they are five to ten times faster than the DBIS (Fig. 15A vs. 15B). This could be related to the shape of the underlying bed. The Malin Shelf displays a clear reverse-sloping bed into the Malin Deep and Hebridean Trough (zones 3 and 5: Fig. 13), where we observe retreat rates of ~25–29 m a⁻¹, that could have contributed to an accelerated ice loss compared to the much more gently inclined DBIS bed (Fig. 15A). When grounding lines retreat onto reverse-sloped beds theoretical and numerical models predict that instability of the ice margin can be triggered by increases in ice thickness at the grounding line, which in turns favours an increase in ice flow across it. This mechanism, termed marine ice-sheet instability (MSI), has been advocated in explanations of the dynamics of many West Antarctic outlets (School, 2007; Favier et al., 2014; DeConto and Pollard, 2016). Whether the water depths are sufficient for MSI to have occurred in the Malin Sea remains to be tested. Overall, the rates of retreat across the margin at this time appear to be between 1.5 and 10 times slower than those of other ice streams of the former BIIS, Laurentide Ice Sheet, Fennoscandian Ice Sheet, and for the Greenland Ice Sheet (Winsborrow et al., 2010; Hughes et al., 2012; Stokes et al., 2014; Scourse et al., 2021).

Foraminiferal and sedimentological data developed for the sector suggest that glaciomarine conditions prevailed during ice margin retreat across the Malin Shelf (Callard et al., 2018). Across the entire ice front there is a distinct reduction in retreat rates once the margins reached constrictions in width at the headlands and islands of Donegal and Scotland; this is particularly the case in the mid-Malin Shelf (10 m a⁻¹ in zone 4; Fig. 13) and outer Donegal Bay (2–5.4 m a⁻¹ in zones 3/4; Fig. 14). Within this area, the Donegal Bay Moraine (zone 4; Fig. 14) represents a major stillstand at 20.5–19 ka. The assessment here of ages developed for the Donegal Bay and Killala Bay moraines cannot be differentiated statistically (Fig. 12B) to distinguish the Killala Bay moraines as a temporally distinct readvance as previously suggested (Ó Cofaigh et al., 2012, 2019). Instead, it is likely that all the moraines mapped within zone 4 (Fig. 14) were the product of oscillating ice positions from different source areas around Donegal Bay and formed around the same time. It appears likely that the Donegal and Mayo headlands and underlying bedrock highs visible in the sub-bottom data (Benetti et al., 2010; Schiele, 2017) acted as shallow and constricted pinning points during the retreat thus slowing ice loss (Favier et al., 2012) and favouring the formation of this moraine complex, at this time fed by entirely Irish-based ice, now a separate Donegal Ice Dome. In attempting to resolve the temporal linkages between MSIS and DBIS we highlight a less well-resolved region between Malin Beg and Bloody Foreland, which occupies both the developing suture between the two ice streams during their respective maximum and later retreat. This sector is rendered even more complex by the growing influence of the ice dome over the Donegal mountains on the geomorphology. The exact timing of the separation of Scottish and Irish Ice in the Malin Sea is resolved for the first time here by the MSIS Bayesian model, which brackets it between 20 and 19.5 ka (Figs. 13 and 16). Thus, separation of Scottish and Irish Ice in the Malin Sea occurs quite early during deglacia-
tion, a feature not present in previous reconstructions (see DATED; Hughes et al., 2016). This timing of 20–19.5 ka coincides with equivalent data from the north Irish Sea basin showing the pullback of ice on land in northeast Ireland (McCabe et al., 2007b; McCabe, 2008; Ballantyne and Ó Cofaigh, 2017; Chiverrell et al., 2018). Here we show the reduced contributions of ice from the North Channel into the Irish Sea, which accords with evidence for an ice-free western Irish Sea and the margins of the Irish Sea Ice-stream positioned to the north of the Isle of Man receiving flows solely from SW Scotland (Galloway Hills Ice Dome) and the English Lake District by 20–19 ka (Chiverrell et al., 2018; Scourse et al., 2021). Ice persisted longer over Donegal Bay than on the Malin Shelf. Compared to the DBIS sector, the ice margin of the MSIS was still straddling the entire width of the Malin Shelf, through a series of deep troughs and smaller headlands (Fig. 13). By 20 ka, Tiree was already seaward of the ice margin, but the remainder of the Inner and Outer Hebrides were still glaciated. Rapid retreat in the Minch Trough between 20 and 18.5 ka and the drawdown of ice led to Hebridean ice masses becoming glaciologically independent shortly before ~18.5 ka (Bradwell et al., 2021) and leading to the development of a separate Outer Hebrides Ice Dome (Small et al., 2017a). A differential pattern of retreat developed to the northeast and southeast once the separation of MSIS and DBIS initiated, and the Outer Hebrides Ice Dome became independent. Seismic and bathymetric data behind Stanton Bank show a stepped retreat to the southeast between Tiree and Mull (Callard et al., 2018) but that is not resolved in terms of timing by the BRITICE-CHRONO sampling.

Around 20–18.5 ka, the retreat of the MSIS was proceeding at a slightly slower pace of ~20 m a⁻¹ compared to earlier retreat and which may reflect stabilization of ice margins at constrained fjord mouths of western Scotland. The net MSIS retreat rates are between 10 and 28 m a⁻¹ throughout and do not vary much at all, so the changes in net pace are subtle. That said, there is better geomorphological evidence for pinning and stabilization points, for example the larger GZWs and moraines, so the actual pace of retreat may have included faster and slower episodes not resolved by the net axial ice margin retreat rate data that emerges from the Bayesian age modelling. This is a pattern of retreat observed commonly in marine-based palaeo-ice streams (Ottosen et al., 2005; Shaw et al., 2006; Larter et al., 2009; Winsborrow et al., 2010; Jakobsson et al., 2012; Newton and Huuse, 2017; Bradwell et al., 2021). Between 21 and 15.4 ka, the reduction in the flux of subglacially derived material, measured using radiogenic Pb isotope data, to the continental shelf is interpreted as the result of the break-up of the ice-stream in western Scotland (Arosio et al., 2018a) and glaciomarine conditions are still indicated in the shelf sediments around the Scottish coastline (Callard et al., 2018). Sedimentological evidence from the Donegal-Barra Fan suggests some marine extension of the BIS until as late as ~16.5 ka that allowed glaciomarine sediment deposition on the fan, with discrete episodes of calcite recorded as peaks in IRD between 18 and ~16.5 ka (Tarlai et al., 2020).

Between 19 and ~16.8 ka, an increase in retreat rate, from an average of around 3.7 to 25 m a⁻¹, is however observed in the inner part of Donegal Bay (Zone 5; Fig. 14), inshore of the extensive Donegal Bay and Killala Bay moraine complex that occupies the outer bay. At this location, the reverse sloped bed (Fig. 15B) is likely to have accelerated through MISI processes (Schoof, 2007; Favier et al., 2014; DeConto and Pollard, 2016). The overall driver, beyond instability, of retreat at this stage is unclear as it is happening within GS-2 and therefore atmospheric warming is unlikely to be a significant control (Fig. 15). Lack of significant change in foraminifera assemblages across the Malin Sea also suggest that the final stages of deglaciation were not likely driven by changes in sea temperature but more probably by local sea level changes and/or thinning of the ice sheet. This is supported by modellled water depths for the inner and outer MSIS derived from a glacio-isostatic adjustment model (Bradley et al., 2011) rerun to account for the ice thicknesses from the latest BRITICE-CHRONO ice sheet reconstruction and accounting for global ice sheet variations (Fig. 15A,B). This suggests maximum water depths occurred at 20–16 ka in the later part of GS-2. Retreat to a fully terrestrial based Donegal Ice Dome occurred within 1–1.5 ka after 16.8 ka and corresponds with the timing of Heinrich event 1. Deglaciation at low ground around Donegal Bay was widespread by 15.3 ka when ice-free conditions are also recorded in the Ox Mountains (zones 6 and 7, Fig. 14).

Conclusions

New OSL ages combined with Bayesian modelling of legacy and BRITICE-CHRONO ages along with consideration of their stratigraphic and landform contexts has allowed us to reconstruct ice advance to the continental shelf edge and withdrawal from here and back across the marine to terrestrial transition (Fig. 16). We summarize the main aspects, and the coastal and inland radiocarbon ages show that Donegal and Scotland were ice free at low elevations around 34–35 ka. However, by 27.9–26.6 ka the BIS had reached its maximum extent, reaching the shelf break of the Malin Sea extending distances of ~190 km from the Donegal and ~280 km from the Scottish coastlines. Geomorphological and sedimentological evidence in the form of subglacial diamict, moraines and GZWs show a continuous ice margin developed at the shelf edge, fed by ice flow from two confluent ice streams, the Malin Sea and the Donegal Bay ice streams. Bayesian modelling of the geochronology shows that retreat from maximum started synchronously along the entire shelf edge of the Malin Sea by 26.3 ka. Compared with the onset of ice retreat globally this is surprisingly early. The MSIS retreated at a rate of ~19 m a⁻¹ and the Donegal Bay Ice Stream at ~20 m a⁻¹ both across the outer shelf between 26.3 and 22.5–23 ka. The outer shelf GZWs in the northern part of the Malin Sea and recessional moraines in the southern part, offshore NW Ireland, indicate that episodic retreat was separated by still-stand or oscillation of the ice margins. The Bayesian modelling struggles to resolve the duration of still-stands, but the scale of the landforms suggests some persistence of the ice margins at these locations. By 23–22 ka the outer shelf (an area of about ~25 000 km²) was already free of grounded ice and ice margin retreat continued at a slower net rate across mid-shelf between 23.5 and 20.5 ka, with the ice margin sitting across the central Malin Sea, near the NW Irish coastline, and across the outer part and mouth of Donegal Bay. The separation between Irish-based and Scottish-based ice seems to have occurred just after this time around 20–19.4 ka, leaving behind an autonomous ice dome over the uplands of Donegal. Thereafter, mass loss of ice on the inner Malin shelf was focused along major submarine troughs and took place over the ensuing 2000 years at a net rate of 16–27 m a⁻¹ with an ice margin positioned close to the present coastline within the Sea of Hebrides at 19 ka. In Donegal Bay retreat during this time was punctuated by still-stands building moraines, and retreat occurred at a much slower pace of 2–5.4 m a⁻¹. The Donegal Bay and Killala Bay moraines at the mouth of the bay record a major ice margin stillstand between 20.5 and 19 ka, with the moraines of different orientations suggesting oscillating ice positions driven by different source areas around Donegal Bay. Once the ice margin started retreating further from this position, the rate of retreat drastically accelerated to 25 m a⁻¹, probably due to the reverse-
slope bed in the inner part of the bay. By 17–16 ka ice had retreated onto land and may have persisted as isolated ice caps in both Scotland and Ireland at least until ca. 14.9–13.9 ka.

Our chronologically constrained reconstruction suggests that the early retreat of the marine-terminating western margin of the BIS was initially driven by local ice loading that increased water depths promoting ice losses by calving, rather than forcing by rises in ocean and atmospheric temperatures. Retreat from the mid-shelf to the coastline proceeded at differing paces between ice-streams and was affected by the presence of topographic controls, including pinning points at underlying bedrock outcrops and constrictions between coastal headlands of Scotland and Ireland, and by the presence of reverse-slope beds underneath portions of the ice streams. Thinning of the ice sheet could have also driven the onset of stages comprising relatively more rapid retreat close to the coastlines of Ireland and Scotland. The timing and rates of retreat for the two ice streams seem largely unrelated to global atmospheric and oceanographic changes, except for the final stage transition into ice-free conditions before 14–13 ka.

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Data availability statement

Background bathymetry and topography presented in the figures are available through the EMODnet data services (https://portal.emodnet-bathymetry.eu/services/). Mapping of onshore and offshore glacial landforms is available from BRITICE Glacial Map v2.0 (www.briticemap.org/). Data in Fig. 15 for ice bed elevations are from the NEXTMap Elevation Data Suite (https://www.nextmap.com/nextmap) and EMODnet bathymetry (www.emodnet-hydrography.eu/). All other data newly presented in this paper are available on request from the authors.

Abbreviations. BIS, British Irish Ice Sheet; DBF, Donegal-Barra Fan; DBIS, Donegal Bay Ice Stream; GS, Greenland stadial; GZW, grounding-zone wedge; ICP-MS, inductively coupled plasma mass spectrometry; IEU, internal–external uncertainty; IBD, ice rafted debris; LGM, Last Glacial Maximum; MAM, Minimum Age Model; MCMC, Markov chain Monte Carlo; MISI, marine ice-sheet instability; MSIS, Malin Sea Ice Stream; OSL, optically stimulated luminescence; SG, single grain; TCN, terrestrial cosmogenic nuclide.

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