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Timing and pace of ice-sheet withdrawal across the marine–terrestrial transition west of Ireland during the last glaciation

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ABSTRACT: Understanding the pace and drivers of marine-based ice-sheet retreat relies upon the integration of numerical ice-sheet models with observations from contemporary polar ice sheets and well-constrained palaeo-glaciological reconstructions. This paper provides a reconstruction of the retreat of the last British–Irish Ice Sheet (BIIS) from the Atlantic shelf west of Ireland during and following the Last Glacial Maximum (LGM). It uses marine-geophysical data and sediment cores dated by radiocarbon, combined with terrestrial cosmogenic nuclide and optically stimulated luminescence dating of onshore ice-marginal landforms, to reconstruct the timing and rate of ice-sheet retreat from the continental shelf and across the adjoining coastline of Ireland, thus including the switch from a marine-to a terrasteriologically-based ice-sheet margin. Seafloor bathymetric data in the form of moraines and grounding-zone wedges on the continental shelf record an extensive ice sheet west of Ireland during the LGM which advanced to the outer shelf. This interpretation is supported by the presence of dated subglacial tills and overridden glaciomarine sediments from across the Porcupine Bank, a westwards extension of the Irish continental shelf. The ice sheet was grounded on the outer shelf at ~26.8 ka cal BP with initial retreat underway by 25.9 ka cal BP. Retreat was not a continuous process but was punctuated by marginal oscillations until ~24.3 ka cal BP. The ice sheet thereafter retreated to the mid-shelf where it formed a large grounding-zone complex at ~23.7 ka cal BP. This retreat occurred in a glaciomarine environment. The Aran Islands on the inner continental shelf were ice-free by ~19.5 ka cal BP and the ice sheet had become largely territorially based by 17.3 ka cal BP. This suggests that the Aran Islands acted to stabilize and slow overall ice-sheet retreat once the BIIS margin had reached the inner shelf. Our results constrain the timing of initial retreat of the BIIS from the outer shelf west of Ireland to the period of minimum global eustatic sea level. Initial retreat was driven, at least in part, by glacio-isostatically induced, high relative sea level. Net rates of ice-sheet retreat across the shelf were slow (62–19 m a⁻¹) and reduced (8 m a⁻¹) as the ice sheet vacated the inner shelf and moved onshore. A picture therefore emerges of an extensive BIIS on the Atlantic shelf west of Ireland, in which early, oscillatory retreat was followed by slow episodic retreat which decelerated further as the ice margin became territorially based. More broadly, this demonstrates the importance of localized controls, in particular bed topography, on modulating the retreat of marine-based sectors of ice sheets. © 2021 The Authors. Journal of Quaternary Science Published by John Wiley & Sons Ltd.

KEYWORDS: British–Irish Ice Sheet; glaciomarine sediments; ice sheet extent; ice sheet retreat; Last Glacial Maximum; Porcupine Bank; radiocarbon dating; subglacial till; western Ireland

Introduction

At its maximum around 27 ka cal BP during the last glacial period, the British–Irish Ice Sheet (BIIS) covered Ireland and much of Britain and had an ice volume with a sea level equivalent of ~2.5 m (Clark et al., 2012). The ice sheet was drained by several large ice streams and extended onto the adjoining continental shelf including the North Sea where it coalesced with the Fennoscandian Ice Sheet.
et al. note that the Malin Sea (Fig. 1) provides evidence for a shelf difference of c. 3 m, ostensibly an extensive, rapid and short-lived advance of the BIIS across the continental shelf, typically to the shelf edge, during the Last Glacial Maximum (LGM) (Benetti et al., 2010; Dunlop et al., 2010; Ó Cofaigh et al., 2012a, 2019; Callard et al., 2018, 2020; Peters et al., 2015, 2016, 2020; Roberts et al., 2020; Craven et al., 2021). These studies have shown that the ice sheet retreated in a glacialmarine environment and radiocarbon dates on marine fauna in deglacial glaciomarine sediments constrain the timing of this retreat. The term ‘global Last Glacial Maximum’ (gLGM) is used in this paper to refer to the period 26.5–19 ka when eustatic sea level was at a minimum because of global ice volume being at its highest (Clark et al., 2009). The term ‘local Last Glacial Maximum’ (lLGM) refers to the period when specific ice sheets reached their maximum extent. The ILGM for the BIS was attained at c. 27 ka (Clark et al., 2012; Scourse et al., 2019), and thus slightly earlier than the gLGM.

Working on the continental shelf west of Galway Bay, Peters et al. (2015, 2016) presented geomorphic and sedimentary evidence for extension of an ice lobe about 200 km westwards from the Irish mainland and onto the Porcupine Bank (Figs. 1 and 2), the outermost part of the Atlantic shelf bordering western Ireland. They dated this advance to sometime after 24.1 cal ka BP, hence during the LGM, and proposed that ice was grounded on the bank as late as 21.8 cal ka BP. Subsequently, however, Callard et al. (2020) presented sedimentological and radiocarbon data from the mid- and inner shelf which showed that initial retreat from Porcupine Bank was underway before 24.4 cal ka BP, and that retreat across the mid-inner shelf was interrupted by oscillations or localized readvances of the ice sheet. These data constrained the age of a large composite grounding-zone wedge on the mid-shelf, variously termed the ‘Mid-Shelf Grounding-Zone Complex’ (Callard et al., 2020) or the ‘Galway Lobe Grounding-Zone Wedge’ (Peters et al., 2016), to c. ≥ 23 cal ka BP. However, subsequent TCN and radiocarbon dates from around the Connemara Peninsula (Peters et al., 2016; Callard et al., 2015, 2016, 2020; Peters et al., 2015, 2016, 2020; Roberts et al., 2020; Craven et al., 2021) or the Porcupine Saddle (derived from the Irish word *Boireann* meaning rocky ground) suggests that initial retreat inland took place at 17.5–17.0 ka, suggesting that the ice sheet remained pinned on the Aran Islands/inner shelf for ~2–3 ka before retreating onshore (Roberts et al., 2020).

Recent work on the shelf offshore of northwest Ireland and in the Malin Sea (Fig. 1) provides evidence for a shelf-edge terminating BIS during the last glacial period, with initial retreat dated to 25.9±1.6 ka BP (Callard et al., 2018) and > 24.8±0.6 ka BP (Ó Cofaigh et al., 2019), and hence early in the gLGM. Ice sheet retreat across the shelf was driven by glacioeustatic depression and high relative sea level (RSL). Further south, the Irish Sea Ice Stream reached a maximum extent in the Celtic Sea (Fig. 1) at ~27 ka (Smedley et al., 2017; Scourse et al., 2019), during what was ostensibly an extensive, rapid and short-lived advance (Ó Cofaigh et al., 2012b; Small et al., 2018). Collectively this implies a difference of c. 3–4 ka between the timing of initial BIS retreat offshore of central western Ireland (cf. Peters et al., 2015, 2016), compared to the retreat of marine-based ice sheet outlets to the north and south.

In this paper we present new geophysical and dated sediment core records from the Porcupine Bank which provide insights into the extent and timing of BIS advance and retreat on the Atlantic shelf west of Ireland during the LGM (Figs. 2–10). We then combine these new data with all previous geochronological data (Peters et al., 2016; Callard et al., 2020; Roberts et al., 2020) in a Bayesian temporal model (Bromke Ramsey, 2009) to reconstruct the timing, pace and pattern of BIS withdrawal across the shelf and adjacent coastline. The paper is the final synthesis of the BRITICE-CHRONO project for the Atlantic shelf sector west of Ireland.

### Regional setting

The study area encompasses central western Ireland, and the adjoining continental shelf (Fig. 2). The mountains of Connemara form an area of rugged topography, incised by glacial troughs and corries. The mountains reach a maximum elevation of 814 m and are mostly composed of quartzite with some schist or gneiss. They are separated from the coast by an extensive area of lowland topography developed chiefly on the Caledonian Galway granite (Davies and Stephens, 1978). To the south of Galway Bay, in County Clare (Fig. 2), the topography is developed on Carboniferous limestone and Namurian sandstones and shales. The limestone area of north County Clare comprises the karstic region known as the ‘Burren’ (derived from the Irish word *Boireann* meaning rocky ground). The three Aran Islands at the mouth of Galway Bay (Fig. 2) form the offshore extension of the limestone terrain of the Burren.

The continental shelf extends for about 150 km westwards from the present coastline to the Slyne Trough (sometimes referred to as the Porcupine Saddle) which separates the

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**Figure 1.** Location of the study area in the context of Ireland and Britain, as well as locations referred to in the text. IS = Irish Sea, CS = Celtic Sea, MS = Malin Sea, DB = Donegal Bay, PB = Porcupine Bank, ST = Slyne Trough, PS = Porcupine Seabight, DBF = Donegal-Barra Fan. SF = Sula Sgeir Fan. Location of Fig. 2 is outlined.
mid-shelf from the Porcupine Bank to the west. The Slyne Trough is approximately 70 km wide and reaches water depths of >300 m. Water depths across the inner to mid-shelf are <200 m. The inner shelf is underlain by an offshore extension of the Precambrian metasedimentary rocks of Connemara and the Carboniferous limestone of the Clare Basin (Naylor et al., 1999), overlain in turn by Pliocene and Quaternary sediments.

The Porcupine Bank forms the outermost part of the continental shelf west of Ireland between 51–54°N and 11–15°W. It is located approximately between 150 and 250 km from the Irish coastline and covers an area of more than 40 700 km². It is the westwards projection of the main Irish continental shelf to which it is linked by the Slyne Trough. The Porcupine Bank itself comprises a north-south-trending plateau and is the north-western margin of the Porcupine Seabight Basin (Naylor and Shannon, 2009). Water depths across the bank are typically 200–400 m (Thébaudeau et al., 2016). The SW–NE-orientated Porcupine Ridge in the north forms the shallowest area with water depths of < 200 m, and the shallowest point on the bank overall is 145 m (Thébaudeau et al., 2016). Along the northern and western sides of the bank the shelf break occurs at c. 400 m water depth and the continental slope is steep (Sacchetti et al., 2012). This contrasts with the southern Porcupine Bank where it slopes gently down into the Porcupine Seabight (Fig. 1). Slope gradients over most of the bank are generally < 0.5° (Thébaudeau et al., 2016).

**Methods**

Multibeam bathymetric data were collected by the Geological Survey of Ireland and the Irish Marine Institute during cruises of
the RSV Siren in 2000 and 2001 as part of the Irish National Seabed Survey (INSS) and Integrated Mapping for the Sustainable Development of Ireland’s Marine Resource (INFORMAR) programmes. The bathymetric data were collected using a hull-mounted Simrad EM1002 multibeam system. They provide a geomorphological context for the new seismic records, sediment cores and dating results that we discuss in this paper. The multibeam system has an operational frequency of 93–98 kHz, pulse lengths of 0.5 ms (150–250 m) and 0.7 ms (250–500 m), and decimetric vertical and horizontal accuracy of 50 cm or better according to water depth (Thébaudeau et al., 2016). Data were gridded at a cell size of 25 m. Linear data artefacts are visible at the edges of overlapping survey lines across the study area. Their visual effect was minimized using a range of different sun-illumination angles (Thébaudeau et al., 2016). New information on acoustic stratigraphy and sediment thickness were acquired during cruise JC106 of the RRS James Cook in 2014 using a hull-mounted Kongsberg SBP120 sub-bottom profiler. The SBP120 system uses a transducer with a frequency sweep range of 2.5–7 kHz installed as part of the EM120 wideband receiver array. It has a typical maximum penetration depth of 50 m (depending on the nature of the sediments) and a nominal depth resolution of 0.3 ms. Two-way travel times in seconds were converted to depth below sea level at the time of surveying using typical values of sound velocity (1500 m s⁻¹ through the water column and 1600 m s⁻¹ through soft sediments). The data were visualized in IHS Kingdom.

Sediment cores were collected from across the Porcupine Bank using a 6-m-long British Geological Survey vibrocorer during cruise JC106 (Figs. 2, 3, 8 and 9; Table 1). Core sites were selected to target glacigenic seafloor landforms, particularly grounding-zone wedges. Following collection, the sediment cores were cut into 1-m-long sections, split, and information recorded on grain size, sedimentary and deformation structures, sorting, colour, the nature of bedding contacts, clast abundance and shape, and macrofaunal content. Measurement of sediment shear strength in kPa was recorded using a Torvane. Magnetic susceptibility and wet bulk density were measured on-board during cruise JC106 using a GEOTEK multi-sensor core logger (MSCL). Post-cruise, the cores were X-rayed using a GEOTEK MSCL-XCT scanner to obtain additional information on sedimentary structures. The cores were stored on ship and subsequently in Durham University at 4 °C.

Samples of marine molluscs (typically single or broken valves) and benthic foraminifera were collected for radiocarbon dating (Table 2). Samples typically targeted subglacial to deglacial lithofacies boundaries or, where cores bottomed out in deglacial sediment, the base of the core to obtain a minimum age constraint on ice sheet retreat. In addition, individual reworked shells from subglacial till were dated to provide a maximum age on till formation and thus ice sheet advance.

The radiocarbon dates were corrected for isotopic fractionation and then calibrated using the Marine13 calibration curve that incorporates a standard marine reservoir effect of 400 years. Delta-R values of 0, 300 and 700 years were then also applied. The radiocarbon and calibrated ages with Delta-R of 0 years are used in the text due to spatial and temporal uncertainties in the radiocarbon reservoir ages in the North Atlantic and adjoining continental shelves since the LGM (cf. Wanamaker et al., 2012).

A Bayesian temporal model implemented using OxCal 4.3 (Bronk Ramsey, 2009) was used to analyse the geochronological data and so reconstruct the pattern of deglaciation. The Bayesian modelling follows that described in Chiverrell et al. (2018). The Bayesian approach allows us to integrate different types of chronological data [radiocarbon, TCN and optically stimulated luminescence (OSL); Tables 2, 3 and 4] and to...
Figure 4. Glacial geomorphology of the western Irish shelf and Porcupine Bank as mapped from multibeam swath bathymetric data. Source: INFOMAR (Integrated Mapping for the Sustainable Development of Ireland’s Marine Resource; Geological Survey of Ireland and Marine Institute). (A) Colour-shaded relief bathymetric image of the seafloor of the mid-outer western Irish shelf and Porcupine Bank showing location of subsequent figures; (B) mapped geomorphology of the seafloor shown in panel A and modified from Thébaudeau et al. (2016). PB = Porcupine Bank, ST = Slyne Trough, WIM = West Ireland Moraine. The dashed box outlines the zone where moraines on the Porcupine Bank transition with increasing water depth into grounding-wedges in the Slyne Trough. [Color figure can be viewed at wileyonlinelibrary.com]
Figure 5. Details of grounding-zone wedges in the Slyne Trough and outer Porcupine Bank showing their seafloor morphology in plan-view and associated topographic profiles. Locations shown in Fig. 4. Note the characteristic ‘ramp-step’ form of the grounding-zone wedges and the heavily iceberg-furrowed ramp slopes. (A) Colour-shaded relief bathymetric image of grounding-zone wedges in the Slyne Trough. The arrow marks the location of a crudely circular depression, interpreted as an iceberg wallow pit, which occurs the end of a curvilinear iceberg furrow. (B,C) Topographic profiles across the grounding-zone wedges shown in A. (D) Colour-shaded relief bathymetric image of grounding-zone wedge on outer Porcupine Bank from which core 158VC was recovered. The core site location is shown on the image. (E) Topographic profile across the grounding-zone wedge shown in D. Data source for the multibeam imagery: INFOMAR (Integrated Mapping for the Sustainable Development of Ireland’s Marine Resource; Geological Survey of Ireland and Marine Institute). [Color figure can be viewed at wileyonlinelibrary.com]
identify data outliers. It uses a ‘prior model’ (the order of events) to refine probability distributions when presented as a relative order of events (Chiverrell et al., 2013, 2018). The prior model is constructed using the events or ice-marginal positions interpreted from the geomorphological and stratigraphic evidence, and arranged in the order of ice retreat, determined independently of the geochronological data. The Bayesian analysis then uses this prior model to assess the possibility of outliers but does not automatically reject these depending on the landform and/or stratigraphic context or the measurement. The approach uses the relationship between dated sites [grounding-zone wedges (GZW), moraines, core sites and sediment exposures, as well as erratic boulders and ice-moulded bedrock], and the modelling can reduce the uncertainty ranges for individual age estimates (Bronk Ramsey, 2009; Chiverrell et al., 2018). The individual age measurements (radiocarbon, TCN, OSL; Tables, 2, 3 and 4) are expressed as age probability functions that represent the likelihood that any one sample has a particular age (Bronk Ramsey, 2009).

To explore the timing of ice margin retreat, the age control within our prior model has been divided into a series of Phases, each representing the geochronology either for specific sites or zones of the former ice mass. Each Phase, in the Bayesian terminology (Bronk Ramsey, 2009), contains grouped dating information that shares a common relationship with other items in the prior model, and they are separated by Boundaries that delimit the period of each Phase (Fig. 11). The Boundary command in the software generates modelled age estimates that are then used to define the timing of ice margin retreat. The modelling uses Markov chain Monte Carlo sampling to build up a distribution of possible solutions. For each sample it generates a probability called a posterior density estimate which is the product of both the prior model and the likelihood probabilities. This approach generated modelled ages for boundaries in the prior model that separate a series of ice retreat zones, with each zone coded as a Phase. The sequence model was run in an outlier mode to assess for outliers in time. The Bayesian analysis produced a conformable age model for ice sheet retreat on the shelf offshore of western Ireland with an overall agreement index of 136%, thus exceeding the > 60% threshold advocated by Bronk Ramsey (2009).
Figure 7. Representative sub-bottom profiles of grounding-zone wedges (GZWs) from the study area showing the internal acoustic stratigraphy. Locations of sediment cores discussed in the text are also shown. (A) GZW from the outer Slyne Trough shown in Fig. 5A as profile line B–B’. Note the heavily iceberg-furrowed surface of the GZW; prominent iceberg furrows are arrowed. (B) Back-stepping GZWs on the surface of the West Ireland Moraine from the eastern side of the Slyne Trough (See Fig. 4B). Prominent iceberg furrows are arrowed. Both profiles show the presence of two acoustic stratigraphic units labelled ‘X’ and ‘Y’. The lower unit X is acoustically homogeneous and overlies a diffuse to locally strong and irregular basal reflector. The upper unit Y is a conformable sediment drape that is commonly acoustically transparent internally. [Color figure can be viewed at wileyonlinelibrary.com]

Figure 8. Lithofacies logs, associated shear strength measurements in kPa and calibrated radiocarbon dates from sediment cores from the Porcupine Bank. ‘ID’ refers to radiocarbon dates that were indistinguishable from background. Lithofacies codes are shown and ‘LFA’ refers to the lithofacies associations discussed in the text with the different LFAs colour-coded for ease of identification. Sediment core locations are shown in Figs. 2 and 3. [Color figure can be viewed at wileyonlinelibrary.com]
Results and interpretations

Seafloor geomorphology and acoustic stratigraphy

We use the detailed description of the seabed geomorphology of the Porcupine Bank and Slyne Trough by Thébaudeau et al. (2016), concentrating on the northern sector of the Porcupine Bank, including the outer shelf, and the Slyne Trough. We describe new acoustic stratigraphic data from these areas and discuss their origin and significance in terms of ice sheet advance and retreat.

A series of prominent ridges extend across the Slyne Trough and Porcupine Bank (Figs. 3–7). These ridges are generally orientated east to west and north-east to south-west although there is some variability. We distinguish two characteristic forms: wedges and sharper, locally sinuous, ridges. The wedges occur within the Slyne Trough and across the outer shelf of the northern bank. They are typically characterized by a well-defined asymmetrical ramp-scarp form with a steeper north-facing slope or scarp (up to 3°) and a gentler southern slope or ramp (up to 1.5°) (Figs. 4, 5 and 7). They range from

![Image of seafloor geomorphology and acoustic stratigraphy](image-url)
5 to 30 m in height and are up to 20 km long (across-trough). In planform they are curvilinear to locally arcuate. Although the distribution of wedges is discontinuous across the Porcupine Bank and Slyne Trough, an outer shelf set, located ~15 km inshore of the shelf break, can be distinguished as well as a series of back-stepping wedges to the south and east (Fig. 4). On the eastern side of the Slyne Trough back-stepping, east-west-oriented wedges up to 15 m high can be seen to overprint a much larger ridge orientated NNE–SSW (Figs. 4 and 7B). Cores from these wedges confirmed that they are composed of sediment (see below ‘Sedimentology and radio-carbon measurements’ and Fig. 9).

Sub-bottom profiles from across the wedges show a clear asymmetrical, ramp–scarp form with the steeper scarp face on the north side (Figs. 5 and 7). Sediment thicknesses across the wedges range from 1 to 10 m and are thickest on the southern (ramp) slope and at the base of the frontal scarp faces. The profiles show the presence of two acoustic units: a lower acoustically homogeneous unit, up to c. 20 m thick, that overlies a diffuse and irregular basal reflector, and an overlying 3–5 m of conformably draped sediment (Fig. 7). Internally, the drape is often acoustically transparent but occasionally is diffusely stratified; the wedges are transparent and no internal structure was observed. A characteristic feature of the wedges is that they are incised by furrows which can be up to 10 m deep (Figs. 5 and 7). Furrows are particularly well developed on the wedge crests and ramp slopes, and in places incise both stratigraphic units.

Across the northern Porcupine Bank a series of sharp-crested, locally sinuous or crenulated ridges are visible (Figs. 4 and 6). The ridges are narrow and typically 2–3 km wide, up to c. 6 km long and up to a maximum of 14 m high above the surrounding seabed. Similar to the wedges, these narrower ridges are discontinuous but the individual segments appear to align. The ridges occur predominantly in water depths shallower than c. 300–330 m, below which they transition to wedges (Fig. 4). The internal structure of the ridges could not be resolved in detail due to low acoustic penetration. The acoustic stratigraphy is limited to c. 7 m of sediment comprising a lower unit that is semi-transparent to homogenous with a localized distribution. This is overlain by a more continuous, homogenous unit, with a high-amplitude seabed reflector. Sediment cores from these ridges bottomed out in stiff, sandy, matrix-supported massive diamicton (see below).

Thébaudeau et al. (2016) mapped over 30 000 cross-cutting furrows across the study area. The furrows occur in water depths of 180–575 m and two main orientations are recorded: south to north and east to west. In form, the furrows are linear to curvilinear, up to 300 m wide, 10 m deep and 20 km long. Some of the furrows are bounded by 1–2 m-high berms, either singly or paired. Thébaudeau et al. also described craters or pits up to 15 m deep which occur at one end of individual furrows (e.g. Fig. 5A). As noted above, furrows incise the wedges, and are particularly well developed on the southern (ramp) slope and crest (Figs. 5 and 7).

Interpretation

Sediment cores confirm that both the ridges and wedges are composed of sediment. The size and morphology of the wedges within the Slyne Trough and across the northern Porcupine Bank are consistent with descriptions of GZWs from glaciated continental shelves (e.g. Evans et al., 2005; Ó Cofaigh et al., 2005; Dowdeswell and Fugelli, 2012). This interpretation is based on their asymmetrical ramp–scarp form and, in planform, their curvilinear to arcuate expression on the seafloor. The sharp-crested sediment ridges are narrower than the GZWs and on this basis are interpreted as ice-marginal moraines (Bradwell et al., 2008; Ottesen and Dowdeswell, 2009; Ó Cofaigh et al., 2012a; Batchelor and Dowdeswell, 2015). GZWs are usually inferred to represent formation in a sub-ice shelf cavity where vertical accretion is constrained, thereby producing the ramped form, whereas sharp-crested moraines are commonly inferred to form at a
grounded, tidewater ice margin (Batchelor and Dowdeswell, 2015; Smith et al., 2019). Across the Slyne Trough and Porcupine Bank, GZWs occur predominantly in water depths greater than c. 300–330, whereas the moraines occur in shallower water (Fig. 4). This pattern may reflect a transition from an ice shelf within the deeper water of the Slyne Trough to a grounded (tidewater) margin across the shallower bank.

The back-stepping pattern of the moraines and GZWs indicates that grounding-line retreat across the outer Porcupine Bank was episodic and was characterized by periodic stillstands or oscillations (cf. Dowdeswell et al., 2008; Ó Cofaigh et al., 2008). Oscillatory behaviour is suggested by the arcuate and locally crenulate or sinuous form of the GZWs and moraines (cf. Bradwell et al., 2008). The back-stepping east-west-oriented GZWs that overprint the large NNE–SSW ridge on the east side of the Slyne Trough (Fig. 4) postdate that ridge as they are formed on top of it and they also align with GZWs further to the west. Peters et al. (2016) termed the larger ridge the ‘West Ireland Moraine’ (Fig. 4B) and inferred that this represented a push moraine that was ornamented with smaller superimposed GZWs built by a grounding ice shelf during extension over the Slyne Trough.

The acoustically homogenous to transparent facies that makes up the cores of the GZWs and moraines is similar to previous observations of these landforms from high-latitude continental margins (e.g. Shipp et al., 1999, 2002; Evans et al., 2005; Hogan et al., 2012; Batchelor and Dowdeswell, 2015). This acoustic facies is interpreted as a diamicton probably sourced from a subglacial deforming layer that was advected to the grounding line (cf. Ó Cofaigh et al., 2005, 2007; Evans et al., 2009). Subsequent deposition of this diamicton formed a subglacial traction till, although it is likely that it was also at least partly produced by sediment reworking, either through glacitectonism during minor fluctuations of the grounding line and/or through debris flow processes (cf. Hogan et al., 2016; Evans, 2018). The presence of subglacial till is consistent with sediment cores which bottomed out in stiff, matrix-supported diamicton (see ‘Sedimentology and radiocarbon measurements’ below). The acoustic drapes which overlie the GZWs is inferred to be a product of formation in a deglacial and/or postglacial environment by passive suspension settling of fine-grained sediment through the water column.

Cross-cutting, linear to curvilinear furrows with paired or single lateral berms which incise the surfaces of the GZWs and occur across much of the study area (Thebaudie et al., 2016) are interpreted to be iceberg ploughmarks, produced by the intermittent scouring of grounded iceberg keels in contact with the seafloor (e.g. Woodward-Lynas et al., 1991; Dorschel et al., 2010; Sacchetti et al., 2012). The associated craters, which occur at furrow ends, are probably ‘iceberg wallow’ structures related to the stranding of icebergs (cf. Reimnitz and Kempka, 1982; Longva and Bakkejord, 1990; Stewart et al., 2016). An alternative interpretation for furrows on the ramp slopes of the GZWs is that they were formed by erosion by irregularly shaped ice keels at the base of an ice shelf in the transition zone from grounded to floating ice (e.g. Graham et al., 2010). However, such features typically exhibit high linearity and can terminate abruptly at an asymmetrical mound (Smith et al., 2019). This is at odds with the marked curvilinear form of many of these features on the Porcupine Bank; GZWs and their termination in depressions or pits at one end. Hence an interpretation of iceberg ploughmarks is preferred.

Sedimentology and radiocarbon measurements

Sixteen vibrocores, targeting sediment ridges interpreted as GZWs and moraines on the Porcupine Bank and in the Slyne Trough, were investigated in this study (Figs. 2, 3, 8 and 9). Nine lithofacies are identified based on core lithology, sedimentary structures and physical property measurements. These lithofacies are described below along with dating constraints. Core logs and stratigraphic positions of the dates are shown in Figs. 8 and 9, as well as in Supplementary Information figures a and b. Figure 10 shows x-radiograph examples of the various lithofacies.

Lithofacies 1. Massive, matrix-supported diamicton (DmM)

This lithofacies comprises a massive, matrix-supported, dark grey (SY4/1) diamicton that typically forms the basal lithofacies in cores from the Porcupine Bank (Figs. 8 and 10). The matrix ranges from sandy (e.g. cores 167VC, 176VC) to muddy (e.g. cores 158VC, 177VC) in texture. Clasts from within the diamicton matrix are rounded to sub-angular and are granule to pebble in size. Clast abundance is variable and, as a result, the diamicton ranges from quite clast-rich and gravelly (e.g. 158VC), to more matrix-rich in which small clasts are dispersed throughout the matrix (e.g. 197VC). There is no discernible internal structure, either visually or within the X-radiographs (Figs. 8–10). The diamicton is characteristically stiff with shear strengths ranging from 33 to 200 kPa and averaging 100 kPa. The wet bulk density is also high, averaging 2.3 g ml\(^{-1}\) (Supplementary Information Figs. a and b). The upper boundary is predominantly sharp, and occasionally convoluted. Magnetic susceptibility varies within and between cores, averaging 502 SI. However, this high value is biased by core 176VC that has values consistently above 1300 SI and when it is removed from calculation the average magnetic susceptibility reduces to 214 SI (Supplementary Information Figs. a and b).

The age of the DmM facies is constrained by 29 radiocarbon measurements on shell and coral fragments that were sampled from cores 158VC, 159VC, 167VC, 176VC, 177VC and 197VC (Figs. 8 and 9; Table 2). Of these, seven were indistinguishable from background. Most of the ages are between 36.2 and 25.0 cal ka BP. The youngest ages in this group, 25 510 ± 265 (SUERC-58389), 25 398 ± 206 (SUERC-67932) and 25 014 ± 310 (SUERC-67937) cal ka BP, are from core 158VC. Two cores contain samples which yielded considerably younger ages. An age of 18 440 ± 145 (SUERC-68872) cal ka BP was obtained from an articulated bivalve of Yoldia species towards the top of the diamicton in core 177VC. Core 158VC contains several relatively young ages which occur in reverse stratigraphic order and are, from deepest to shallowest, 17 681 ± 184 (SUERC-63582), 19 381 ± 174 (SUERC-67931), 18 390 ± 157 (SUERC-67930), 19 328 ± 186 (SUERC-67358) and 29 375 ± 284 (SUERC-63577) cal ka BP (Fig. 8).

Lithofacies 2. Stratified, matrix-supported diamicton (Dms)

This facies is only observed from 246 to 173 cm depth in core 198VC (Figs. 9 and 10). Stratification is inclined and is imparted by textural banding in which zones of diamicton are interbedded with more silty horizons. The stratification is particularly well developed in the lowermost 30 cm, becoming more diffuse gradationally above. Individual beds are internally massive. The lower contact of this facies is gradational although shear strength shows an abrupt decrease across this transition dropping from 90 kPa at 250 cm depth to 20 kPa at 240 cm. Shear strength through the Dms ranges from 6 to 20 kPa. The upper contact is sharp. Two samples from the close to the base of the Dms were radiocarbon dated to provide age constraint on the timing of formation (Fig. 9; Table 2).
A sample of mixed benthic foraminifera from 243 cm depth gave a non-finite age (SUERC-63563). A second sample from 244 cm depth of monospecific *Elphidium clavatum* dated to 25 556 ± 237 (UClAMS-176368) cal a BP.

**Lithofacies 3. Laminated mud, consolidated (Fld(c))**

This facies is identified in three cores, 197VC, 198VC and 199VC. It is a dark grey (5Y4/1) to olive grey (5Y4/2) laminated clayey silt that contains dispersed granules and occasional pebbles as well as abundant, dispersed shell and coral fragments. Laminations vary from inclined to horizontal and have variably sharp or diffuse upper and lower contacts. In core 197VC the laminations become distorted towards the top of this unit and contain a deformed, 9-cm-thick, grey (5YR5/1) silt bed at 159–168 cm (Fig. 9). The shear strength of this facies ranges from 33 to 112 kPa, with an average of 80 kPa. Both the magnetic susceptibility and the wet bulk density are also relatively high and are 99 SI and 2.02 g ml⁻¹, respectively (Supplementary Information Figure b). Seven radiocarbon ages were obtained from this lithofacies (Fig. 9; Table 2). The youngest, a monospecific *Elphidium clavatum* sample collected from 152 cm in core 199VC, yielded an age of 22 655 ± 221 cal a BP (UClAMS-176369). A second monospecific *E. clavatum* sample from 248 cm depth in 199VC produced an age of 25 886 ± 221 cal a BP (UClAMS-186911) while a shell fragment from 275 cm in the same core was indistinguishable from background. Two shell fragments at 150 and 185 cm in core 197VC yielded ages of 31 621 ± 347 (SUERC-60173) and 34 029 ± 270 cal a BP (SUERC-63583), respectively. A further two shell fragments from core 198VC produced ages that were indistinguishable from background.

**Lithofacies 4. Massive mud with clasts, consolidated (Fmd(c))**

This facies occurs in cores 198VC and 199VC where it is the basal facies in both cores (Fig. 9). It consists of a dark grey to olive-grey (5YR4/1–5YR4/2), massive, silty clay with dispersed, predominately granule-sized clasts. Individual beds are 15–40 cm thick. In 198VC the shear strength of this facies ranges from 112 to 200 kPa. In 199VC the basal Fmd(c) has a shear strength of 60–65 kPa and a 20-cm-thick bed higher up in the core from 100 to 80 cm depth yielded 45–50 kPa. The average wet bulk density is 2.05 g ml⁻¹, similar to facies Fmd (see below), but the magnetic susceptibility is much higher, averaging 103.7 SI and reaching a maximum 134 SI, probably as a consequence of greater abundance of larger clasts within the muddy matrix (Supplementary Information Figure b). Facies Fmd(c) also contains abundant shell fragments and is particularly rich in foraminifera. In both cores Fmd(c) grades subtly into facies Fld(c). Three samples date the same horizon in Fmd(c) at a depth of 312–314 cm, close the bottom of core 199VC (Fig. 9; Table 2). These comprise ages of 37 229 ± 753 (SUERC-63564), 40 452 ± 631 (SUERC-67942) and 31 915 ± 492 (SUERC-67941) cal a BP from a mixed benthic foraminifera sample, a monospecific *Bulimina elongata* sample and a monospecific *Elphidium clavatum* sample, respectively. A further monospecific *Elphidium clavatum* sample from the base of core 198VC produced an age indistinguishable from background (UClAMS-186908).

**Lithofacies 5. Laminated mud (Fld)**

Facies Fld is identified in core 197VC from the Slyne Trough (Fig. 9). It is a dark grey (5Y4/1) laminated, clayey silt that contains dispersed granules and occasional pebbles some of which are draped by overlying laminae. Shell and coral fragments are abundant in this facies. Bioturbation, in the form of burrows, is occasionally present. Lamination style is variable, ranging from sharp and well-defined millimetre to sub-millimetre in thickness to more diffuse. Shear strength of the Fld facies is typically < 25 kPa with a minimum of 9 kPa. Magnetic susceptibility is 1.9 SI and wet bulk density is 92 g ml⁻¹ (Supplementary Information Figure b). There are no radiocarbon measurements from this lithofacies.

**Lithofacies 6. Massive mud with clasts (Fmd)**

Facies Fmd occurs in two cores from the Slyne Trough: 173–55 cm depth in core 198VC and 70–43 cm depth in core 199VC. This facies consists of massive, silty clay with dispersed, predominately granule-sized clasts. Individual beds are 15–40 cm thick. A sample and a monospecific *Elphidium clavatum* foraminifera sample, a monospecific *Elphidium clavatum* sample collected from 152 cm in core 199VC, yielded an age of 22 655 ± 221 cal a BP (UClAMS-176369). A second monospecific *E. clavatum* sample from 248 cm depth in 199VC produced an age of 25 886 ± 221 cal a BP (UClAMS-186911) while a shell fragment from 275 cm in the same core was indistinguishable from background. Two shell fragments at 150 and 185 cm in core 197VC yielded ages of 31 621 ± 347 (SUERC-60173) and 34 029 ± 270 cal a BP (SUERC-63583), respectively. A further two shell fragments from core 198VC produced ages that were indistinguishable from background.

**Table 1. Location, water depth and recovery of sediment cores discussed in text.**

| Core name (IC106-) | Location | Water depth (m) | Recovery (m) |
|------------------|----------|----------------|--------------|
| 158VC            | 53°48.117′N, 13°7.35′W | 295            | 2.28         |
| 159VC            | 53°47.441′N, 12°59.977′W | 316            | 0.76         |
| 161VC            | 53°44.876′N, 12°59.442′W | 267            | 0.51         |
| 167VC            | 53°39.376′N, 12°46.022′W | 257            | 0.98         |
| 168VC            | 53°39.72′N, 12°42.543′W | 246            | 0.67         |
| 169VC            | 53°41.457′N, 12°41.799′W | 277            | 1.02         |
| 170VC            | 53°41.457′N, 12°41.799′W | 267            | 0.38         |
| 172VC            | 53°47.555′N, 12°46.875′W | 271            | 1.35         |
| 176VC            | 53°38.855′N, 12°17.386′W | 292            | 2.08         |
| 177VC            | 53°36.373′N, 12°4.896′W | 277            | 2.34         |
| 180 PC           | 53°18.322′N, 10°12.677′W | 112            | 6.49         |
| 184 PC           | 53°20.049′N, 10°16.551′W | 100            | 7.90         |
| 190VC            | 53°21.818′N, 11°9.173′W | 149            | 1.97         |
| 191VC            | 53°25.008′N, 11°17.092′W | 146            | 2.58         |
| 194VC            | 53°33.31′N, 11°42.048′W | 240            | 5.24         |
| 195VC            | 53°34.237′N, 11°43.13′W | 236            | 4.18         |
| 196VC            | 53°34.414′N, 11°43.295′W | 235            | 2.38         |
| 197VC            | 53°45.064′N, 11°53.655′W | 278            | 2.44         |
| 198VC            | 53°49.389′N, 11°50.428′W | 290            | 3.85         |
| 199VC            | 53°36.677′N, 12°4.23′W | 302            | 3.29         |
| 211VC            | 53°1.556′N, 11°43.657′W | 160            | 1.94         |
| Publication code | Core | Depth (cm) | Sample type | \(^{14}C\) age (a yr) ± 1σ | Calibrated age (cal a yr) ± 2σ \(\Delta R = 0\) years | Calibrated age (cal a yr) ± 2σ \(\Delta R = 300\) years | Calibrated age (cal a yr) ± 2σ \(\Delta R = 700\) years | Bayesian modelled age (ka) | Comment |
|-----------------|------|------------|-------------|-----------------------------|-------------------------------------------------|-------------------------------------------------|-------------------------------------------------|---------------------------------|-----------------------------|
| SUERC-67929     | JC106-158VC | 160       | Shell fragment | Indistinguishable from background | 16 414 ± 46 | 18 328 ± 186 | 18 949 ± 129 | 18 586 ± 135 | Iceberg turbate – maximum age |
| SUERC-63578     | JC106-158VC | 179       | Shell fragment | | 15 333 ± 43 | 18 390 ± 157 | 18 081 ± 161 | 17 588 ± 178 | Iceberg turbate – maximum age |
| SUERC-67930     | JC106-158VC | 184–185   | Shell fragment | | 16 460 ± 44 | 18 381 ± 174 | 18 999 ± 139 | 18 633 ± 122 | Iceberg turbate – maximum age |
| SUERC-67931     | JC106-158VC | 184–185   | Shell fragment | | 14 906 ± 43 | 17 681 ± 184 | 17 294 ± 190 | 16 695 ± 237 | Iceberg turbate – maximum age |
| SUERC-63582     | JC106-158VC | 188–188.5 | Shell fragment | | 25 722 ± 84 | 29 375 ± 284 | 29 053 ± 290 | 28 651 ± 221 | Iceberg turbate – maximum age |
| SUERC-63577     | JC106-158VC | 58        | Shell fragment | | 25 722 ± 84 | 29 375 ± 284 | 29 053 ± 290 | 28 651 ± 221 | Iceberg turbate – maximum age |
| SUERC-58387     | JC106-159VC | 68        | Shell fragment | Indistinguishable from background | 21 202 ± 57 | 25 103 ± 265 | 24 669 ± 302 | 24 161 ± 208 | Till – maximum age |
| SUERC-58389     | JC106-167VC | 48–49     | Shell fragment | | 21 452 ± 60 | 25 398 ± 206 | 25 023 ± 308 | 24 431 ± 237 | Till – maximum age |
| SUERC-67932     | JC106-167VC | 50–51     | Shell fragment | | 23 109 ± 66 | 27 263 ± 192 | 26 932 ± 312 | 26 413 ± 263 | Till – maximum age |
| SUERC-58392     | JC106-167VC | 61–62     | Shell fragment | | 47 431 ± 928 | 53 308 ± 358 | 52 308 ± 358 | 51 308 ± 358 | Till – maximum age |
| SUERC-67933     | JC106-167VC | 74        | Shell fragment | | 21 146 ± 59 | 25 014 ± 310 | 24 581 ± 298 | 24 104 ± 208 | Till – maximum age |
| SUERC-67937     | JC106-167VC | 80–81     | Shell fragment | | 26 895 ± 89 | 30 786 ± 206 | 30 533 ± 298 | 29 997 ± 404 | Till – maximum age |
| SUERC-58393     | JC106-167VC | 81–82     | Shell fragment | | 69 332 ± 117 | 81 180 ± 388 | 79 629 ± 496 | 78 064 ± 525 | Till – maximum age |
| SUERC-59511     | JC106-176VC | 69        | Coral | | 29 328 ± 117 | 33 140 ± 388 | 32 691 ± 496 | 32 064 ± 525 | Till – maximum age |
| SUERC-59512     | JC106-176VC | 70.5      | Shell fragment | Indistinguishable from background | 24 541 ± 79 | 28 168 ± 282 | 27 876 ± 190 | 27 618 ± 167 | Till – maximum age |
| SUERC-59518     | JC106-177VC | 125       | Shell fragment | Indistinguishable from background | 24 529 ± 82 | 28 155 ± 284 | 27 868 ± 190 | 27 611 ± 169 | Till – maximum age |
| SUERC-59519     | JC106-177VC | 132       | Shell fragment | Indistinguishable from background | | | | | Till – maximum age |
| SUERC-59520     | JC106-177VC | 158.5     | Shell fragment | Indistinguishable from background | | | | | Till – maximum age |
| SUERC-60159     | JC106-177VC | 164–165   | Shell fragment | | 24 688 ± 171 | 36 171 ± 390 | 35 885 ± 402 | 35 465 ± 447 | Till – maximum age |
| SUERC-59521     | JC106-177VC | 168.5     | Shell fragment | | 24 365 ± 80 | 27 986 ± 241 | 27 754 ± 161 | 27 515 ± 174 | Till – maximum age |
| SUERC-60160     | JC106-177VC | 196       | Shell fragment | | 24 365 ± 80 | 27 986 ± 241 | 27 754 ± 161 | 27 515 ± 174 | Till – maximum age |
| SUERC-60163     | JC106-177VC | 214       | Shell fragment | Indistinguishable from background | | | | | Till – maximum age |
### Table 2. Continued

| Publication code | Core | Depth (cm) | Sample type | 14C age (aBP) ±1σ | Code Core Depth (cm) Sample type | 14C age (aBP) ±1σ | Comment |
|------------------|-----|------------|--------------|--------------------|---------------------------------|--------------------|---------|
| SUERC-60167-24   | 190C | 176 | Shell fragment | 28 867 ± 1834 | SUERC-60167-24 | 190C | 176 | Shell fragment | 28 867 ± 1834 | Consolidated glacimarine mud | 197VC, 2020 |
| SUERC-60167-25   | 190C | 154 | Shell fragment | 20 695 ± 53  | SUERC-5832 | 190C | 176 | Shell fragment | 29 214 ± 125 | Maximum age | 197VC, 2020 |
| SUERC-60167-26   | 190C | 150 | Shell fragment | 19 387 ± 58  | SUERC-60167-26 | 190C | 150 | Shell fragment | 11 265 ± 45  | Maximum age | 197VC, 2020 |
| SUERC-60167-27   | 190C | 125 | Shell fragment | 20 849 ± 231 | SUERC-60167-27 | 190C | 125 | Shell fragment | 23 312 ± 117 | Maximum age | 197VC, 2020 |
| SUERC-60167-28   | 190C | 100 | Shell fragment | 26 446 ± 204  | SUERC-60167-28 | 190C | 100 | Shell fragment | 32 407 ± 56  | Maximum age | 197VC, 2020 |
| SUERC-60167-29   | 190C | 85 | Shell fragment | 26 134 ± 206 | SUERC-60167-29 | 190C | 85 | Shell fragment | 31 982 ± 53  | Maximum age | 197VC, 2020 |
| SUERC-60167-30   | 190C | 70 | Shell fragment | 25 029 ± 359 | SUERC-60167-30 | 190C | 70 | Shell fragment | 25 029 ± 359 | Maximum age | 197VC, 2020 |
| SUERC-60167-31   | 190C | 55 | Shell fragment | 25 414 ± 241 | SUERC-60167-31 | 190C | 55 | Shell fragment | 25 414 ± 241 | Maximum age | 197VC, 2020 |
| SUERC-60167-32   | 190C | 40 | Shell fragment | 19 470 ± 100 | SUERC-60167-32 | 190C | 40 | Shell fragment | 19 470 ± 100 | Maximum age | 197VC, 2020 |
| SUERC-60167-33   | 190C | 25 | Shell fragment | 19 216 ± 71 | SUERC-60167-33 | 190C | 25 | Shell fragment | 19 216 ± 71 | Maximum age | 197VC, 2020 |
| SUERC-60167-34   | 190C | 10 | Shell fragment | 20 695 ± 53 | SUERC-5832 | 190C | 10 | Shell fragment | 20 695 ± 53 | Maximum age | 197VC, 2020 |
| SUERC-60167-35   | 190C | 5 | Shell fragment | 20 000 ± 13 | SUERC-5832 | 190C | 5 | Shell fragment | 20 000 ± 13 | Maximum age | 197VC, 2020 |
| SUERC-60167-36   | 190C | 0 | Shell fragment | 20 000 ± 13 | SUERC-5832 | 190C | 0 | Shell fragment | 20 000 ± 13 | Maximum age | 197VC, 2020 |

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| Publication code | Core | Depth (cm) | Sample type | \(^{14}C\) age (a BP)± 1σ | Calibrated age (cal a BP)± 2σ ΔR = 0 years | Calibrated age (cal a BP)± 2σ ΔR = 300 years | Calibrated age (cal a BP)± 2σ ΔR = 700 years | Bayesian modelled age (ka) | Comment |
|-----------------|------|------------|-------------|---------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------|---------|
| SUERC-60177     | IC106-197VC | 227–228  | Shell fragment | 24 438 ± 83                 | 28 061 ± 269                    | 27 802 ± 169                    | 27 559 ± 172                    | 27.88 ± 0.15 | Till – maximum age |
| SUERC-58395     | IC106-197VC | 230  | Shoe | 30 244 ± 126              | 33 944 ± 247                     | 33 738 ± 243                     | 33 369 ± 363                     | 33.63 ± 0.21 | Till – maximum age |
| SUERC-60178     | IC106-197VC | 230  | Shoe | 23 393 ± 76              | 27 322 ± 192                     | 27 038 ± 314                     | 26 532 ± 325                     | 27.18 ± 0.18 | Till – maximum age |
| UCIAMS-176385   | IC106-198VC | 77–80   | Foraminifera, monospecific *Elphidium clavatum* | 18 410 ± 60                        | 21 807 ± 210                      | 21 429 ± 253                      | 20 870 ± 216                      | 21.81 ± 0.11 | Glacimarine mud – ice sheet retreat |
| SUERC-63563     | IC106-198VC | 243–245  | Mixed benthic foraminifera | 45 223 ± 723                        |                              |                              |                              |                              | Glacimarine debris-flow – ice sheet retreat |
| UCIAMS-176368   | IC106-198VC | 244–246  | Foraminifera, monospecific *Elphidium excavatum* | 21 620 ± 90                          | 25 556 ± 237                      | 25 256 ± 282                      | 24 706 ± 355                      | 25.61 ± 0.11 | Glacimarine debris-flow – ice sheet retreat |
| SUERC-59522     | IC106-198VC | 327  | Shell fragment | Indistinguishable from background | 19 630 ± 70                          | 23 171 ± 270                      | 22 783 ± 239                      | 22 403 ± 157 | Glacimarine mud – ice sheet retreat |
| SUERC-59523     | IC106-198VC | 340  | Shell fragment | Indistinguishable from background | 19 200 ± 70                          | 22 655 ± 221                      | 22 374 ± 171                      | 21 936 ± 241 | Consolidated glacimarine mud – retreat and readvance |
| UCIAMS-186908   | IC106-198VC | 378–382 | Foraminifera, monospecific *E. excavatum* | Indistinguishable from background | 28 829 ± 115                          | 32 349 ± 539                      | 31 916 ± 500                      | 31 449 ± 247 | Consolidated glacimarine mud – retreat and readvance |
| UCIAMS-176386   | IC106-199VC | 55–57   | Foraminifera, monospecific *E. clavatum* | 21 990 ± 130                          | 25 866 ± 221                      | 25 616 ± 280                      | 25 192 ± 404                      | 25.79 ± 0.12 | Consolidated glacimarine mud – retreat and readvance |
| UCIAMS-176369   | IC106-199VC | 152–154  | Foraminifera, monospecific *E. clavatum* | 28 531 ± 110                          | 31 915 ± 492                      | 31 533 ± 282                      | 31 264 ± 188                      | 31.51 ± 0.2 | Consolidated glacimarine mud – retreat and readvance |
| SUERC-58396     | IC106-199VC | 275  | Shell fragment | Indistinguishable from background | 36 212 ± 259                          | 40 452 ± 631                      | 40 120 ± 637                      | 39 657 ± 651 | Consolidated glacimarine mud – retreat and readvance |
| SUERC-67941     | IC106-199VC | 312–314  | Foraminifera, monospecific *Bulimina eebonga* | 33 480 ± 182                          | 37 229 ± 753                      | 36 775 ± 633                      | 36 272 ± 427                      | 36.06 ± 0.14 | Consolidated glacimarine mud – retreat and readvance |
| SUERC-67942     | IC106-199VC | 312–314  | Foraminifera, monospecific *E. clavatum* | 36 212 ± 259                          | 40 452 ± 631                      | 40 120 ± 637                      | 39 657 ± 651                      | 39.65 ± 0.24 | Consolidated glacimarine mud – retreat and readvance |
| SUERC-63564     | IC106-199VC | 312–314  | Mixed benthic foraminifera | 33 480 ± 182                          | 37 229 ± 753                      | 36 775 ± 633                      | 36 272 ± 427                      | 36.06 ± 0.14 | Consolidated glacimarine mud – retreat and readvance |

(Continued)
| Publication code | Sample type | 14C age (aBP) ± 2σΔR | Calibrated age (cal a BP) ± 2σΔR | Comment |
|-----------------|-------------|----------------------|--------------------------------|---------|
| SUERC-60158     | Shell fragment | 22 211 ± 176         | 20 592 ± 176                  | Iceberg turbate – maximum age (Callard et al., 2020) |
| SUERC-60179     | Shell fragment | 24 627 ± 45          | 24 710 ± 45                   | Iceberg turbate – maximum age (Callard et al., 2020) |
| Beta-334419     | Coral fragment | 180 118 ± 90         | 108 ± 90                      | Till – maximum age (Peters et al., 2015, 2016) |
| Poz2-66430      | Single bivalve shell | 94 733 ± 167        | 108 ± 90                      | Till – maximum age (Peters et al., 2015, 2016) |

199VC (Fig. 9). It is a massive, well-sorted, dark grey (5Y4/1) silt or silty clay with shell and coral fragments. Occasional sandy pods, stringers and black mottles occur, and may represent bioturbation. Bounding contacts are sharp to gradational. The shear strength of this lithofacies ranges from 8 to 35 kPa and averages 15 kPa. The wet bulk density is low, averaging 1.97 g ml⁻¹. The magnetic susceptibility is also low, averaging 88 SI (Supplementary Information Figure b). Two monospecific samples of benthic foraminifera Elphidium clavatum were collected from close to the top of this facies in both 198VC (77 cm) and 199VC (54 cm). They dated to 21 807 ± 210 (UCIAMS-176385) and 23 171 ± 270 (UCIAMS-176386) cal a BP, respectively (Fig. 9; Table 2).

### Lithofacies 7. Clast-supported sandy gravel (Gs)

This lithofacies comprises a clast-supported medium to coarse sandy gravel that contains angular to rounded clasts up to 4 cm in diameter as well as abundant shell and coral fragments (Figs. 8 and 9). Individual beds range from 5 to 65 cm in thickness. The upper contact is either sharp or gradational. Where the contact is gradational, facies Gs fines into graded or massive sand (Sg and Sm facies; see below). Facies Gs is unconsolidated but it has a high average wet bulk density of 2.30 g ml⁻¹ and a very high average magnetic susceptibility of 329 SI due to its gravelly nature (Supplementary Information Figures a and b).

### Lithofacies 8. Gravelly sand (Sg)

This lithofacies is a massive, unconsolidated, gravelly, medium to coarse sand that contains a high abundance of shell and coral fragments (Fig. 8). It is invariably overlain by massive sand (Sm). Clasts range from gravel to pebble-sized, reaching a maximum of 4 cm, and range from angular to rounded. The sand is olive (5Y4/3) or very dark grey (5Y3/1). Individual beds range from 3 to 50 cm in thickness. The upper contact can be sharp or gradational. Where gradational, Sg fines into facies Sm. This lithofacies is unconsolidated and has a high average wet bulk density and magnetic susceptibility of 217 g ml⁻¹ and 296 SI, respectively (Supplementary Information Figure a).

### Lithofacies 9. Massive sand (Sm)

The uppermost lithofacies in all the cores consists of a saturated, massive, fine to medium sand that contains frequent shell fragments and occasional pebble and gravel-sized clasts (Figs. 8 and 9). The colour varies from olive (5Y4/3), olive grey (5Y4/2) or dark grey (5Y4/1) and often becomes darker downcore. The upper and lower contacts can be either sharp or gradational. The massive sand is underlain predominantly by gravelly sand (Sg). Individual beds are typically a few tens of centimetres thick. Average wet bulk density and magnetic susceptibility is 1.98 g ml⁻¹ and 118.3 SI, respectively, but both values vary between cores and within cores, with high values often correlating with the presence of isolated pebble-sized clasts (Supplementary Information Figures a and b).

### Lithofacies associations and interpretations

The nine individual lithofacies identified in cores from the Porcupine Bank are grouped into four lithofacies associations (LFAs) (Figs. 8 and 9). LFA 1: consolidated diamicton association (facies Dmm(c); LFA2: consolidated mud association (facies Fmd(c), Fmd(c)); LFA3: muddy diamicton association (facies Dms, Fld, Fmd); LFA4: sand-gravel association (facies Gs, Sm, sg).
LFA 1. Diamicton association (Dmm(c))

Massive matrix-supported diamicton is often the lowermost lithofacies in the cores. It is typically stiff with shear strength values up to 200 kPa. Stratification and grading were not observed. This lithofacies is interpreted as a subglacial till with the high shear strengths consistent with loading and compaction by grounded glacier ice (Evans et al., 2006; Ó Cofaigh et al., 2007, 2013; Evans, 2018). Shell fragments within the diamicton indicate that pre-existing marine sediments were incorporated into the till during glacier overriding, and radiocarbon ages on these reworked shells provide a

| Sample code | Location | Outcrop type | Sample lithology | Bayesian modelled age (ka) |
|-------------|----------|--------------|------------------|---------------------------|
| T5BH01      | Black Head | Erratic | Granite | 13.8 ± 1.1 |
| T5BH02      | Black Head | Erratic | Granite | 16.99 ± 0.31 |
| T5BH03      | Black Head | Erratic | Granite | 11.8 ± 1.0 |
| T5CL01      | Claddaghuff | Erratic | Granite | 16.98 ± 0.29 |
| T5CL02      | Claddaghuff | Erratic | Granite | 17.0 ± 0.31 |
| T5CL03      | Claddaghuff | Erratic | Granite | 17.0 ± 0.30 |
| T5CL04      | Claddaghuff | Erratic | Granite | 17.0 ± 0.30 |
| T5CL06      | Claddaghuff | Erratic | Granite | 17.0 ± 0.33 |
| T5CL07      | Claddaghuff | Erratic | Granite | 16.99 ± 0.31 |
| T5IE01      | Illion east | Erratic | Metasandstone | 15.53 ± 0.81 |
| T5IE03      | Illion east | Erratic | Quartzite | 15.81 ± 0.71 |
| T5IM01      | Inis Meain | Erratic | Granite | 17.0 ± 1.2 |
| T5IM02      | Inis Meain | Erratic | Granite | 19.09 ± 0.108 |
| T5IM03      | Inis Meain | Erratic | Metasandstone | 22.3 ± 1.2 |
| T5KK01      | Kilkierian | Erratic | Granite | 16.99 ± 0.3 |
| T5KK02      | Kilkierian | Bedrock | Granite | 17.0 ± 0.3 |
| T5KK03      | Kilkierian | Bedrock | Granite | 17.0 ± 0.3 |
| T5KK04      | Gowan East | Erratic | Granite | 16.56 ± 0.3 |
| T5KK05      | Gowan East | Erratic | Granite | 16.54 ± 0.29 |
| T5KK06      | Gowan East | Erratic | Granite | 16.57 ± 0.28 |
| TSMOY01     | Moycullen | Erratic | Granite | 16.51 ± 0.32 |
| TSMOY02     | Moycullen | Erratic | Granite | 16.52 ± 0.32 |
| TSMOY03     | Moycullen | Erratic | Granite | 16.54 ± 0.3 |
| TSMOY04     | Moycullen | Erratic | Granite | 16.56 ± 0.28 |
| TSMOY05     | Moycullen | Erratic | Granite | 16.57 ± 0.28 |
| TSOU04      | Rossaveel | Erratic | Granite | 16.99 ± 0.3 |
| TSOU05      | Rossaveel | Erratic | Granite | 13.5 ± 1.3 |
| TSOU06      | Rossaveel | Bedrock | Granite | 17.03 ± 0.34 |
maximum age for till formation and thus ice sheet advance. The youngest ages are the most instructive in this regard and indicate till formation and the presence of a grounded ice sheet on Porcupine Bank after ~25 cal ka BP. This is also consistent with work by Peters et al. (2016) who obtained an age of 24.1 cal ka BP from a reworked shell in till from the bank. Collectively this indicates the presence of a grounded ice sheet on the Porcupine Bank during the LGM and until ~24 cal ka BP.

The matrix-supported diamicton in core 158VC from the north-western Porcupine Bank is also massive and very stiff (average shear strength of 102 kPa and maximum of 125 kPa). This might be consistent with an origin as a subglacial till deposited by grounded ice, but we consider this less likely for the following reasons. First, radiocarbon ages on reworked shells within the diamicton are in the age window of 17.8–19.3 cal ka BP, Callard et al. (2020) have recently shown that deglaciation of Porcupine Bank was largely complete by about 23 cal ka BP with the ice sheet grounded on the mid-shelf by this time. This is considerably earlier than the ages on reworked shells in core 158VC and is also consistent with ages of ~22–23 cal ka BP from core site 199VC on the eastern Porcupine Bank. Hence the continued presence of a grounded ice sheet on the outer Porcupine Bank as late as 17.8–19.3 cal ka BP is considered unlikely. Second, seafloor bathymetric data from the site of 158VC show that the surface of the GZW from which this core was collected is covered in an extensive series of elongate furrows consistent with iceberg scours (Fig. 5D). We therefore suggest that the massive, matrix-supported muddy diamicton in 158VC is probably a product of formation in a glacimarine environment in which grounded iceberg keels intermittently scoured the seafloor and consolidated the sediment (cf. Dowdeswell et al., 1994; Sacchetti et al., 2012). The date of 18.4 cal ka BP on the articulated bivalve Yoldiella species from close to the top of the diamicton in core 177VC is also anomalously young. It may represent incorporation of younger material into the top through localized reworking, e.g. by mass flow or bottom current activity.

**LFA2. Consolidated mud association (facies Fld(c), Fmd(c))**

This lithofacies association comprises laminated and massive muds with clasts. In many respects these sediments are similar to LFA3 (notably facies Fld and Fmd) but are characterized by higher shear strengths of up to 200 kPa. Furthermore, in some cores (e.g. 197VC) facies Fld(c) laminations are contorted and/or interbedded with deformed silts. Laminated and massive muds with clasts are characteristic of deposition by a range of subaqueous processes in glacimarine and glaciolacustrine environments by suspension settling through the water column, iceberg rafting and sediment gravity flows (cf. Powell, 2003). A glacimarine environment is consistent with the presence of abundant *E. clavatum*, a benthic foraminifer that tolerates low-salinity, cold water environments. However, the highly consolidated nature of the matrix of facies Fld(c) and Fmd(c) is difficult to reconcile with in situ glacimarine sediments. Rather, it points to a two-stage depositional process in which initial subaqueous deposition of laminated and massive glacimarine muds was followed by ice-marginal overriding and glaciectonic compaction (cf. Ó Cofaigh et al., 2011). Hence the sediments of LFA2 are regarded here as glacitectonites (sensu Benn and Evans, 1996; Evans et al., 2006; Evans, 2018). This implies a period of ice-free conditions before ice-sheet readvance and overriding (cf. Peters et al., 2015, 2016).

Radiocarbon measurements from LFA2 were obtained on a mix of samples, comprising individual shell fragments, monospecific samples of benthic foraminifera and mixed benthic foraminiferal samples. All of these ages provide maximum age estimates for the overriding and glaciectonism. The youngest, a monospecific *E. clavatum* sample from core 199VC dated at 22 655 ± 221 cal a BP indicates the core site was ice-free at this time but was subsequently overridden. A second sample of monospecific *E. clavatum* from 248 cm depth in the core dated at 25 866 ± 221 cal a BP also indicates ice-free conditions and a glacimarine sedimentary environment at this time and before overriding (Fig. 9; Table 2). However, most of the samples are considerably older. Indeed, monospecific samples of *E. clavatum* from the base or close to the base of cores 198VC and 199VC gave a non-define age and an age of 31 915 ± 492 cal a BP, respectively (Fig. 9; Table 2). The latter age on *E. clavatum* was obtained on a sample from 312 to 314 cm depth where we also dated a mixed benthic foraminiferal sample (37 229 ± 753 cal a BP) and a monospecific sample of the warm water species *Bulimina elongata* (40 452 ± 631 cal a BP). This was done to assess the impact of a mix of warm and cold species on the age of the mixed sample. The monospecific *E. clavatum* age is the youngest of the three and implies the presence of glacimarine conditions on the bank at ~32 cal ka BP, before ice advance and overriding.

**LFA3. Mud-diamicton association (facies Dms, Fld, Fmd)**

Collectively this LFA is interpreted as recording deglacial glacimarine sedimentation associated with retreat of grounded ice on Porcupine Bank. Stratified, matrix-supported diamicton is interpreted as the product of subaqueous cohesive debris flow deposition based on the inclined stratification, textural banding and interbedding of poorly sorted diamictons with more silty horizons (Postma, 1986; Eyles and Eyles, 2000; Mulder and Alexander, 2001; Talling et al., 2012). The change in facies upcore from consolidated laminated muds (Fld(c)) interpreted as glacitectonites (see above) into stratified debris flow diamictons with low shear strengths (Dms) in core 198VC is consistent with a change in depositional environment from...
subglacial to deglacial associated with the retreat of grounded ice. A monospecific sample of *E. clavatum* from 2 cm above this contact is dated at 25 556 ± 237 cal a BP (Fig. 9) and provides the best constraint on the timing of final ice-sheet retreat from the outer Porcupine Bank.

Laminated mud with dispersed clasts (facies Fld) is inferred to be a product of meltwater sedimentation in which mud emplacement was by suspension settling from turbid meltwater plumes issuing from the retreating grounding-line during deglaciation and supplemented by fine-grained turbidity currents (cf. Mackiewicz *et al.*, 1984; Hesse *et al.*, 1997; Ó Cofaigh and Dowdeswell, 2001; Lucchi *et al.*, 2013). The presence of clasts draping by overlying laminae supports an interpretation of suspension settling through the water-column of fine-grained sediment interspersed with the episodic delivery of coarser ice-rafted debris (Thomas and Connell, 1985).

Massive mud with clasts (facies Fmd) is also interpreted as a product of rapid suspension settling of fine-grained material from meltwater plumes combined with iceberg rafting (Domack, 1984; Dowdeswell *et al.*, 1994). The presence of well-preserved *E. clavatum* indicates that this occurred in a glacimarine setting. The production of such massive glacimarine muds may reflect changes in the position of the grounding-line with the massive structure signifying a more distal setting. Alternatively, it could signify a more proximal location where rapid sedimentation acted to suppress laminae formation. Radiocarbon ages from monospecific samples of *E. clavatum* (see above) indicate the presence of glacimarine conditions, and thus recession of grounded ice, by ~23 cal ka BP.

LFA4. Sand-gravel association (facies Gs, Sm, Sg)

These facies form the uppermost LFA in the sediment cores (Figs. 8 and 9). They comprise a series of variably sorted sand and gravelly facies that are massive and contain abundant shell and coral fragments. Such facies have been commonly described from the uppermost part of the Quaternary sediment sequence on the Atlantic shelf along the Irish margin (Peters *et al.*, 2015, 2016; Callard *et al.*, 2018; Ó Cofaigh *et al.*, 2019) and we interpret them similarly as postglacial in origin, formed predominantly by bottom current activity on the shelf (cf. Howe *et al.*, 2001). Dating constraints on this LFA, however, are poor. The lower boundary of LFA4 is typically erosional and the closest ages to it are from the top of the underlying muds of LFA2 and LFA3. These provide a range of ages of ~21–23 cal ka BP on monospecific samples of benthic foraminifera (*E. clavatum*). We therefore infer the presence of a hiatus due to bottom current erosion associated with formation of LFA4. We do not consider LFA4 further in this paper.

**Advance and retreat of the last BIIS across Porcupine Bank**

GZWs and moraine ridges combined with dated subglacial tills in sediment cores from across the northern Porcupine Bank indicate an extensive BIIS, which reached the outer shelf during the glGM. Radiocarbon dates on reworked shells within subglacial tills and glaciectonites provided a range of ages from non-finite to finite. Ages on these shells constrain the timing of open-marine conditions on the shelf, as the marine fauna must have originally lived in open water. However, they also provide a maximum age for the enclosing subglacial deposits and therefore ice sheet advance during which the marine fauna and their host sediments were cannibalized and reworked subglacially. Such glacially reworked shell popula-

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Glacitectonized deglacial subaqueous facies are characteristic of many areas of the last BIS where it underwent retreat in a marginal glacimarine or glaciacustrine environment (e.g. Hambrey et al., 2001; Ó Cofaigh and Evans, 2001; Evans and Ó Cofaigh, 2003; Roberts and Hart, 2005; Hiemstra et al., 2006; Chiverrell et al., 2018; Roberts et al., 2018; Ó Cofaigh et al., 2011, 2012b, 2019; Callard et al., 2018, 2020).

Grounding-line retreat occurred in a glacimarine environment. Peters et al. (2016) first proposed that this was an ice shelf rather than a tidewater margin. The presence of GZWs within deeper water, notably within the Slynne Trough, suggests that vertical accommodation space in front of the grounding-line was limited and is consistent with formation in an ice-shelf cavity (Dowdeswell and Fogelli, 2012). However, the sharp-crested moraines in shallower water across Porcupine Bank (Fig. 6) are more indicative of a grounded tidewater ice margin (see Batchelor and Dowdeswell, 2015 and references therein). We suggest that the nature and distribution of the deglacial landforms across the Porcupine Bank and Slynne Trough record local variability in accommodation space along the BIS grounding-line during retreat, as reflected in lateral transitions from a grounded tidewater front to an ice shelf. In terms of lithofacies, sediments deposited at ice-shelf grounding lines are characterized by coarse-grained facies, often represented as stratified diamictics which transition upwards into laminated and massive muds (e.g. Kilfeather et al., 2011; Smith et al., 2019). Core 198VC from the outer Slynne Trough preserves a sequence of subglacial sediments overlain by glacially stratified diamict and massive mud. It is therefore possible that this reflects sedimentation with increasing distance from the grounding-line in an ice shelf cavity. However, it is equally plausible that this sequence and the massive-laminated glacimarine muds found in other cores from the Slynne Trough represent tidewater glacimarine sediments produced by ice-marginal debris flows, suspension-settling and rain-out of ice-rafter debris (Mackiewicz et al., 1984; Powell and Domack, 1995). Thus, although there is geomorphological support for ice-shelf development within the Slynne Trough during deglaciation, the sedimentological data are more equivocal.

Bayesian analysis and palaeoglaciological reconstruction

We now use a Bayesian temporal model to reconstruct the pattern of ice sheet retreat from the outer shelf offshore of western Ireland to a few tens of kilometres inshore of the present coastline (Figs. 11 and 12). For full details of the Bayesian model see the Methods section. We divide our reconstruction into three geographical areas: (1) Porcupine Bank and the Slynne Trough, (2) the mid- to inner shelf, including the Aran Islands, and (3) inland from the present coastline. We thus capture the ‘marine to terrestrial transition’.

Porcupine Bank and the Slynne Trough

Across the northern Porcupine Bank, GZWs and moraines, combined with subglacial tills and glacitectonites, provide evidence for a grounded ice sheet that extended to the outer shelf during the glGM. The Bayesian analysis indicates that the maximum position was attained by 26.8 ± 1.33 ka BP (Boundary Layer 1, Fig. 12), and that retreat was underway by 25.9 ± 1.8 ka BP (Boundary Layer 2, Fig. 12). The presence of well-preserved glacimarine benthic foraminifera in deglacial lithofacies implies that retreat took place in a glacimarine environment, and that glacimarine conditions prevailed across the bank until 25.4 ± 0.25 ka BP (Boundary Layer 3, Fig. 12). Ages on reworked shells from tills and overridden glacimarine sediments on the outer shelf indicate that the ice sheet readvanced onto the bank, and probably oscillated there, until c. 24.3 ± 0.19 ka BP (Boundary Layer 4, Fig. 12) when it underwent retreat towards the mid-shelf (Fig. 12). The geomorphology, sedimentology and radiocarbon chronology are therefore consistent with an extensive ice sheet that was grounded on the outer shelf at the glGM (cf. Peters et al., 2015, 2016), which commenced retreat also during the glGM, but then oscillated during deglaciation. Such early BIS retreat is consistent with recent work from further to the north in Donegal Bay (Ó Cofaigh et al., 2019) and the Malin Sea shelf (Callard et al., 2018), and also from the Celtic Sea sector to the south (Scourse et al., 2019).

Mid-inner shelf

Sub-bottom profiler data from across the mid- to inner shelf reveal a series of well-developed GZWs and intervening smaller moraines indicating episodic retreat across the shelf (Callard et al., 2020). The largest of these GZWs has been referred to variously as the ‘Mid-Shelf Grounding-Zone Complex’ (Callard et al., 2020) or ‘Galway Lobe Grounding-Zone Wedge’ (Peters et al., 2016). It is a composite, multi-crested feature containing over-ridden glacimarine muds and formed at an oscillating grounding line which glacitectonized deglacial subaqueous sediments. Across the mid-shelf, individual GZWs are separated by sedimentary basin fills, and cores from these basins recovered overridden glacimarine sediments. Callard et al. (2020) interpreted the basin fills as having formed by a two-stage depositional process in which initial glacimarine sedimentation took place in front of the grounding line when the ice sheet was positioned at a GZW, followed by grounding-line readvance(s) which overrode and deformed the deglacial sediments (cf. Peters et al., 2016).

The BIS grounding-line had retreated to the Mid-Shelf Grounding-Zone Complex and was oscillating in this position by 23.7 ± 0.45 ka BP (Boundary Layer 5, Fig. 12). Hence the Porcupine Bank and Slynne Trough were ice-free by this time. Retreat from the Mid-Shelf Grounding-Zone Complex back to the Aran Islands on the inner shelf had occurred by c. 22.0 ± 1.12 ka BP (Boundary Layer 6, Fig. 12). Grounding-line retreat at this time is consistent with glacimarine sedimentation in the Slynne Trough dated to ~22.8 cal ka BP (Callard et al., 2020). A minimum age for the timing of deglaciation of the inner shelf is provided by an age of 17.1 cal ka BP in glacimarine sediments (Callard et al., 2020). However, the Bayesian modelling of TCN ages obtained for glacially transported, erratic boulders of Galway granite resting on Carboniferous limestone bedrock on Inish Mean, the middle of the three Aran Islands (see Roberts et al., 2020), suggests that these sites were ice-free by 19.5 ± 0.8 ka BP (Fig. 12; Table 3).

Terrestrial transition

TCN ages primarily obtained from glacially transported, erratic boulders constrain the timing of ice sheet retreat onto mainland Ireland and then further inland into the mountains of Connemara (Roberts et al., 2020) (Table 3). These ages in the Bayesian analysis indicate that the ice sheet had retreated from the inner shelf onto mainland Ireland by 17.3 ± 0.46 ka BP (Boundary Layer 7, Fig. 12) and imply a relatively slow retreat from the Aran Islands (deglaciated by 19.5 ± 0.8 ka BP) back onto the coastline. The Aran Islands may have acted as a pinning point on the inner shelf, stabilizing the BIS grounding-line and thus slowing overall ice sheet retreat.
TCN ages on erratic boulders along the Connemara coastline or from a few kilometres inland constrain the timing of ice sheet retreat inland from the coast (see Roberts et al., 2020). This phase of retreat was underway by 16.8 ± 0.2 ka BP (Boundary Layer 8, Fig. 12). At this time the BIIS grounding-line had stepped back from the inner shelf and the ice sheet was largely terrestrially based, although it is possible that, locally, parts of the ice sheet were marine-terminating and still calving icebergs onto the inner shelf. TCN ages on glacial erratics from sites further inland (Roberts et al., 2020) constrain the timing of subsequent north-eastward ice sheet recession inland to the foothills of the mountains. The Bayesian analysis...

TCN ages on erratic boulders along the Connemara coastline or from a few kilometres inland constrain the timing of ice sheet retreat inland from the coast (see Roberts et al., 2020). This phase of retreat was underway by 16.8 ± 0.2 ka BP (Boundary Layer 8, Fig. 12). At this time the BIS grounding-line had stepped back from the inner shelf and the ice sheet was largely terrestrially based, although it is possible that, locally, parts of the ice sheet were marine-terminating and still calving icebergs onto the inner shelf. TCN ages on glacial erratics from sites further inland (Roberts et al., 2020) constrain the timing of subsequent north-eastward ice sheet recession inland to the foothills of the mountains. The Bayesian analysis...
places this phase of retreat at 16.3 ± 0.42 ka BP (Boundary Layer 9, Fig. 12) and the ice sheet had receded back to the foothills by 15.0 ± 1.35 ka BP (Boundary Layer 10, Fig. 12). We infer that at this time the ice sheet was predominantly terrestrially based, and this is supported by reconstructions of RSL in this area, which indicate that RSL remained below present between 20 and 10 ka BP (Edwards and Craven, 2017; Roberts et al., 2020). Interestingly, however, Callard et al. (2020) suggested that the ice sheet may locally have had a marine-terminating margin on the shelf as late as ~15.6 ka BP based on dated evidence of iceberg turberation on the mid-shelf slope. This is also consistent with Peters et al. (2016) who proposed that their ‘Connemara Lobe’, an outlet glacier emanating from Killary Harbour (Fig. 12), overrode marine sediments dating to c. 15.1 cal ka BP.

Discussion
Extent of the BIIS on the Atlantic shelf west of Ireland during the LGM

Early terrestrially-based models of the extent of the last BIIS in western Ireland can be divided into two broad hypotheses (Ballantyne et al., 2008; Ballantyne and Ó Cofaigh, 2017): (1) at the LGM the ice sheet limit was located onshore at a moraine belt termed the Ballycastle–Mulrany limit leaving ice-free terrain beyond this limit, both north and south of Clew Bay (e.g. Charlesworth, 1928; Orme, 1967; Synge, 1968, 1979; Bowen et al., 2002; Knight et al., 2004); and (2) at the LGM the ice sheet extended offshore onto the adjoining continental shelf (e.g. Hallissy, 1914; Warren, 1991, 1992). As noted by Ballantyne et al. (2008), however, both hypotheses contain uncertainties. These mainly centred on, firstly, the age of the ‘unglaciated’ terrain beyond the LGM limit as to whether this was ‘Munsterian’ (Knight et al., 2004) or Early Midlandian in age (Bowen et al., 2002; McCabe et al., 2007), and, secondly, whether the mountains of Connemara represented a barrier to ice flowing west from the Irish Midlands or acted as an independent centre of ice dispersal.

Ballantyne et al. (2007, 2008) used trimline evidence and TCN dating to constrain the altitude and deglaciation chronology of the last ice sheet in western Ireland. Although the trimlines were initially interpreted as marking the upper limit of the ice sheet, subsequent research showed that they represent an englacial transition between warm-based erosive ice on lower ground and cold-based ice over higher ground and summits (Ballantyne et al., 2011; Ballantyne and...
Stone, 2015), implying that the trimlines record the minimum, rather than maximum, altitude of the LGM ice surface. It is likely that the last ice sheet overtopped all the Connemara mountains, which appear to have acted as an independent centre of ice dispersal throughout the lifetime of the last ice sheet, diverting ice from the Irish Midlands both north-west into Clew Bay and south-west into Galway Bay (Ballantyne and Ó Cofaigh, 2017).

The earliest direct evidence supporting the presence of grounded ice on the Atlantic shelf west of Ireland was reported by King et al. (1998) who used reflection seismic data to document a series of north–south-oriented moraines. They proposed that the outermost moraine positioned at the shelf edge, although not dated, represented the LGM limit (Sejrup et al., 2005). Subsequent glacial geomorphological mapping from the Malin Sea and offshore of Donegal Bay (Fig. 1) using the Olex bathymetric database and high-resolution multibeam swath bathymetry data provided compelling evidence for the presence of grounded ice in the form of drumlins and nested arcuate moraines (Benetti et al., 2010; Dunlop et al., 2010, 2011; Ó Cofaigh et al., 2012a). A shelf-edge-terminating ice sheet during the LGM offshore of northwest Ireland and western Scotland was subsequently confirmed by the evidence provided by dated sediment cores, which indicated that the ice sheet was grounded at the Malin Shelf edge at ~26.5–26.2 cal ka BP (Callard et al., 2018; Ó Cofaigh et al., 2019). Farther south on the shelf offshore of Galway Bay, 14C dates obtained for marine fauna in sediment cores from the Porcupine Bank by Peters et al. (2015, 2016) suggested that grounded ice advanced onto the bank after ~24.1 cal ka BP and did not commence retreat until ~21.8 cal ka BP.

The data presented here support these earlier findings and conclusively demonstrate BIS advances to the outer Atlantic shelf offshore of western Ireland early in the gLGM (Fig. 12). This finding is consistent with a step-wise increase after 26.5 ka BP in the flux of ice-rafted debris to core site MD01-2461, located on the north-western flank of the Porcupine Seabight (Figs. 1 and 2) (Peck et al., 2007). A consistent picture therefore emerges of an extensive BIS offshore of west and north-west Ireland during the gLGM (Ó Cofaigh et al., 2012a, 2019; Peters et al., 2015, 2016; Callard et al., 2018, 2020; this study). This revises previous interpretations that the LGM ice sheet was not as extensive as that of Marine Isotope Stage 3 (MIS3) (cf. Bowen et al., 2002; McCabe et al., 2007), or terminated at the Mid-Shelf Grounding-Zone Complex (cf. Clark et al., 2012; McCarron et al., 2018). The evidence presented here suggests that the LGM advance was probably at least as extensive as any preceding earlier advance(s) across the western shelf.

The Bayesian analysis indicates that ice sheet retreat from outer Porcupine Bank was underway by 25.9 ka BP (Fig. 12). This indicates a relatively short period (<1000 years) of ice sheet grounding at its maximum position on the outer shelf before retreat commenced. Thus, the maximum position may not have been a steady-state ice sheet configuration. However, retreat from Porcupine Bank was not a continuous process. It was punctuated by a series of readvances of up to a few kilometres in extent, as indicated by dated, overridden glaciomarine sediments and tills, as well as GZWs and moraines. These provide compelling evidence for an ice sheet that was intermittently grounded on Porcupine Bank as late as c. 24.3 ka BP based on the Bayesian analysis (cf. Peters et al., 2016; Callard et al., 2020; this study). This implies that although the ice sheet underwent early initial retreat and was characterized by oscillatory behaviour during deglaciation, it maintained an outer shelf position for c. 1500 years. However, deglacial sediments dated to 25.5 ka BP in core 198VC are not subsequently overridden; hence, any such readvances/oscillations did not reach the shelf edge.

**Timing and controls on ice sheet retreat**

Retreat of the last BIS from the outer shelf west of Ireland was underway by 25.9 ka BP and thus early during the gLGM (Fig. 12). Early deglaciation of the Porcupine Bank is consistent with recent studies from the continental shelf offshore of northwest Ireland and Scotland, where retreat was underway by 25–26 ka BP (Callard et al., 2018; Ó Cofaigh et al., 2019). It is also consistent with the retreat of the Irish Sea Ice Stream from the shelf edge of the Celtic Sea at c. 25 ka BP (Scourse et al., 2019). Initial deglaciation of the BIS from the Atlantic shelf west of Ireland and in the Celtic Sea was therefore largely synchronous. Retreat west of Ireland commenced during Greenland Stadial 3 (GS-3, 27.5–23.3 ka; Rasmussen et al., 2014), and this period also encompasses final retreat of the grounding-line off Porcupine Bank and back towards the Mid-Shelf Grounding Zone Complex at c. 24.3 ka BP (Fig. 13C). We therefore rule out atmospheric warming as the trigger for initial deglaciation of the bank.

Recent studies of deglacial benthic foraminiferal populations from the Porcupine Bank and western Irish shelf indicates the persistence of glacialic conditions until c. 20.9 cal ka BP on the outer shelf, and at least c. 18.5 cal ka BP on the mid-shelf (Peters et al., 2020). Several deep-water cores from along the Irish margin, notably core OMEX-2K from the Goban Spur (Scourse et al., 2009; Haapaniemi et al., 2010) and core MD01-2461 from the Porcupine Saddle (Peck et al., 2007), contain high percentages of *N. pachyderma sinistral* during this period, implying low sea-surface temperatures (Fig. 13D). The absence of a clear ocean warming signal suggests that initial retreat was not triggered by ocean forcing.

Rather, we propose that initial deglaciation and retreat across the Porcupine Bank was driven, at least in part, by high RSL. Current water depths in the outer Slyne Trough are 290–340 m. Assuming eustatic sea level was c. 134 m lower than present early in the gLGM (Lambeck et al., 2014), water depths in the Slyne Trough would still have been in the range of c. 156–206 m. Indeed, this would have been a minimum estimate, as it does not take into account the effects of glacioisostatic depression of the outer shelf related to the presence of an extensive grounded Irish Ice Sheet at glacial maximum. Glacioisostatic adjustment modelling indicates water depths of ~203 m in the central Slyne Trough at 25–26 ka BP (Fig. 13E). High RSL is thus a plausible explanation for triggering initial pull-back and retreat across the bank (cf. Eyles and McCabe, 1989; Ó Cofaigh et al., 2019; Scourse et al., 2019). The inference of high RSL and low water temperatures is supported by the presence of benthic glaciomarine foraminifera and associated lithofacies in sediment cores from the Slyne Trough (see ‘Sedimentology and radiocarbon measurements’ above; cf. Peters et al., 2020).

The Bayesian analysis indicates that the gLGM ice-sheet offshore of western Ireland had a duration of <1000 years on the outer shelf. It also implies that retreat from the bank was interrupted by a series of readvances/oscillations as recorded by back-stepping GZWs, moraines and overridden glaciomarine sediments (Figs. 3–9). The ice sheet had retreated to, and was oscillating at, the Mid-Shelf Grounding Zone Complex by 23.7 ka BP (Fig. 12). This feature marks a major grounding-line position as implied by its large size (>20 km wide), composite, multi-crested form and internal structure which contains overridden glaciomarine sediments. Seismic records from the grounding-zone complex indicate that it forms the upper part of over 70 m of Quaternary sediments and that most likely represents formation during multiple glacial cycles (McCarron et al., 2018). Stabilization at this position during the last deglaciation may have been facilitated by the bathymetric
control of a pre-existing glacialic depocentre, and the normal bed slope of the seafloor. A positive feedback may have ensued in which sediment delivery to the grounding-line resulted in a shoaling effect acting to reduce water depths at the grounding line promoting further stabilization (cf. Alley et al., 2007).

Ice-sheet retreat across Porcupine Bank to the Mid-Shelf Grounding Zone Complex (Boundary Layers 1 and 5, respectively, Fig. 12) occurred at a net average rate of ~19 m a⁻¹. However, this estimate does not consider the evidence for oscillations or readvances during deglaciation of the bank. If we estimate the retreat rate from the readvance position at ~25 ka yr⁻¹ (Boundary Layer 3) back to the Mid-Shelf Grounding Zone Complex (Boundary Layer 5) the retreat rate increases to ~30 m a⁻¹. The ice sheet was oscillating at the Mid-Shelf Grounding Zone Complex by 23.7 ka yr⁻¹, and had receded to the Aran Islands by 22.0 ka yr⁻¹ where it apparently underwent a stillstand until at least 19.5 ka yr⁻¹. This implies a retreat rate across the mid-shelf of ~62 m a⁻¹. The duration of the stillstand on Aran probably reflects the physiographic control imparted by the islands, which would have acted to stabilize the grounding-line on the inner shelf. Deglaciation of Aran was complete by ~19.5 ka yr⁻¹ but the ice sheet took ~2 ka to retreat back to the mainland, giving a net average retreat rate of ~8 m a⁻¹. Therefore, retreat from the Mid-Shelf Grounding Zone Complex onto mainland Ireland took place during Greenland Stadial 2.1 (22.9–14.7 ka; Rasmussen et al., 2014) (Fig. 13C), again implying that atmospheric warming was unlikely to have been a significant control on retreat across the mid- and inner shelf.

Collectively, the above figures imply relatively slow, oscillatory retreat across the Porcupine Bank, even allowing for retreat along the reverse bed slope of the outer Slynne Trough, until the ice sheet grounded on the mid-shelf. The net retreat rate doubled across the normal bed slope of the mid-shelf to the Aran Islands where the ice sheet stabilized, and then underwent slow retreat back to the mainland. It is notable these rates are 1–2 orders of magnitude lower than retreat rates for present-day ice streams (e.g. 0.5–2 km a⁻¹; Park et al., 2013; Rignot et al., 2014; Scheuchl et al., 2016). Such slow, episodic and oscillatory retreat is consistent with the geomorphological and sedimentological record of retreat across the Malin Shelf north-west of Ireland, which is characterized by numerous GZWso, moraines and overridden deglacial sediments (cf. Callard et al., 2018; Ó Cofaigh et al., 2012a, 2019). The overall slow retreat is inferred to reflect the control of bathymetry. Shallow water depths on the Porcupine Bank itself would have facilitated grounding-line stabilization and oscillations, as would the normal bed slope across the mid-inner shelf (cf. Peters et al., 2016; Callard et al., 2020). A topographic control on retreat across the mid- and inner shelf may have been augmented by climate cooling during retreat, as it occurred during Greenland Stadial 2.1 (Rasmussen et al., 2014).

Conclusions

- Marine-geophysical data and radiocarbon-dated sediment cores, combined with TCN and OSL dating of onshore ice-marginal landforms, are used to reconstruct the timing and style of ice sheet retreat across the Atlantic shelf west of...
Ireland and the adjoining coastline during the last deglaciation.

- Subglacial tills, moraines and GZW s record an extensive ice sheet at the LGM that terminated on the outermost Porcupine Bank, a westwards extension of the Irish continental shelf.
- Radiocarbon dating on reworked shells in the tills indicate the ice sheet was grounded on the outer shelf at ~26.8 ka BP.
- Grounding-line retreat in a glaciomarine environment was underway by 25.9 ka BP. Retreat was punctuated by stillstands and oscillations/readvances, as recorded by overridden glaciomarine sediments, moraines and GZW s. The grounding-line had retreated inshore of Porcupine Bank by 24.3 ka BP and stabilized on the mid-shelf by 23.7-ka BP, forming a large depocentre, the ‘Mid-Shelf Grounding-Zone Complex’. Overall net retreat rates during this phase were 19–30 m a⁻¹.
- The ice sheet had receded to the Aran Islands by ~22.0 ka BP where it stabilized until ~19.5 ka BP, probably reflecting the topographic control of the islands which would have acted as pinning points. Overall net retreat rates across the mid-shelf were ~62 m a⁻¹.
- The timing of ice-margin recession from the Aran Islands onto mainland Ireland is constrained by numerous TCN ages on ice-transported boulders and glacially eroded bedrock as well as several OSL ages on deglacial outwash. These indicate that retreat across this marine–terrestrial transition took ~2 ka and was slow (8 m a⁻¹).
- Retreat of the BIIS on the Atlantic shelf west of Ireland early during the glGM implies that external forcing by atmospheric warming was not the main driver of initial retreat. Rather deglaciation was probably at least partially triggered by high RSL reflecting a combination of eustatic and glaciostatic components.
- Ice-sheet retreat across the shelf was characterized by numerous stillstands and oscillations of the grounding-line, but it is notable that overall retreat rates were slow. Slow retreat across the mid-inner shelf was probably controlled by a normal bedslope and the substantial pinning point of the Aran Islands. The findings here underline the importance of localized controls, specifically bed topography, on modulating the rate of retreat of marine-based sectors of ice sheets.

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Data availability
All relevant data will be made available in the forthcoming BRITICE-CHRONO online data repository, or upon reasonable request from the lead author.

Supporting information
Additional supporting information can be found in the online version of this article.

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