# Bed erosion during fast ice streaming regulated the retreat dynamics of the Irish Sea Ice Stream.

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# Bed erosion during fast ice streaming regulated the retreat dynamics of the Irish Sea Ice Stream.

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**ABSTRACT**

Marine-terminating ice stream behaviour often defines the stability of ice sheets and is driven by a complex interplay of climatic, oceanic, topographic and glaciological factors. Here, we use new integrated high resolution, extensive (2100 km$^2$) and continuous geophysical, sedimentological and geotechnical data to reconstruct past glacial environments during the Last Glacial Maximum from a well-preserved palaeo-landscape. The data is from the axial centre of the Irish Sea Ice Stream (ISIS), which drained > 17% of the former British-Irish Ice Sheet. Recent geochronological data of the palaeo-ISIS show a build-up and advance of ice to marine-terminating maximum limits in the southern Celtic Sea 27–25 ka BP, followed by rapid ice margin retreat into the northern Irish Sea Basin (ISB) by 20.8 ± 0.7 ka BP. However, the flow dynamics in the central and axial bed of the ISIS through this timeframe are not well understood. Here, we use our new glacial landscape reconstruction to identify the spatial and temporal patterns of flow re-organisation and re-activations for the marine-terminating ISIS. From this we infer how ice streaming was driven by a variety of factors through advance, deglaciation and towards a temporary lift-off of ice from its bed and an ultimate demise. Overprinted subglacial bedforms with differing ice flow directions indicate an on/off behaviour to the ice-streaming, an increasing topographical influence and substantial realignment of ice flows. Subsurface geophysical data reveal the erosive capability of the ice stream through time, with a first erosive component in the formation of mega-scale glacial lineations leaving bedrock exposed at the ice stream bed. The depositional component of MSGL crest building occurred in the same ice-flow phase. Whilst the ice stream was laterally constricted in two locations, likely contributing to changes in ice margin retreat rates, we also propose that changes in basal drag associated with exposed bedrock at the ice–bed interface influenced the retreat dynamics, particularly when this exposure was near the grounding zone. The wider implications of this work are that episodic and highly erosive ice streaming during ice advance and early retreat can change ice–bed conditions radically and in turn influence glacial dynamics during later retreat episode, thus constituting a feedback process to be considered in modelling the dynamics of marine-terminating ice streams.

**KEYWORDS**

Quaternary; Glaciology (incl. palaeo-ice sheets); Europe; Geomorphology, glacial; Geophysics; Subglacial Bedforms; British-Irish Ice Sheet; Palaeo-ice streaming; Deglaciation

## 1. INTRODUCTION

### 1.1. Grounded ice stream dynamics and aim of the study

Deglaciation patterns of major ice sheets are determined by responses to climatic and oceanographic changes (external forcing), topographic and basal thermal conditions (internal forcing), with margin retreat rates varying by as much as an order of magnitude (Stroeven et al., 2016). Ice streams are conduits of grounded ice flowing faster than adjacent areas (Paterson, 1994) and account for most of the ice discharge in Antarctica (Bamber et al., 2000). Realistic ice sheet drainage simulations therefore require knowledge of the onset of, variations in, and shut-down of ice streaming (Jamieson et al., 2012; Schoof and Hewitt, 2016). The former Irish Sea Ice Stream (ISIS) drained >17% of the former British-Irish Ice Sheet (BIIS). During the last deglaciation (25–18 ka BP) the ISIS experienced fluctuations between rapid retreat (in the range 100–550 m $a^{-1}$), stabilisation and short-lived re-advances of the ice margins (Chiverrell et al., 2018; Praeg et al., 2015; Scourse et al., 2019; Small et al., 2018; Smedley et al., 2017a; Smedley et al., 2017b). Whilst the ice margin chronology of ISIS is constrained well, no high-resolution shallow sub-seafloor information of the central part of the ice flow conduit, the Irish Sea Basin (ISB – see Fig. 1) has thus far been integrated to understand consecutive flow phases. We have only small snapshots of one preserved surface of the former ice bed to inform us of the ice dynamical processes associated with the ice margin fluctuations. On the seafloor NW of Anglesey, grounded ice streaming is evidenced, and numerous iceberg scours show that retreat of the ISIS was in part accompanied by a calving ice margin (Van Landeghem et al., 2009). New and extensive data are provided here for both the surface and subsurface of the central sector of the ISB (see Fig. 1), making it an ideal testing ground for exploring subglacial evidence for changes in ice stream dynamics during the well-dated advance, retreat and ultimate demise in the period of 27–18 ka BP.

Grounded marine margins can become unstable influencing ice sheet mass flux and rates of deglaciation, with bed topography, subglacial bed properties and meltwater processes as first order controls (Kleman and Applegate, 2014; Rignot et al., 2014; Bradwell et al., 2019). For the former ISIS a high amplitude tidal regime and rising relative sea-levels (Bradley et al., 2011; Ward et al., 2016) are further factors affecting potentially the stability of the ice margin. Observations of ice streams in the palaeo-domain can provide a longer temporal range and more extensive spatial perspective to that available typically in the observational ice stream record. For example, Lakeman et al. (2018) used geophysical, stratigraphical and chronological data in the western Canadian Arctic to reconstruct a 250 km retreat of a marine-terminating ice stream in ~250 years during the Younger Dryas. Direct comparison of bedforms observed developing at contemporary ice stream beds with equivalent bedforms from palaeo-ice streams, provides basis for interpretation of deglacial processes (Van Landeghem et al., 2009) and surge-stagnation-reactivation cycles (Kurjanski et al., 2019). Observations of how ice–bed interfaces have evolved through time can be discerned from information on (sub-)seabed geophysical surveys (see reviews in Stokes, 2018) and provide potentially important constraint for simulations of glacial dynamics (Hindmarsh, 2018).

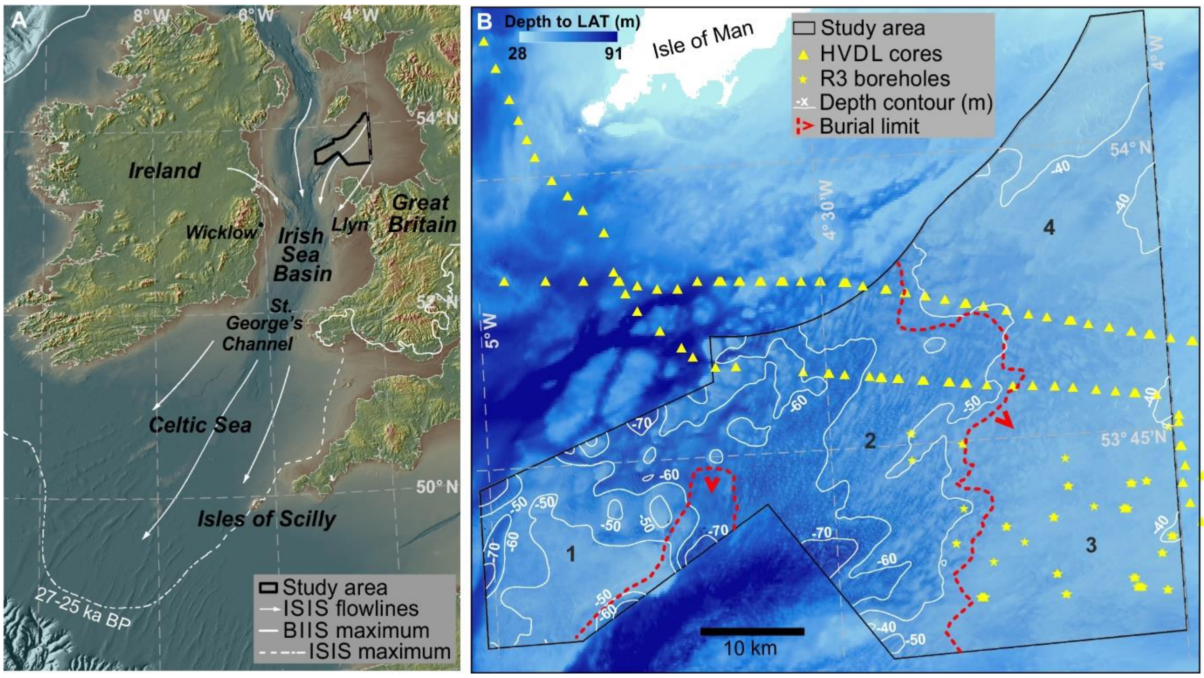



Figure 1. Panel (A) shows the Irish Sea Basin (ISB) with indicative Irish Sea Ice Stream (ISIS) flow lines and the maximum extent of the British-Irish Ice Sheet and ISIS at ca. 27–25 ka BP (Chiverrell et al., 2013; Praeg et al., 2015). Panel (B) shows the study area divided in 4 zones: (Zone 1) a western till plateau, (Zone 2) subglacial bedforms, (Zone 3) buried subglacial bedforms, and (Zone 4) an eastern till plateau. The extent of the geophysical and sedimentological data collected in preparation of a Round 3 (R3) Windfarm development

project is displayed, as are the High Voltage Direct Link (HVDL) cable installation cores. Depth to seafloor is relative to the Lowest Astronomical Tide (LAT).

The aim of this study is to use reconstructed ISIS dynamics as a case study to explain the processes behind changes in ice stream advance, retreat and fluctuations in relation to geological factors of the subglacial bed. We achieve this by reconstructing the temporal and spatial evolution of a palaeo-glacial landscape for the axial central sector of the ice stream (Fig. 1A), contextualised by the palaeo-ISIS geochronology during the Last Glacial Maximum (27–18 ka BP – see next section). A new 2100 km$^2$ seabed geophysical and sedimentological dataset is analysed to capture evidence of changing topography, sediment distribution and glacial bedform assemblages from which we interpret changes in thermal regime, bed topography, bed roughness, drainage patterns, ice flow direction and velocity for the axial sector of the ISIS.

### 1.2. The palaeo-ISIS geochronology

The time series for grounded ice margin dynamics in this area is divided here into five stages. The maximum extension of grounded ice reached the Celtic Sea shelf break around 27–25 ka BP (Stage 1) (Praeg et al., 2015; Scourse et al., 2019; Smedley et al., 2017b). It then had retreated rapidly by 400 km to the south coast of Ireland at 25.1 ± 1.2 ka BP (Stage 2), eventually stabilizing at the constriction of the St-George’s Channel at 24.2 ± 1.2 ka BP (Small et al., 2018). The Irish Ice Sheet may have advanced into the Celtic Sea again (Tótz et al., 2020), but the margin of the Irish Sea Ice Stream retreated rapidly towards the constriction between Wicklow – Llŷn Peninsula at 22.6 ± 1 ka BP, where it stabilised (Stage 3) (Small et al., 2018; Smedley et al., 2017a). The western and deeper (100–140 m below mean sea level (bmsl)) ISB had deglaciated by 19.8 ± 0.8 ka BP (Ballantyne and Ó Cofaigh, 2017; McCabe, 2008), whereas the retreat of ice margins in the adjacent shallower (50–20 m bmsl) eastern ISB was slower crossing and moving north of the Isle of Man from 20.7 ± 0.7 and 19.3 ± 0.8 ka BP onwards (Stage 4) (Chiverrell et al., 2018). Stage 5 (between 19.3 ± 0.8 ka and 18.3 ± 1.1 ka BP) represents an ice-free study area, but a re-advance of Scottish ice on the northern edge of the Isle of Man after 18.3 ± 1.1 ka BP may have seen ice encroach on the area (Chiverrell et al., 2018).

### 1.3. Glacial bedforms in palaeo-environmental reconstructions

The type of glacial bedform can be indicative of cold- or warm-based basal conditions (Kleman and Glasser, 2007; Kleman et al., 2006) whilst bedform geometry can reflect changes in ice flow direction and relative changes in velocity (King et al., 2009; Landvik et al., 2014; Stokes and Clark, 2002). Cold-based ice is below the pressure melting point with no water available at the bed and has a limited ability to transport subglacial sediment and to erode its bed. Thus cold-based ice is mainly associated with preservation of the pre-existing land surface (Cuffey et al., 2000; Paterson, 1994). Warm-based ice is at the pressure melting point with water available at the bed, and therefore associated with more intensive and widespread erosion, deposition and reshaping of the subglacial bed (Kleman and Glasser, 2007; Paterson, 1994). Interpreting ice flow direction and velocity from subglacial bedforms has contributed towards a better understanding of the basal thermal organisation and temporal changes in flow regime for former ice streams (for example: Hogan et al., 2010).

## 2. METHODS

Our reconstruction of changing flow dynamics that characterised the collapsing ISIS stems from new, extensive and high resolution geophysical and geotechnical evidence that document a time and space continuum of an exceptionally preserved glacial landscape. The evidence comprises the surface (multibeam data) and sub-surface (acoustic and geotechnical data) expression of an assemblage of glacial bedforms. Acoustic, geotechnical and sedimentological datasets were obtained for Celtic Array LTD for a proposed Round 3 (R3) offshore wind development (Fig. 1B). From this integration of the geomorphological, sub-bottom acoustic, sedimentological and sediment geotechnical evidence, we have mapped a series of glacial bedform assemblages using ESRI ArcGIS software.

### 2.1. Acoustic surveys

The acoustic data were acquired by the Swedish surveying company MMT (Summer 2010). Bathymetric data were collected with a ship-born Kongsberg EM3002 Multibeam Echosounder (MBES) and processed with CARIS HIPS by MMT. The EM3002 system uses frequencies in the 300 kHz band, ideal for the shallow waters in this area of the Irish Sea (28–92 m). The pulse length emitted by the EM3002 transducer is 150 μs and the depth resolution is 10 mm. Vessel positioning and orientation systems achieved a vertical accuracy < 5 cm and a horizontal accuracy < 0.1 m. The bathymetric data were reduced to Lowest Astronomical Tide (LAT), corrected with the UK Hydrographic Office Vertical Offshore Reference Frame (VORF) model and gridded at 2×2 m. A total of circa 16500 km of sub-bottom seismic data were collected using a 1kJ GeoSpark 200 tip Sparker source and a Chirp Edgetech 512i (0.5–12 kHz) in a dense 2-dimensional grid at line spacings of 150 m and 500 m for cross-lines, all visualised with the KingdomSuite software package. This provided shallow sub-bottom information with a vertical resolution up to 30 cm, from which bedrock horizons and the top of over-consolidated tills were interpreted and digitised initially by MMT. After a thorough review of the data analyses, the geophysical data outputs were analysed further in ArcGIS. A multi-directional hillshading was applied to the bathymetric grids, reducing a potential directional bias in bedform detection. Focal statistics were performed on the irregular bathymetric surfaces to visualise the background break in slope, whereby the mean bathymetric values were calculated in a raster with 500m cell size and contour lines smoothened to eliminate the influence of the slope of smaller bedforms.

### 2.2. Geotechnical and sedimentological surveys

Seabed drilling was performed by Fugro Survey. The seabed drilling frame was positioned using ultra-short baseline (USBL) acoustics with respect to Fugro's Starfix global positioning system. Additional sediment descriptions were performed by Reynolds International Ltd. and Parsons Brinckerhoff delivered a geotechnical interpretative report. These results were integrated with other borehole surveys (Fig. 1B; 2) and the gridded seismic analyses to gain confidence in the stratigraphic interpretations. Normally consolidated sand was generally described as very loose to loose, slightly silty to silty gravelly fine to coarse sand, normally consolidated clay as very soft to stiff slightly sandy slightly gravelly clay. Over-consolidated sand was generally described as dense to very dense slightly silty to silty, gravelly fine to medium sand. Band of gravels and cobbles and closely spaced thin laminae of clay have been recorded in this unit. Over-consolidated clay was generally described as very stiff to very hard slightly sandy to sandy, slightly gravelly clay. Soft bands occur within this unit. Occasional soft bands and sand pockets are recorded within this unit, and a possible boulder. Values for the key geotechnical parameters in Table 1, based on the Recommended Practice for Statistical Representation of Soil Data (DNV, 2007), allowed normally consolidated and over-consolidated sediments to be distinguished from soil profiles, and integrated with seismic sub-bottom profiles. The thickness of the over-consolidated till (Fig. 4; 6; 8) was derived from subtracting Two-Way-Travel Times (TWT) from sound waves reflecting off the top of the bedrock with the TWT from sound waves reflecting off the top of the over-consolidated till (Fig. 6). The speed of sound through the sediments is variable (for example 1700–2700 ms$^{-1}$ measured for clay) and conversion from TWT to meters can only be estimated.

| Parameter | Profile | Normally consolidated clay | Normally consolidated sand | Over-consolidated clay | Over-consolidated sand |
|---|---|---|---|---|---|
| Relative density (%) | characteristic lower bound | | 7 | | 40 |
| | characteristic mean | | 65 | | 85 |
| Undrained shear strength (kPa) | characteristic lower bound | 10 | | 50 | |
| | characteristic mean | 50 | | 300 | |
| CPT tip resistance (MPa) | characteristic lower bound | 0.3 | 2 | 4 | 5 |
| | characteristic mean | 1.2 | 15 | 10 | 30 |

Table 1. Values for the key parameter for the normally and over-consolidated sediments, measured and reported on by Parsons Brinckerhoff.

# 3. OBSERVATIONS AND INTERPRETATIONS

## 3.1 Stratigraphic units underpinning interpretation of the glacial landscape.

The stratigraphy of the sub-surface can be summarised as normally consolidated sediments, over-consolidated sediments and bedrock, and these divisions underpin the interpretations of the glacial landscape. Boreholes (BH) 32a and 33 are representative of this simplified lithostratigraphy. They were recovered along a sub-bottom profile with lateral variation in acoustic properties, and the integration of borehole descriptions and sub-bottom profile is presented in Fig. 2, taking 1800 $ms^{-1}$ as the speed of sound through the sediments to match the down-core depth in meters to two-way travel time in milliseconds. BH32a was extracted exactly on the sub-bottom profile line. The down-core transition between over-consolidated sediments and the underlying mudstone corresponds with the high-amplitude seismic reflector between chaotic acoustic facies with discontinuous internal reflections, and the acoustic facies underneath with continuous internal reflections at a high angle and the upper boundary reflection forming a sharp angular unconformity. BH33 is offset by 100 m from the sub-bottom profile line, yet the transition between the normally consolidated sediments at the top and the over-consolidated sediments underneath coincides with a clear distinction between low-medium amplitude acoustic facies and high-amplitude acoustic facies underneath. The transition to bedrock underneath is offset by circa three meters.

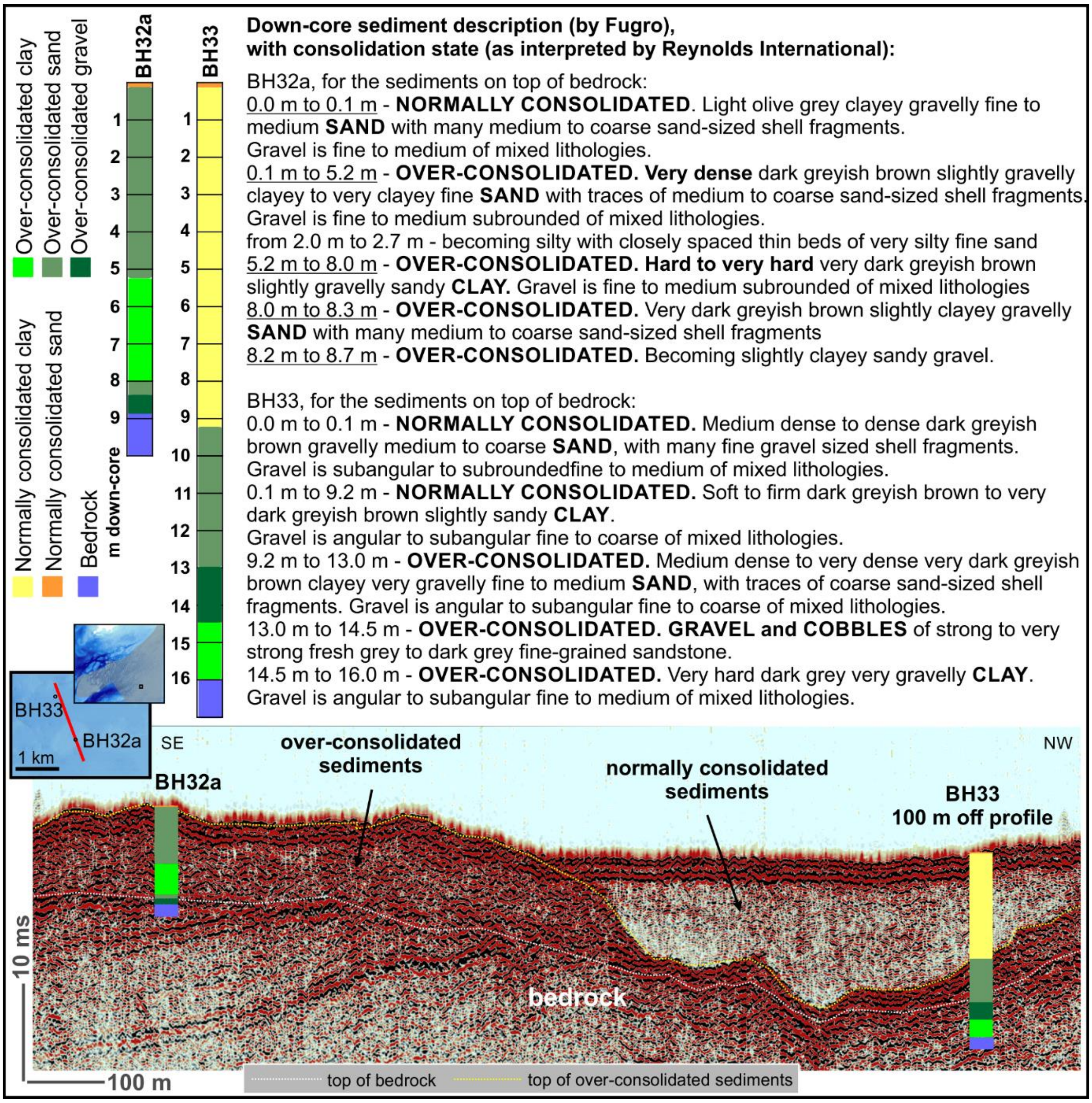



Figure 2. Integration of the sedimentological description of two boreholes with the sub-bottom acoustic profile along which they were extracted.

### 3.2 Mapping the glacial landscape with inferred ice flow directions

The bedform record in this study area is dominated by glacial bedforms. These bedforms were mapped on both the seabed surface and in the sub-surface. With locations of examples presented in Fig. 3., Fig. 4 summarises in three page-wide panels the interpretive lexicon used for the glacial bedforms analysed in this area. From their geomorphological and spatial analyses, it was possible to reconstruct the patterns of ice flow and to infer the changing properties in subglacial bed conditions in the centre of the ISB. To assess wavelengths in between flat topped subglacial bedforms, it was more accurate to delineate the trough in between the bedforms, which had a steeper dip and could be better observed and delineated.

Groupings of glacial bedforms with typically a spatial extent > 50 $km^2$ and a coherence in terms of morphology, proximity and orientation are used here to identify a series of ice-flow sets. To facilitate discussion of the bedform sequence, the glacial landscape is subdivided into four broad morphological regions (Fig. 1B): (Zone 1) a western till plateau, (Zone 2) subglacial bedforms, (Zone 3) buried subglacial bedforms,  and (Zone 4) an eastern till plateau.

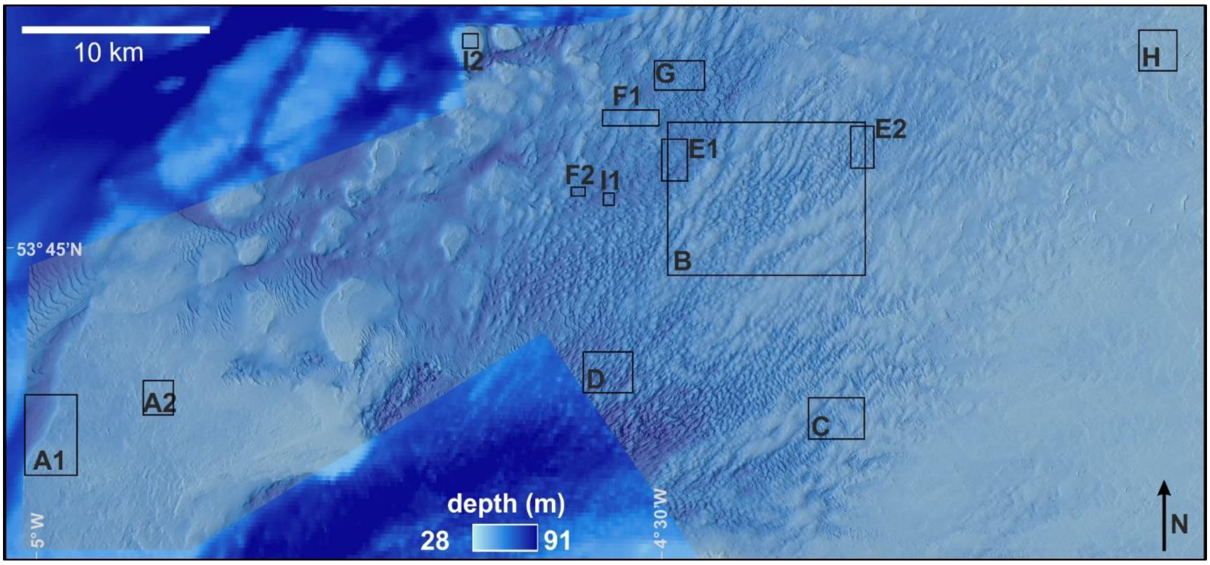
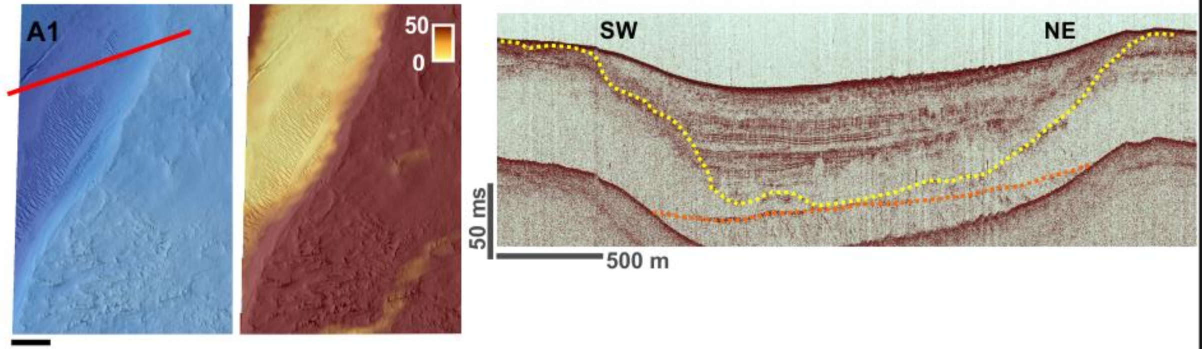
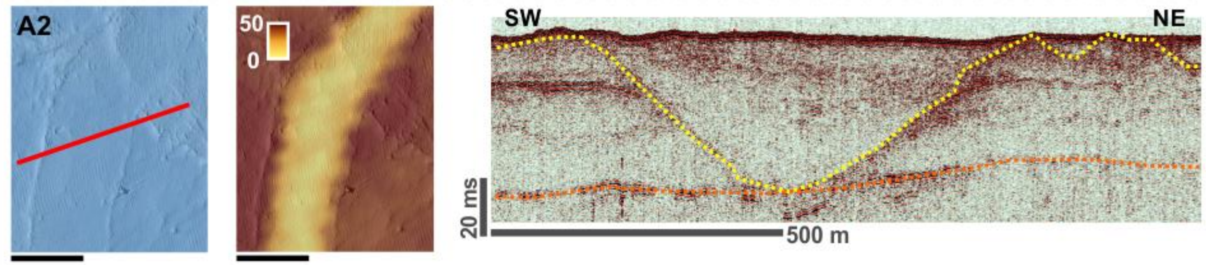
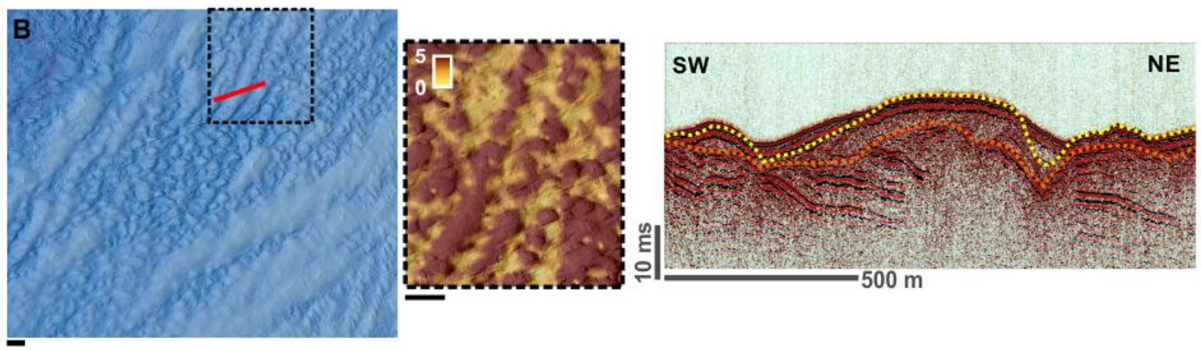
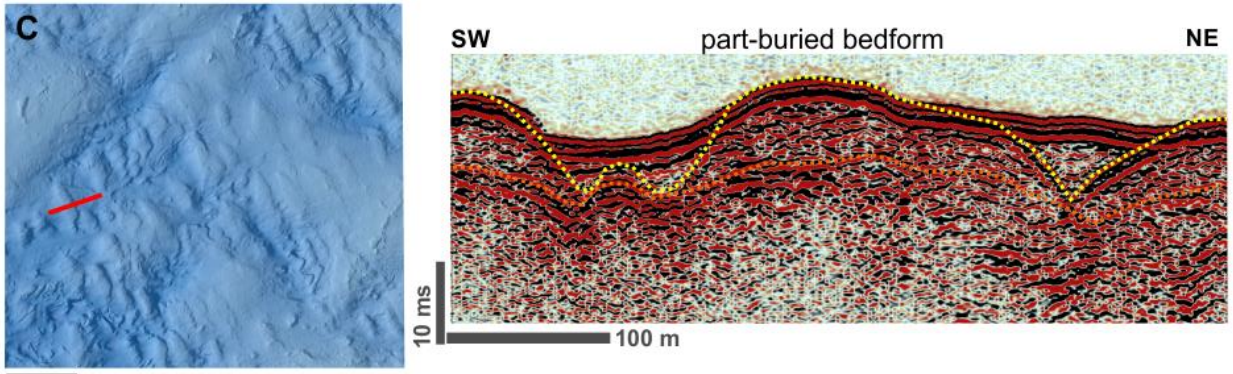
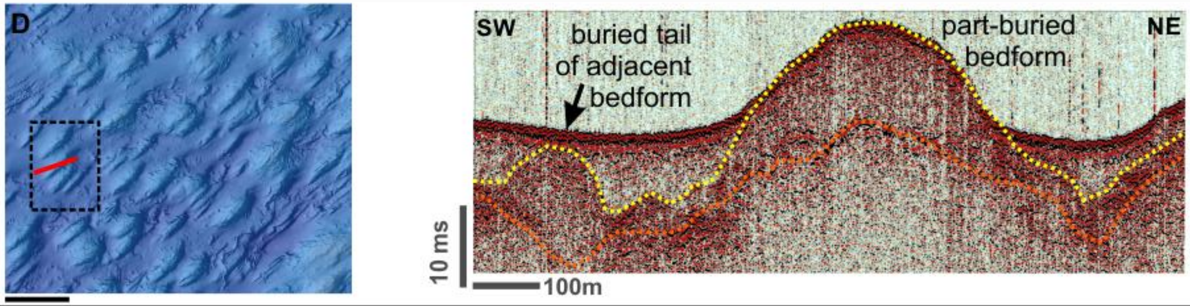
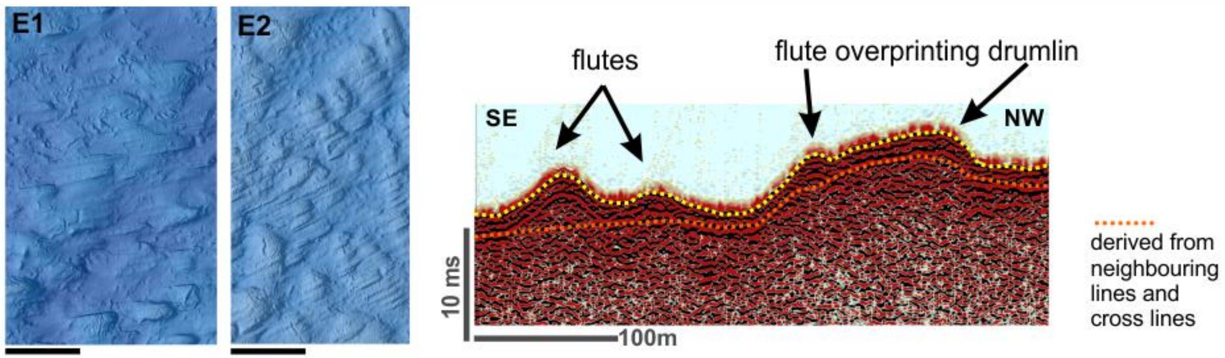



Figure 3: Seabed bathymetry of study area illuminated with multi-directional hillshading, identifying the location of exemplar bedforms presented in Fig. 4, where individual glacial bedforms are shown and analysed.

| N 500m depth (m) 28 91 TWT (ms) — Seabed bathymetry (m), thickness of over-consolidated till (TWT in ms), both with multi-directional hillshade of seabed bathymetry, and sub-bottom profile (SBP), located via ⁄ — top of bedrock, top of over-consolidated sediments | Description | Interpretation Ice flow |
| --- | --- | --- |
|   | Channels cut through over-consolidated sediments, subsequently partly filled with normally consolidated sediments | Subglacial drainage channels. |
|   | Channels cut through over-consolidated sediments, subsequently entirely filled with normally consolidated sediments | Subglacial drainage channels. |
|   | Elongated bedforms from over-consolidated sediments, often partly eroded into bedrock.Thicker sediment deposits on bedform crest (consistent with ploughing process). Length >35 km, fragmented (overprinting by smaller subglacial bedforms). Elongation ratio >60:1, spacing 500–1000 m. | Mega-scale Glacial Lineations (MSGLs) (formed subglacially in direction of ice flow) |

| N 500m depth (m) 28 91 location SBP — Seabed bathymetry (m) with multidirectional hillshade and sub-bottom profile (SBP). top of bedrock; top of over-consolidated sediments | Description | Interpretation Ice flow |
|---|---|---|
|   | Elongated bedforms from over-consolidated sediments, typically 300–1000 m long, 100 m wide and 5–10 m high. Often fit neatly together like a jigsaw. Gentle slopes (rarely >10°) and variable symmetry. Smaller bedforms towards the SE. Bedforms found buried up to 20 m deep. | Ribbed moraines (formed subglacially transverse to direction of ice flow). |
|   | Elongated bedforms from over-consolidated sediments, sometimes partly eroded into bedrock. Tapered side towards deeper ISB. Elongation ratios < 15:1. Length typically 500 m, width 150–250 m, height 10–15 m. Steep eastern slopes (< 30°). | Drumlins (formed subglacially in direction of ice flow). |
|   | Small bedforms from over-consolidated sediments, partly overprinting drumlins (left) and ribbed moraines (right). They are typically 100–150 m long, 20–50 m wide and 1–2 m high. Too small to detect in the subsurface. | Flutes / flutings (formed subglacially in direction of ice flow). |

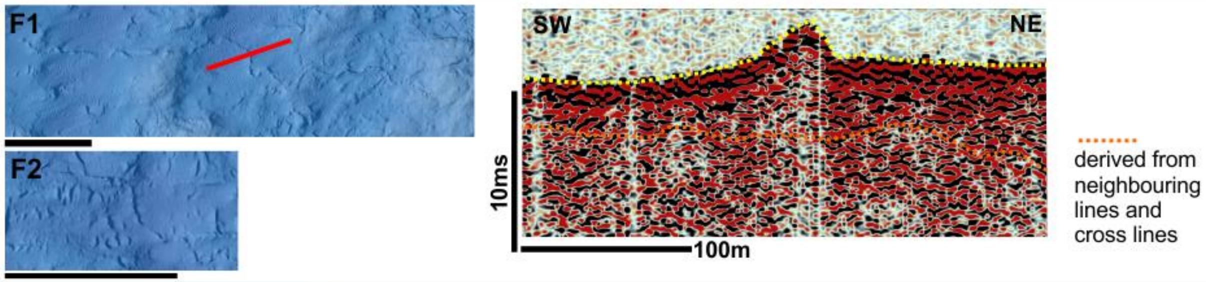
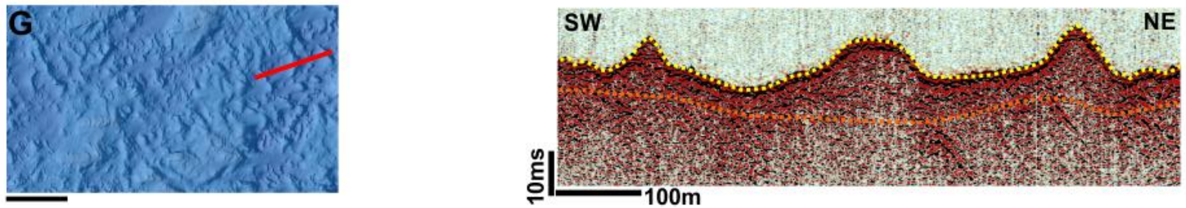
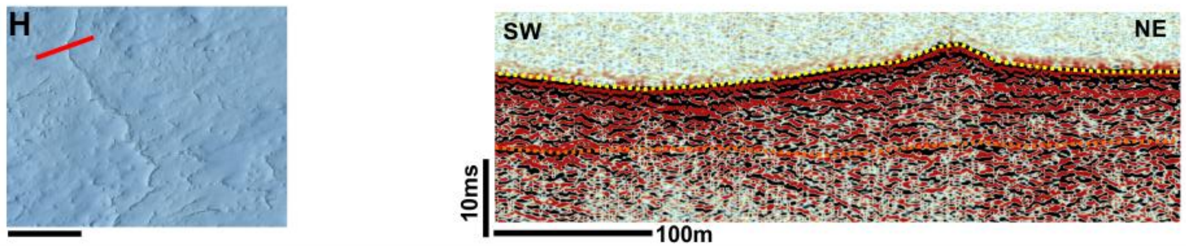
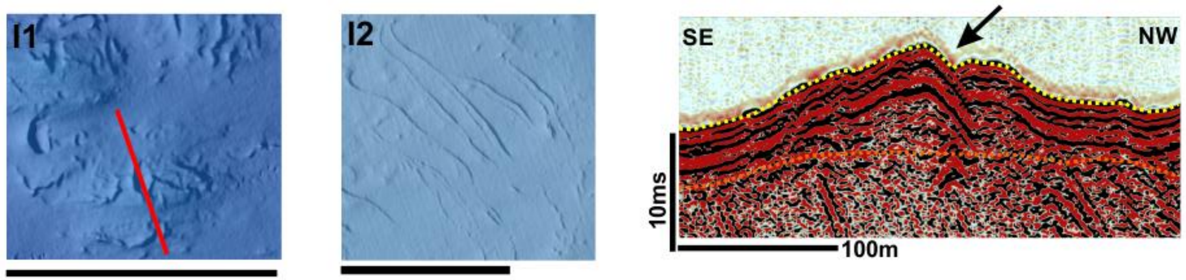
| N<br>500m<br>depth (m) 28 91<br>location SBP<br>**Seabed bathymetry (m) with multidirectional hillshade and sub-bottom profile (SBP).**<br>top of bedrock · top of over-consolidated sediments | **Description** | **Interpretation**<br>*Ice flow* |
|---|---|---|
|  | Long, winding ridge, up to 4 m high. Overprinting flutes, but some parts of the eskers have been streamlined as well. | Eskers (formed subglacially). |
|  | Ridges of diamicton, typically 5 m high, 1km (yet variable) in length, 100-150m wide, variable in asymmetry. Surficial sand ripples over the top of the features alter their shape in places. | Push moraines (formed near/at ice margin). |
|  | Ridges of diamicton, typically 2-3 m high. Fragmented, several kilometres long. Slightly steeper towards the eastern side. | End moraines (formed at ice margin). |
|  | Parallel sets of berms around pit or elongated depression with various shapes & sizes. Round pits can be 100 m in diameter, 3 m deep. Small marks overprinting drumlins (left); long marks on a plateau of over-consolidated sediments (right). | Iceberg scour marks. |

Figure 4. Three panels with examples (A–I) of bedforms identified in the multibeam echosounder data and sub-bottom profiles (SBP), with delineation of bedrock and over-consolidated till surfaces on the acoustic profiles, description and interpretation of the bedforms and the implied direction of ice flow.

Figure 5 presents maps of the axes of these individual bedforms identified and exemplified in Fig. 4. Only the axes of the MSGL could be traced into the subsurface from the 500×150 m grid of sub-bottom profiles. The precise axis of individual smaller and (partly) buried bedforms could not be fully delineated from sub-bottom profiles, but their presence could be detected. The ice-flow sets (FS) interpreted from these bedform assemblages are summarised in Fig. 5J and in Fig. 8 and contextualised with the five stages of the advance and retreat chronology for the ISIS defined in Section 1.2.

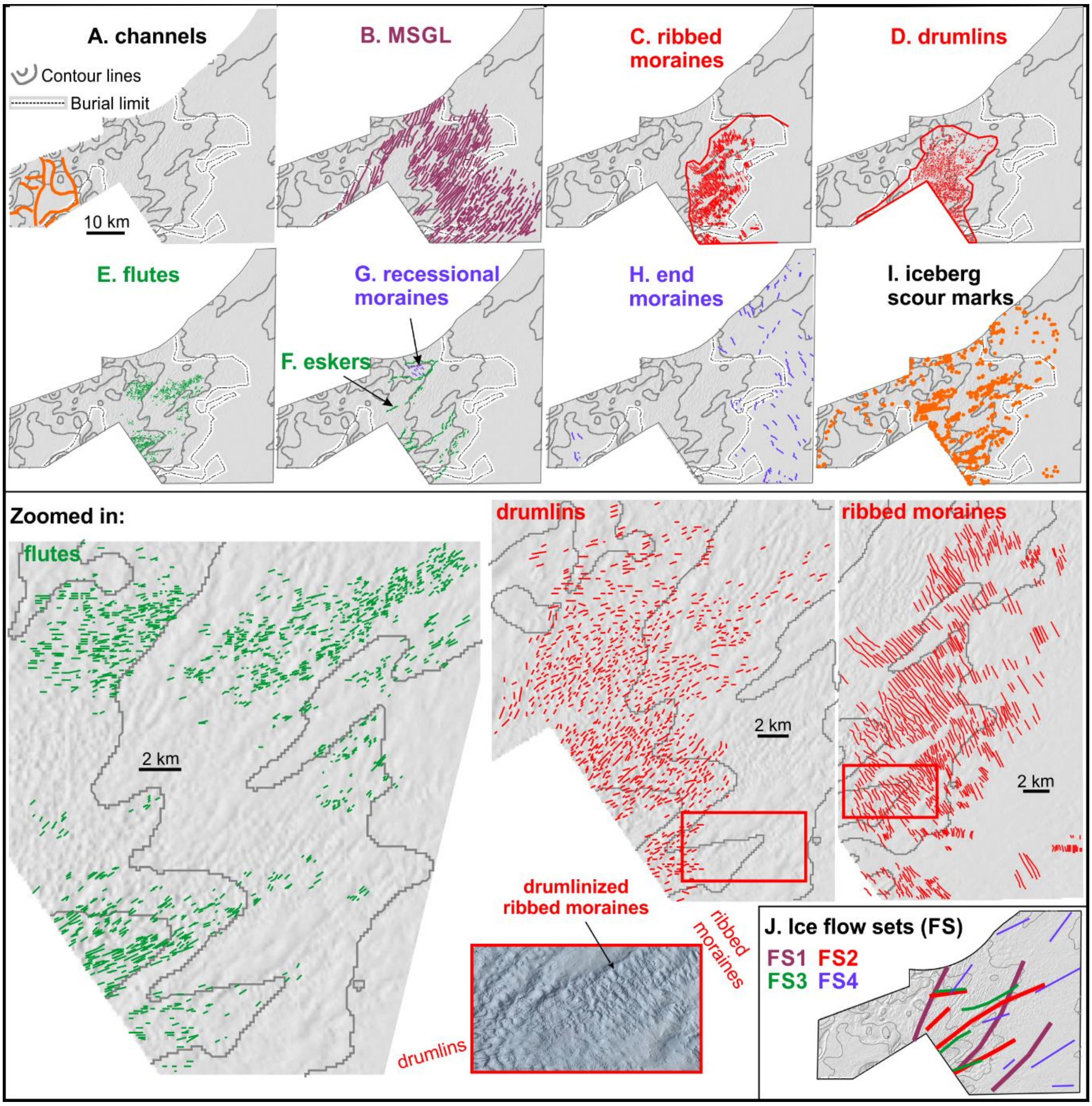



Figure 5. A–I: The axes of the individual bedforms as identified and exemplified in the panels of Fig. 4. Only the axes of the MSGL could be digitised from the sub-bottom profiles, so the axes of the ribbed moraines and drumlins only represent the bedforms present on the seabed surface. The areas with all ribbed moraines, drumlins and flutes have been zoomed in to better discern the direction of the longest axes of the bedforms, and a hillshaded seabed bathymetry displays the drumlinised ribbed moraines. J: schematic representation of ice-flow sets (FS) interpreted from these bedform assemblages and their overprinting relationships (Fig. 4).

### 3.3 Mega-scale glacial lineations (MSGLs) forming and exposing bedrock – ISIS maximum extent and rapid initial retreat

The western till plateau (Zone 1) comprises over-consolidated tills that are incised by linear channels (Fig. 4A) dissecting in places the entire sedimentary sequence with a similar NNE-SSW orientation to the subglacial drainage systems on Anglesey (Lee et al., 2015). The channels have been filled partially or fully by normally consolidated sediments, and some infills post-date moraine formation on the plateau, as the crests of some moraines are interrupted across the infilled channels (Fig. 4A). Thick accretionary wedges of normally consolidated glacigenic sediments formed in places at the edge of the till plateau, seemingly fed in part from these incised channels. This plateau of till has been eroded and the eroded sediments moulded into a series of subglacial bedforms (Fig. 6), creating the central glacial terrain with subglacial bedforms on the seabed surface (Zone 2).

Forming Flow Set (FS) 1, large-scale ridges of >30 km long, about 20–25 m high and with elongation ratios approaching 60:1 occur as a broad 45 km wide swathe across this Zone 2. Interpreted as mega-scale glacial lineations (MSGLs – Fig. 4B; Fig. 6), these bedforms are partly buried by normally consolidated sediments in the east of Zone 2 and their sizes compare well with reported MSGL (Spagnolo et al., 2014). The integration of BH 32a and BH33 with the sub-bottom profile over an MSGL exemplifies the differences in sediment properties between the MSGL crest and trough (Fig. 2; 6), with over-consolidated gravelly sands and clays forming the MSGL crest, and mainly normally consolidated gravelly clays filling in the troughs between MSGLs. Across the large study area the orientation of MSGLs gradually changes. In the northwest and central region of Zone 2, the MGSLs are orientated 205°, parallel to the main axis of the ISB, but curve westerly 240° moving progressively southeast (Fig. 5B;6), and thus guided probably by the topography of the island of Anglesey (Phillips et al., 2010 - Fig. 1).

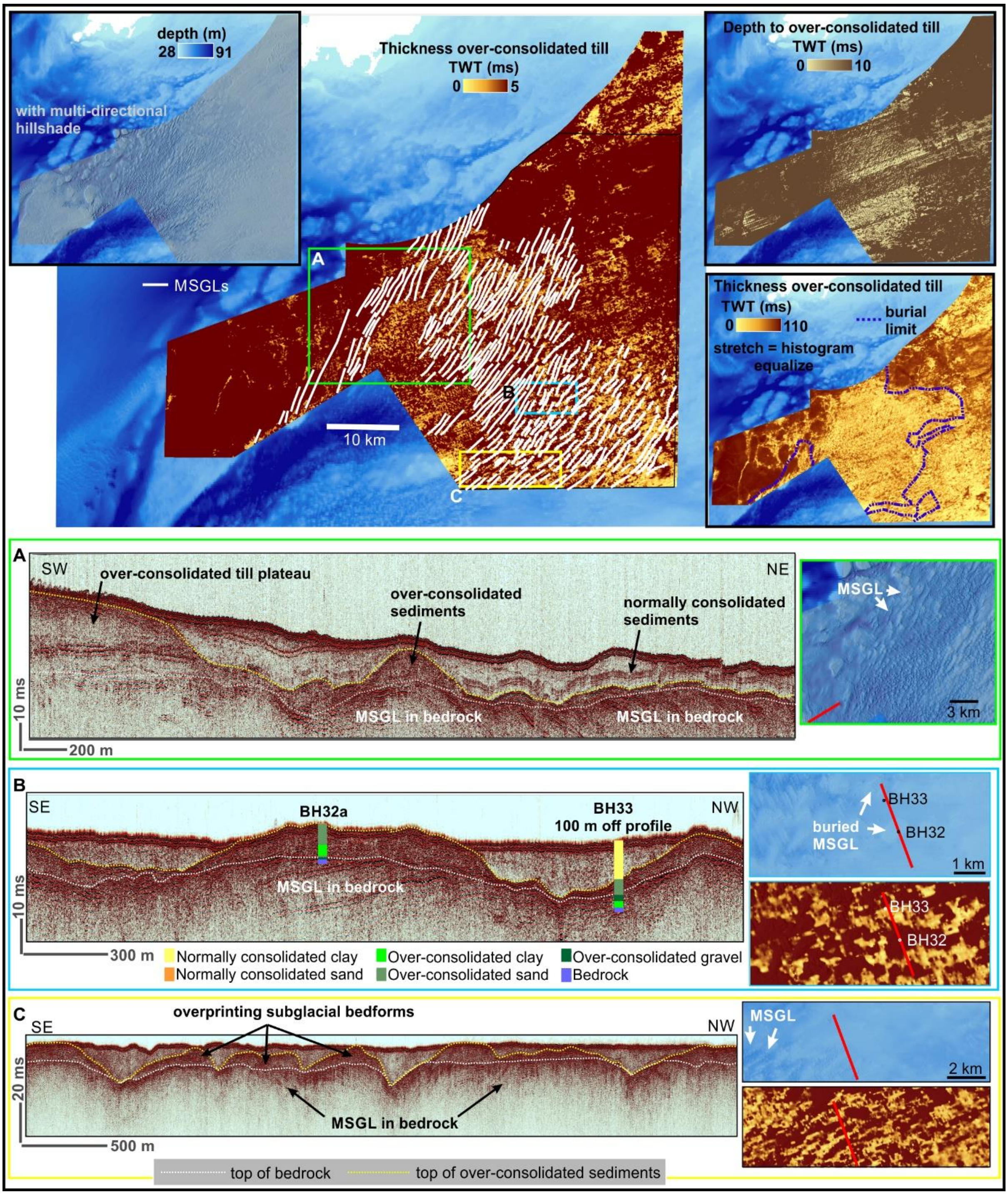



Figure 6. Data of the surface and sub-surface expression of mega-scale glacial lineations (MSGLs) from 3D grids of depth to the seabed surface, depth to the over-consolidated till and

the thickness of the over-consolidated till. The latter was presented with both a linear color gradient and with a stretched classification system using histogram equalisation to enhance contrast across this large study area with large variations in till thickness. Panels A, B and C display 2D seismic profiles perpendicular to the long axis of MSGLs, with a depth scale in Two-Way-Travel Time (TWT). Panel A shows the erosional edge of the subglacial till plateau and subglacial bedforms buried underneath normally consolidated sediments and formed partly erosional into the underlying bedrock. Panels B and C display partly and fully buried subglacial MSGLs constructed from erosion into bedrock and deposition of till.

A key interpretation integrated from the analysis of 16500 km of sub-bottom profiles, is that the swathe of MSGLs is flanked to the west, north and east by the remaining plateau of over-consolidated till and that the MSGLs have interspaced grooves that incise into the bedrock (Fig. 6), creating lower-amplitude MSGL in the bedrock surface over which over-consolidated till was deposited (Fig. 2; 6). No meltwater-related deposits were observed on-lapping the over-consolidated sediments in the MSGLs and the bedrock in between the MSGLs is overlain by a very thin layer of over-consolidated till, or directly by normally consolidated sediments. There is thus no evidence for bedforms to have formed via subglacial floods, and the MSGLs were seemingly formed via subglacial erosion into bedrock and deposition of till. The bedrock, predominantly Carboniferous sand stones, siltstones and mudstones in this area, was thus eroded into and left temporarily exposed as part of the subglacial bed due to MSGL formation. This erosional phase of ice streaming has an ISIS axis-parallel alignment. To sustain ice flow parallel to the axis of the ISIS on the seafloor and across Anglesey (Phillips et al., 2010) requires an ice front significantly to the south of the constriction between Wicklow and the Llŷn Peninsula. This erosional phase is therefore associated with geochronological Stages 1 and 2, during extension to maximum limits and/or with ice streaming during the initial rapid retreat (27–25 ka BP) (Praeg et al., 2015; Scourse et al., 2019; Smedley et al., 2017b, Small et al., 2018).

### 3.4 Frozen bed and onset of warm-based ice streaming

Fields of moraines (Fig. 4C) in the central Irish Sea Basin with often jigsaw-puzzle arrangements have been interpreted as ribbed moraines (Van Landeghem et al., 2009), with a formation mechanism of brittle fracture of frozen till aligned with that proposed by Kleman and Hättestrand (1999). These ribbed moraines gradually evolve down-ice where moraines start having a clearly tapered side interpreted as drumlins (Fig. 4D; Van Landeghem et al., 2009), and can be explained by time and space transgressive model by Kleman and Hättestrand (1999) where this transition occurs due to a phase change of frozen bed to thawed bed. The data presented here, show that ribbed moraines and drumlins overprint the MSGL terrain (Fig. 3; 5B–D). The ribbed moraines are transversely orientated to a 225–255 °N ice flow and display evidence for topographical focusing of ice flow as the long axes of ribbed moraines are transverse to ice flow direction, and these axes broadly follow the contour lines of the background bathymetry (Fig. 5C). Identical ice flow alignment is displayed by the drumlins, elongated parallel to the ice flow direction, and together with the ribbed moraines they form FS2. The area of drumlinised ribbed moraines between the ribbed moraine and drumlin field (Fig. 5; Van Landeghem et al., 2009) reflects time-transgressive changes to a lubricated and less rigid bed rheology (cf. Kleman and Hättestrand, 1999). Towards the western steeper slopes of Zone 2, drumlin elongation ratios increase from <1.5:1 to 5:1 and the converging directions of the long drumlin axes are indicative of a convergence of south-westerly ice flows guided by the topography. To the east and south-east of Zone 2 the glacial landscape is buried (Zone 3), and MSGLs, ribbed moraines and drumlins are visible in the sub-bottom acoustic data (Fig. 4C; 4D). Both drumlins and ribbed moraines display an ice-flow alignment more in keeping with geochronological Stage 3. This westerly - south-westerly ice flow alignment corresponds with short-lived stabilisations of the ice margin at the constriction of the Irish Sea between Wicklow and Llŷn Peninsula at 22.6 ± 1 ka BP (Small et al., 2018; Smedley et al., 2017a), draw-down of the ice surface, greater topographical focusing and the inception of ice-free marine conditions in the western ISB (Chiverrell et al., 2018).

### 3.5. Ice flows into ice-free western ISB

Smaller bedforms overprint and have moulded locally all the previous ice-flow sets and are interpreted as flutes or flutings (Fig. 4E). The numerous lineations vary from a westerly direction on flat seabed to a south-westerly direction where they are conditioned by and follow the seabed topography (FS3 - Fig. 5E). The flutings often overprint multiple bedforms (e.g. drumlins and ribbed moraines) and their dimensions of ~60 m wide and 1500 m long approach those of mega-flutes (Bluemle et al., 1993). Formation of such elongated flutings are primarily associated with a temperate and grounded glacial land-system (Evans and Twigg, 2002; Stokes and Clark, 2002) and have been linked with surge type ice flow behaviour and high basal ice velocities (Bluemle et al., 1993; Waller et al., 2008).

Our interpretation of flow sets 2 and 3 (ribbed moraines, drumlins and flutes), is that a phase of reduced ice streaming allowed frozen bed conditions and ribbed moraines to form. A resumption of warm-based ice streaming cased the drumlinisation of the ribbed moraines as the thawing front moved eastwards. The subsequent ice flow surge stopped the thawing front migrating eastwards and resulted in flutings over the top of drumlins and the most westerly ribbed moraines. The ice margin then lifted off temporarily to preserve the remaining ribbed moraines from further subglacial reworking.

Eskers are also preserved well in Zone 2 and their direction is seemingly controlled by the changes in topography (Fig. 4F; 5F). Their spacing varies between 2 and 8 km and their general direction aligns with the ice flow directions derived from FS2 (ribbed moraines and drumlins). They are found mostly superimposed on FS1–3 (MSGLs, ribbed moraines, drumlins and flutes) (Fig. 5F), but occasionally are moulded into flutes and so may be contemporary with the end of FS3 (Fig. 4F2). As the eskers align with ice flow directions, we favour esker formation by a time-transgressive build-up of sediment  beneath a more slowly retreating ISIS margin as opposed to instantaneous formation due to large meltwater floods, which would see more frequent eskers in a larger range of directions (Hewitt, 2011).

Glacifluvial bedforms and associated sediments are sparsely distributed in Zone 2, limited to these confined englacial/subglacial tunnels systems and isolated large depositional forms. These forms include subaqueous fans formed against topographic highs that provided pinning points for probably short-lived grounding-line stabilisation (Fig. 7). The restricted nature of outwash deposits prevented potentially the burial of the glacial landscape of Zone 2. This phase of warm-based ice streaming requires an ice-free western ISB, and is most likely associated with the transition from geochronological Stage 3 to 4, between 21.4 ± 1.0 ka and 19.8 ± 0.8 ka BP during which the western and deeper (100–140 m bmsl) ISB had deglaciated (Ballantyne and Ó Cofaigh, 2017; Chiverrell et al., 2018; McCabe, 2008).

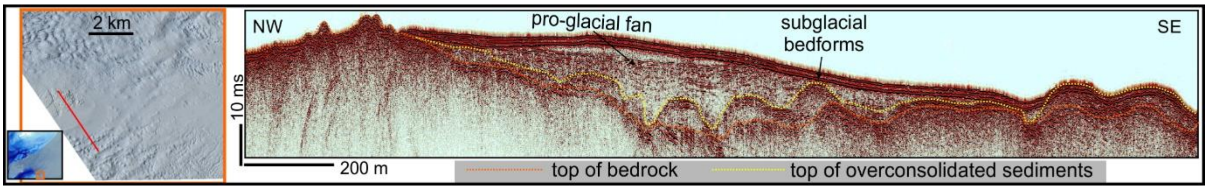



Figure 7. A fan of proglacial sediments deposited down-ice from a bedrock outcrop. The subglacial bedforms buried by the proglacial sediments were constructed by erosion into bedrock and deposition of till.

### 3.6 Ice margin retreat towards east and northeast

Small ice-flow transverse ridges in a lobate planform overprint FS1–3 in the north of Zone 2 (Fig. 4G; 5G). Typically 5 m in height, 100–150 m in width and about 1 km long, these regularly-spaced ridges are often broken up, but can be interpreted as ice marginal or recessional moraines formed by small-scale oscillations of a retreating lobate ice margin. The recessional moraines were the last grounded bedforms created in the area, with few flutes discernible and probably reworked by moraine formation during retreat characterised by backstepping and minor re-advances of the ice margin. Zones 1 and 2 are densely overprinted by numerous well-preserved iceberg pits and elongated scour marks (Fig. 4I; Van Landeghem et al., 2009), suggesting the generation of large numbers of icebergs and the possible collapse of the ice stream in a subaqueous environment (Van Landeghem et al., 2009). The eastern extent of the iceberg scour field coincides with a break in the background topographic slope of the region at 40–50 m water depths, in turn matching the burial limits of the glaciated terrain (Fig. 1B; 5I). A series of long moraines in Zone 4 (Fig. 1B; 4H;5H) are interpreted as ice-front moraines formed parallel to the retreating ice margins. This last phase of ice retreat is attributed here to geochronological Stages 4–5, and retreat of the ice margin across the shallower (50–20 m bmsl) eastern ISB moving north across the Isle of Man between 20.7 ± 0.7 and 19.3 ± 0.8 ka BP (Chiverrell et al., 2018).

## 4. DISCUSSION

### 4.1 Changing ice flow dynamics with evolution of the Irish Sea Ice Stream

The evolution of the ISIS from 27 to 18 ka BP is described here in five chronological stages (Section 1.2, and all the references herein). The ice-flow sets (FS) interpreted from glacial bedform assemblages (Fig. 5J) are organised chronologically in Fig. 8E–I, contextualised with these five stages of the advance and retreat. Ice flow is evidenced in Stage 1 (27–25 ka BP) by large-scale MSGL (FS1 – Fig. 8A), providing the first definitive subglacial bedform evidence for ice streaming within, and axis-parallel to, the ISB, and thus attributed to ice margins located south of the Llŷn Peninsula – Wicklow constriction (Fig. 8F). Channelised meltwater floods created large and deep tunnel valleys in Zone 1, at a time when meltwater production under the ISIS potentially increased as ice flow was faster. The dataset presented here shows two components to MSGL formation: subglacial erosion and till deformation. During ISIS advance, bedrock erosion resulted in grooves and low-amplitude MSGLs, leaving bedrock exposed on the ice–bed interface. A second and less erosive component to MSGL formation left deformable till redistributed and deposited as crests of the MSGLs. These two components must have occurred within the same flow phase as the MSGL grooves align perfectly with the crests. The FS1 flow alignment and long MSGLs favour formation by discharge of an ice stream thick enough to generate this erosion (cf. Motyka et al., 2006) and to reach the Celtic Sea (e.g. Praeg et al., 2015; Lockhart et al., 2018 – Fig. 8F), at net advance rates of ~350 m $a^{-1}$ (Smedley et al., 2017a). MSGL formation probably continued during the early phases of margin retreat (Stage 2), but then a significant change occurs. Analyses of FS2 reveals a 20° and 60° change in ice flow direction, and formation of ribbed moraines implies a slow-down in ice flow as they form in the ice stream onset zone where the bed is frozen (e.g. Kleman and Hättestrand, 1999). At the end of Stage 3, ice margins retreated and stabilised north of the Wicklow-Llŷn limit during early Stage 4, developing westerly ice flow trajectories and oscillatory dynamics of the ice margin (Smedley et al., 2017a; Thomas and Chiverrell, 2007), conditions that are consistent with observations from FS2 (Fig. 8b; 8G). The drumlinisation has an erosional component carving into the bedrock surface (Fig. 4C; 7) and delineates the onset of warm-based ice streaming in FS2. As Zone 2 is progressively drumlinised (Fig. 5), faster ice was flowing (c.f. King et al., 2016; Smith and Murray, 2009) from the shallow Zone 4 (Fig. 1B) towards accommodation space growing in the western ISB (end of Stage 3). Drumlinisation was encouraged by the preceding ice stagnation during ribbed moraine formation and the associated ice thickening providing sufficient frictional heat to potentially melt basal ice and for the renewed ice streaming to deform the stiffer basal diamictons. The timing of Stages 3 and 4 (22.6±1 to 20.7 ± 0.7 ka BP) overlaps with cool conditions in the North Atlantic (Rasmussen et al., 2014) and thermodynamic glacial processes are thus favoured as the main driver for this flow-reactivation. This is an important consideration in understanding present-day ice stream behaviour, as ice flow shut down and abrupt changes in ice flow directions are observed and predicted to occur more frequently for Antarctic ice streams (c.f. Conway et al., 2002). Fluting reflects a later realignment and last phase of faster ice flow (FS3) at the start of Stage 4, which is attributed to ice surging into deep waters to the west (Fig. 8C; 8H). The end of the fluting phase is associated with a lubricated subglacial environment and extensive drainage patterns, with extensive iceberg calving. The collapse of the ice stream was imminent with ice margin retreat patterns seemingly topographically conditioned as small recessional moraines and large end-moraines formed in a direction consistent with ice pulling back towards the Isle of Man and NW England (Fig. 8D; 8I).

Sectors of the MSGL and ribbed moraine terrain were not drumlinised and we attribute this to temporary lift-off of the ice bed in those areas driven by ice thinning due to intense iceberg calving rather than sea level rise, which was probably limited over the estimated time of grounding line fluctuations. Where the ice margin grounded again at the top of the break in regional slope, it stagnated, with restricted calving and prolonged subglacial meltwater drainage providing sediments to bury the eastern edge of the glacial terrain. Burial may have continued during Stage 5, when Scottish ice re-advanced towards the northern edge of the study area (Fig. 8D; 8I).

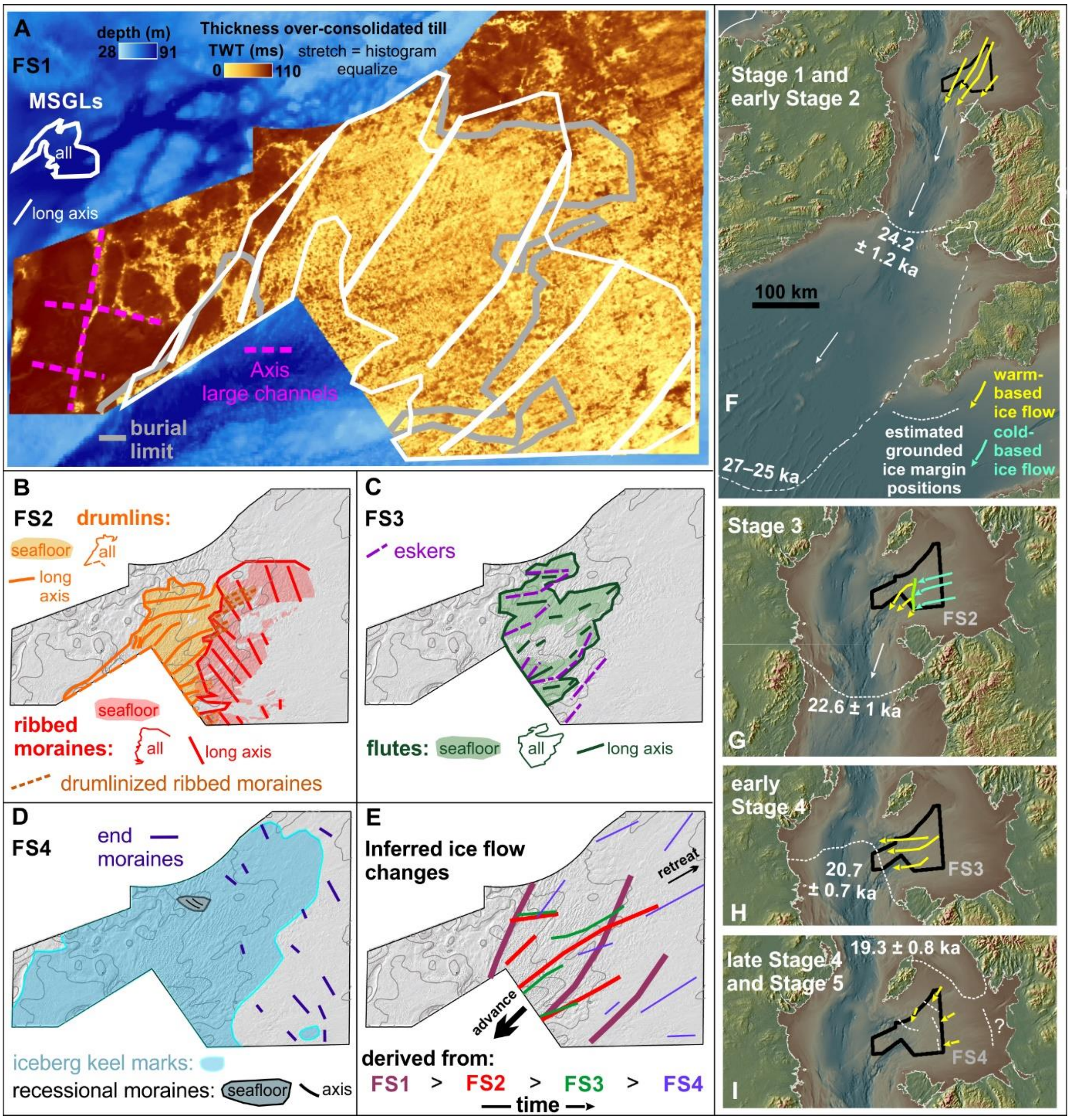



Figure 8. Summary geomorphology with Flow Sets (FS1–3) represented schematically. Panel (A) shows FS1 on over-consolidated glacigenic sediment thickness data in Two-Way-Travel Time (TWT) and with a stretched classification system using histogram equalisation to enhance contrast. Panels (B–E) display Flow Sets on contoured present-day seabed topography (see Fig. 1B). Panels F–I contextualises the temporal and spatial evolution of the palaeo-glacial landscape in the studied area (outlined in black) with the published

geochronology for the retreat of the former ISIS (see Section 1.2) to visualise the changes in ice flow dynamics.

**4.2 Geological factors regulating ice sheet drainage?**

Our qualitative observations of changing bed conditions provide an insight into the link between ice streaming and subglacial geological factors. Regional-scale erosion occurred into the original till plateau and into the bedrock during formation of meltwater flow channels in Zone 1 and the >30 km long MSGLs in Zone 2. In Zone 2, the deformable sediments were largely removed where MSGLs were formed. This is evidenced through the integrated interpretation of 16500 km of individual sub-bottom geophysical profiles (Fig. 4; 6; 8), where the interpretation of the base of the over-consolidated glacigenic sediments (as gridded from the seismic horizons) show clearly that the swathe of MSGLs is flanked to the west, north and east by the remaining plateaus of over-consolidated till (Fig. 6; 8), where the deviated ice flow path around the Isle of Man was likely most erosive in the immediate pathway of the deviation. This pervasive erosional character in Stages 1 and 2 of ISIS advance and the initial retreat agrees with observations beneath contemporary (Jezek et al., 2011) and palaeo-ice streams with limited deformable sediment at the ice–bed (e.g. Bradwell, 2013; Krabbendam et al., 2016; Smith, 1948). We suggest that the erosional processes involved during fast ice streaming in the ISB induced a change in the geological factors steering the ice streaming, as after the near complete removal of a deformable substrate from Zone 2, the bedrock became exposed under the ice. These changes in ice–bed conditions would have adjusted ISIS dynamics internally, as the style of basal sliding and subglacial drainage changes when the bedrock underneath the eroded till is reached. Increased basal drag can slow down ice flows and trigger basal freezing (c.f. Jacobson and Raymond, 1998). This may have happened to the ISIS after formation of the MSGLs, because the ice bed must have frozen for ribbed moraines to form. Passage of the ice margins northwards of the constriction between Wicklow and the Llŷn Peninsula may also have contributed towards a slow-down in ice stream velocity. When the downstream end of the retreating ISIS reached this area of exposed bedrock, the ice flow changed direction with ribbed moraines and drumlins (FS2). The change from fast flow over warm-based water-saturated soft diamictons, to a temporarily compressional, re-aligned and slower flow over cold-based and stiffer basal diamict is a significant change in ice stream dynamics. These observations from the palaeo-ISIS agree with a recent sensitivity analyses in ice stream models, finding changes in basal friction particularly influential for ice stream dynamics when these changes are felt near the grounding zone (Alley et al., 2019). In our case study, where the retreating palaeo-ISIS experienced topographical confinement and where the western ISB started to become ice free, we propose that the changes in basal friction will have been an additional contributing factor in ice stream shut-down, ice flow re-alignment and ultimate demise.

## 5. CONCLUSION

The evolution of a well-preserved glacial landscape in the central part of a large palaeo-ice stream details changes in ice flow dynamics during its advance and retreat that parallel changes in ice thickness, ice bed conditions and proglacial open water accommodation space. This reconstruction, linked to the robust geochronology of ice margin retreat, reveals that the erosive ice stream processes exposed the bedrock underneath the ice during ice advance and/or the early phases of deglaciation as MSGLs were formed via an erosive and subsequent depositional component within one ice-flow phase. We propose that the changes in basal friction due to bedrock exposure to the ice may have contributed to defining the nature of dynamics during deglaciation through slower ice flow velocities and basal freezing. During later phases of deglaciation, topographic variations in the subglacial bed near the grounding line increasingly conditioned ice flow as re-activation and re-alignment of ice streaming led to an ultimate demise. This new analysis highlights that ice streaming can be determined by how ice flow impacts local geology both prior to and during deglaciation, by altering the availability of deformable bed and bed roughness underneath the ice. The dynamics of a rapidly retreating palaeo-ice streams can thus be steered by the processes prior to the final retreat phase, identifying a need for detailed parametrisation of ice–bed conditions throughout all flow phases when forecasting episodes of changing retreat rate, shutdown and reactivation of present-day ice streams, still the least understood components of ice sheet dynamics.

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