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## Retreat and stabilization of a marine-based ice margin along a high arctic fjord-cross-shelf trough system

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1 **Retreat and stabilization of a marine-based ice margin along a High**  
2 **Arctic fjord-cross-shelf trough system**

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19 **Abstract**

20           Multibeam bathymetric and seismostratigraphic data collected in the Clyde fjord-  
21 cross-shelf trough system (eastern Baffin Island, Canadian Arctic Archipelago) display  
22 glacial landforms and depositional assemblages that enable the identification of the  
23 maximal extent of the Laurentide Ice Sheet (LIS) margin and delineating the patterns and  
24 controls on its subsequent retreat. Additionally, 10 new sediment cores – from which  
25 seven radiocarbon ages were acquired – allow the recognition of depositional processes.  
26 Results show that, during the Last Glacial Maximum, the LIS margin extended almost to  
27 the edge of the continental shelf. Early deglaciation of the trough was marked by an initial  
28 ice-shelf collapse and rapid retreat of the ice stream, as evidenced by the absence of ice  
29 marginal landforms and the presence of extensive iceberg ploughmarks across a large  
30 portion of the outer trough. It was followed by a slow retreat and successive stabilizations  
31 of the ice margin that led to the deposition of recessional moraines and GZWs.  
32 Deglaciation of the fjord in the early Holocene occurred in an episodic style, whereby  
33 rapid retreat was punctuated by relatively long standstills that enabled major moraine  
34 formation. Long-term stabilizations of the ice margin in the Clyde fjord-cross-shelf  
35 trough system are interpreted to coincide with major climatic cooling events, such as the  
36 Younger Dryas and early Holocene cold reversals. Ages derived from sediment cores and  
37 previous work suggest that higher retreat rates correspond with periods of significant  
38 global sea level rise, suggesting that oceanic forcing exerted a minor control on the  
39 deglaciation. GZWs and large moraine ridges are observed at pinning points in the trough,  
40 indicating that the location of ice margin stabilizations was influenced by topography.  
41 The reconstruction of the deglaciation of the Clyde fjord-cross-shelf trough system allows  
42 us to refine deglacial models for similar systems of northeastern Baffin Island, in  
43 particular beyond the coast and along the steeper section of the fjord where chronological  
44 gaps remained.

45

46 **Keywords:** Laurentide Ice Sheet; Baffin Island; Deglaciation; Holocene; Submarine  
47 landforms; Marine sediment cores; Abrupt climatic events

## 48 **1. Introduction**

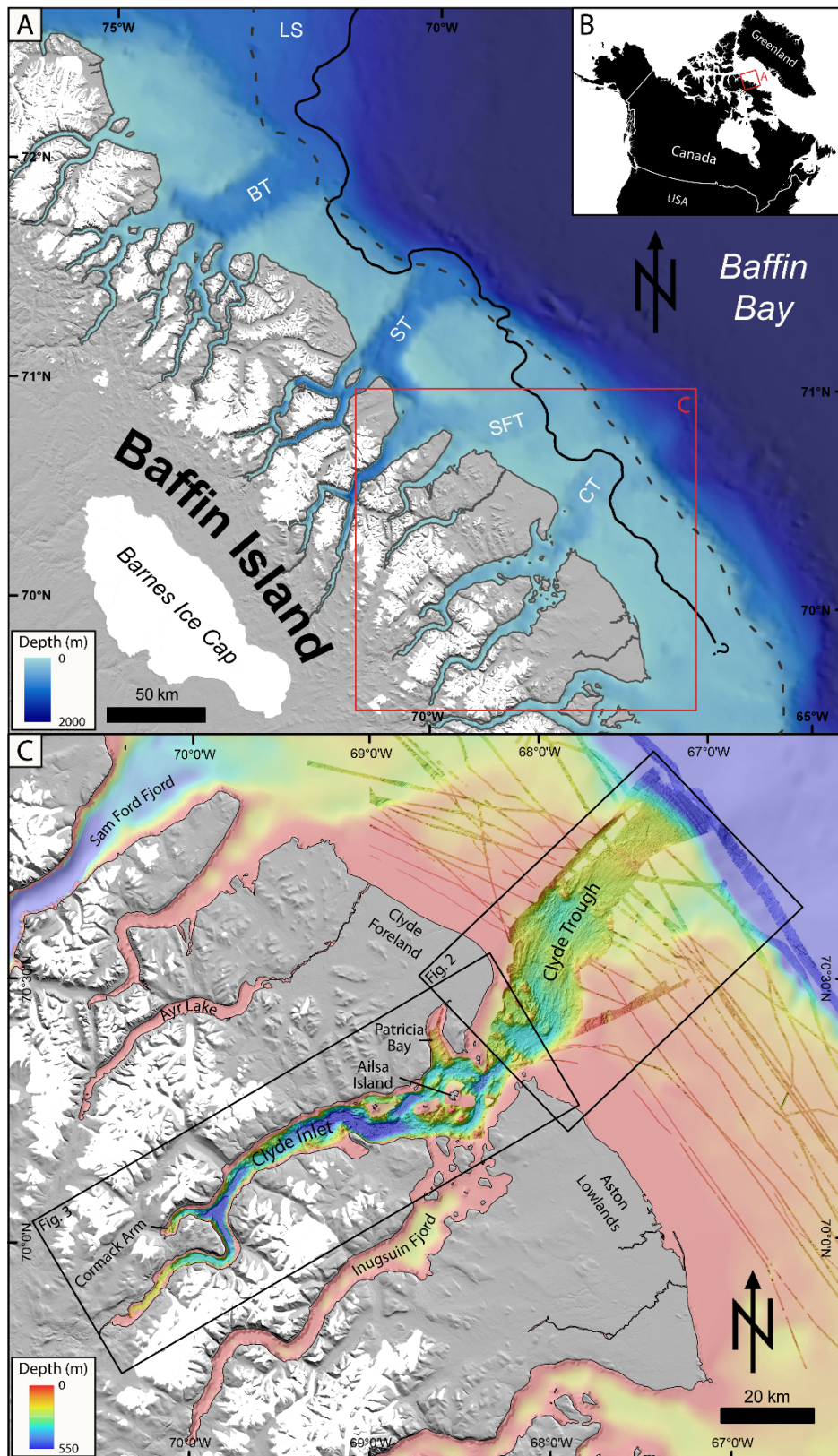
49           Reconstructing the history of past ice sheets allows to document their inter-  
50 relationship with the climate system, especially by providing boundary conditions to test  
51 models that aim to simulate climate and ice sheet evolution (Joughin et al., 2012; Stocker  
52 et al., 2013; Rasmussen et al., 2014; Tinto et al., 2019; Briner et al., 2020; Lowry et al.,  
53 2020). Past marine-based ice sheet margins such as those of the Laurentide Ice Sheet  
54 (LIS), which was in contact with the western Atlantic Ocean during the Last Glacial  
55 Maximum (LGM; ~24 ka BP), are believed to have operated in a comparable way to the  
56 present Antarctic Ice Sheet, with a network of ice streams channeled in bathymetric  
57 troughs and mass loss dominated by calving (Ottesen et al., 2005; De Angelis and  
58 Kleman, 2007; Margold et al., 2018). Reconstructing palaeo-ice sheet dynamics and  
59 deglaciation patterns is therefore fundamental for understanding the long-term behaviour  
60 of modern ice sheets (i.e., Antarctica and Greenland) and their interconnection with global  
61 climate (i.e., Danielson and Bart, 2019; Smith et al., 2019; Briner et al., 2020; Young et  
62 al., 2020, 2021).

63           Controls and ice-retreat patterns during the deglaciation that followed the LGM  
64 still remain poorly documented in the eastern Canadian Arctic Archipelago (CAA)  
65 compared to other deglaciated regions of the Northern Hemisphere, such as Norway (e.g.,  
66 Lyså and Vorren, 1997; Aarseth et al., 2007; Laberg et al., 2009; Bjarnadóttir et al., 2012),  
67 Svalbard (e.g., Winsborrow et al., 2010; Streuff et al., 2018; Flink and Noormets, 2018;  
68 Howe et al., 2019; Allaart et al., 2020) and the British Isles (e.g., Arosio et al., 2018;  
69 Callard et al., 2020). Off Baffin Island, marine geophysical data suggest that ice retreat  
70 was punctuated by successive stabilizations of the late glacial ice margin on the  
71 continental shelf (Brouard and Lajeunesse, 2017), while onshore data suggest a rapid ice  
72 decay into the fjords during the early Holocene (Miller et al., 2005; Briner et al., 2005,



73 2007, 2009a; Young et al., 2015; Margreth et al., 2017). However, recent offshore  
74 investigations show that multiple ice-margin stabilizations occurred within the fjords of  
75 northeastern Baffin Island during the last deglaciation (Brouard and Lajeunesse, 2019a).  
76 Stabilizations both on the shelf and in the fjords of northeastern Baffin Island were found  
77 to be strongly influenced by bed geometry (Brouard and Lajeunesse, 2017, 2019a),  
78 similarly to observations made in other deglaciated fjords and continental shelves (i.e.,  
79 Aarseth et al., 1997; Hodgson et al., 2014; Batchelor et al., 2019a). These contrasting  
80 patterns of ice retreat, which probably arise from the scarcity of valuable data from the  
81 submarine domain, complicate the establishment of a reliable model of deglaciation and  
82 outline the need for improved understanding of LIS extent and retreat patterns in the fjord-  
83 cross-shelf trough systems of Baffin Island by bridging submarine and continental data.

84         Here we map and analyse submarine landforms and sediment assemblages in the  
85 180 km-long Clyde fjord-cross-shelf trough system, with the aim of refining our  
86 understanding of the retreat patterns of the marine-based margin of the northeastern LIS.  
87 By using a combination of multibeam bathymetric data, acoustic and seismic profiles as  
88 well as sediment cores, the key objectives are to 1) define a chronology for the  
89 deglaciation of the LIS in the region; 2) reconstruct changes in the morpho-sedimentary  
90 system and dynamics on the shelf and in the fjord; and 3) identify the factors that  
91 controlled ice margin retreat rates and stabilization during deglaciation.



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**Figure 1** (A) Baffin Bay and the proposed LGM limit on northeastern Baffin Island, modified from Brouard and Lajeunesse, 2017. LS: Lancaster Sound; BT: Buchan Trough; ST: Scott Trough; SFT: Sam Ford Trough; CT: Clyde Trough. The dashed line represents the LIS extent at the LGM from Dalton et al., 2020. (B) Location of the study area. (C) Map showing the multibeam bathymetric data used in this study draped on the International Bathymetric Chart of the Arctic Ocean data gridded at a 500 m cell-size resolution (IBCAO; Jakobsson et al., 2014). The hillshade is from the Canadian Digital Elevation model (CDEM).

## 100 **2. Regional setting**

### 101 **2.1. Study area**

102 The studied area comprises the fjord of Clyde Inlet and its offshore extension  
103 across the continental shelf, namely Clyde Trough (Fig. 1). This fjord-cross-shelf trough  
104 system stretches from the interior plateau of Baffin Island to the shelf break facing Baffin  
105 Bay.

106 The continental shelf off Clyde Inlet is generally shallow ( $< 200$  m) with a deeper  
107 trough carved by repeated occupation by ice streams during Quaternary glaciations  
108 (Løken and Hodgson, 1971; Praeg et al., 2007). Precambrian crystalline rocks extend  
109 halfway across the continental shelf, where they are overlain by upper Cretaceous-  
110 Tertiary strata of the Baffin Bay Basin (Jackson et al., 1984; Fader et al., 1989; Praeg et  
111 al., 2007). Clyde Trough is 20 to 30 km-wide, extends from SW to NE, on  $>60$  km, and  
112 bends slightly toward the east near the shelf break. Existing coarse resolution bathymetry  
113 data (GEBCO) show that the trough is flat-bottomed and has a reverse-gradient slope  
114 (Fig. 1C), characteristic of troughs in formerly glaciated areas (Ottesen et al., 2007;  
115 Slabon et al., 2016; Arndt et al., 2017; Bart et al., 2017; Brouard and Lajeunesse, 2017),  
116 with water depths decreasing seaward from  $\sim 375$  m at the mouth of the fjord to  $\sim 150$  m  
117 at the shelf break.

118 Clyde Inlet is a fjord that has been glacially incised into the Precambrian  
119 crystalline rocks of the eastern coastal mountains of Baffin Island during Quaternary  
120 glaciations (Jackson et al., 1984; Kessler et al., 2008). With its U-shaped profile and a  
121 succession of deep basins separated by intervening sills, it has a typical mid- and high-  
122 latitude fjord morphology (Syvitski and Shaw, 1995). Clyde Inlet is 120 km-long,  
123 between 3 and 20 km-wide and consist of basins between 200 and 500 m-deep (Fig. 1C).  
124 Many hanging valleys of various widths and depths drain into Clyde Inlet, providing

125 localized sediment input through proglacial rivers. Inugsuin Fjord, a 100 km-long fjord,  
126 and Patricia Bay also merge into Clyde Inlet near its mouth. Clyde Inlet is directly  
127 connected to the cross-shelf trough, forming a 180 km-long submarine glacial valley  
128 system connecting Baffin Island to Baffin Bay and separated by a shallow bedrock sill at  
129 the fjord mouth.

130 Most of the valleys, lowlands, fjords and trough of Clyde Inlet and nearby areas  
131 are covered by a thick sequence of Quaternary deposits (Jackson et al., 1984; Praeg et al.,  
132 2007; Brouard and Lajeunesse, 2017, 2019a, 2019b).

## 133 **2.2. Late Quaternary glacial history**

134 Late Wisconsinan (MIS 2) ice sheets started building up at ~28 ka BP (Stokes et  
135 al., 2012; Klassen et al., 2010; Batchelor et al., 2019b) and reached their maximal extent  
136 by 24 ka BP in many sectors of North America (Dyke, 2002; Andrews and Dyke, 2007;  
137 Hughes et al., 2013). Although many models have been proposed for establishing the  
138 maximal extent of the LIS at the LGM, recent studies indicate that the ice margin  
139 extended at or near the edge of the northeastern Baffin Island shelf (Briner et al., 2006;  
140 Li et al., 2011; Brouard and Lajeunesse, 2017; Jenner et al., 2018; Dalton et al., 2020;  
141 Lévesque et al., 2020; Couette et al., 2022). In between the fjords, cosmogenic exposure  
142 ages on glacial erratics and bedrock suggest an extensive LIS cover during the LGM  
143 (Briner et al., 2005; Davis et al., 2006). Marine geophysical data also show that an ice  
144 shelf covered northern Baffin Bay during the LGM, possibly buttressing peripheral ice  
145 streams and impacting their flow to the ocean (Couette et al., 2022).

146 Following the LGM, the break up of the Baffin Bay ice shelf between 16 and 14.6  
147 ka BP provoked a major reorganization of the ice-sheet drainage system by removing the  
148 buttressing effect on ice streams (Couette et al., 2022), and therefore led to ice-flow  
149 acceleration in systems from northeastern Baffin Island (Jenner et al., 2018). This event

150 was found to be more or less coeval with the deposition of a detrital carbonate sediment  
151 layer in Baffin Bay (BBDC-1), corresponding to massive ice discharge from ice streams  
152 of the CAA and northwest Greenland and characterized by Ca-rich sediments (Hiscott et  
153 al., 1989; Andrews et al., 1998; Simon et al., 2014; Jackson et al., 2017; Jenner et al.,  
154 2018). This iceberg discharge event was followed by a period of important ice retreat on  
155 the shelf off eastern Baffin Island coinciding with the Bølling-Allerød warm period (Dyke  
156 et al., 2002; Briner et al., 2005; Margreth et al., 2017; Brouard and Lajeunesse, 2017;  
157 Jenner et al., 2018). The subsequent retreat towards Baffin Island mainland was  
158 punctuated by glacier readvances or stabilizations during the cold phase of the Younger  
159 Dryas (12.9-11.7 ka BP) which led to the building of major moraine systems (Briner et  
160 al., 2007; Margreth et al., 2017; Young et al., 2020). Following the Younger Dryas, and  
161 under the warmer climate of the early Holocene period, the ice margin retreated rapidly  
162 to reach the inner fjord (Andrews and Ives, 1978; Briner et al., 2007, 2009b). This rapid  
163 ice-margin retreat was temporarily interrupted by ice-margin stabilizations during short  
164 cold events at 10.3, 9.3 and 8.2 ka (Briner et al., 2007, 2009b; Young et al., 2012, 2020;  
165 Crump et al., 2020). The latter event is recorded at the fjord heads on eastern Baffin Island  
166 and referred to as the Cockburn stage moraine (Andrews and Ives, 1978; Briner et al.,  
167 2007, 2009b; Brouard and Lajeunesse, 2019a; Young et al., 2021). Subsequently, the LIS  
168 margin retreated steadily from the fjord head and separated to form what are today's  
169 Barnes and Penny ice caps (Dyke, 2004; Miller et al., 2005; Briner et al., 2009b).

### 170 **3. Material and methods**

171 This study is based on multibeam bathymetric data, shallow acoustic data, seismic  
172 data, and sediment core data. The main bathymetric dataset was acquired during  
173 expedition MSM66 of the German research vessel RV Maria S. Merian in August 2017  
174 (Dorschel et al., 2017). The MSM66 data are complemented by bathymetric data collected

175 during ArcticNet cruises onboard the CCGS Amundsen (2003-2017). Shallow acoustic  
176 data and sediment cores were also collected during MSM66 (Dorschel et al., 2017).  
177 Seismic data were acquired by the Geological Survey of Canada during airgun surveys in  
178 1978 and 1980.

### 179 **3.1. Multibeam bathymetric data**

180 MSM66 bathymetric data were collected with a Kongsberg Simrad EM122 (12  
181 kHz) multibeam echosounder (MBES), while the CCGS Amundsen data were collected  
182 with a Kongsberg Simrad EM302 (30 kHz) MBES. These datasets were processed for  
183 anomalous data points and artefacts removal using Caris Hips and Sips software. The  
184 datasets were gridded with a 10 m cell-size resolution and then imported into ESRI  
185 ArcGIS 10.8 software for geomorphological mapping and interpretation.

### 186 **3.2. Shallow acoustic and seismic data**

187 Shallow acoustic data were recorded with an Atlas Parasound DS P-70 system (5-  
188 33 kHz) during MSM66. The raw data were recorded into PS3 format and then converted  
189 into SEG Y using ps32segy software of Dr. Hanno Keil (University of Bremen). The  
190 acoustic profiles were then imported into the *SMT Kingdom Suite* software for processing  
191 and interpretation.

192 Seismic lines 78029\_AG\_275\_0130 and 80028\_AG\_RAYT\_257\_0200 were  
193 acquired on expeditions 78029 (1978) and 80028 (1980) by Brian MacLean of the  
194 Geological Survey of Canada-Atlantic (Cruise reports available via [https://ed.marine-  
195 geo.canada.ca/cruise\\_report\\_e.php](https://ed.marine-geo.canada.ca/cruise_report_e.php)). Extraction and interpretation was performed using  
196 the *LizardTech GeoViewer* software. Both acoustic and seismic data were transferred into  
197 Adobe Illustrator for figure production. Thicknesses and water depth were calculated  
198 using a velocity of 1500 m/s.

199 **3.3. Core data**

200 The sediment cores collected during expedition MSM66 were between 137 and  
201 965 cm long (Table 1). On board, all cores were split, visually described with a particular  
202 focus on noting the lithology, texture, contacts, sedimentary structures and Munsell  
203 colour, as well as digitally photographed.

204 **Table 1** Information on cores collected in the Clyde fjord-cross-shelf trough system.

Core name	Latitude	Longitude	Water depth (m)	Core length (cm)
GeoB22344-3	70°03.90'N	70°02.93'W	367	483
GeoB22346-3	69°54.18'N	70°13.54'W	203	783
GeoB22348-3	69°58.47'N	69°57.47'W	362	896
GeoB22350-3	70°08.52'N	69°44.36'W	435	137
GeoB22351-3	70°10.16'N	69°38.30'W	364	523
GeoB22353-3	70°13.35'N	69°00.16'W	489	862
GeoB22356-3	70°27.68'N	67°58.36'W	338	965
GeoB22357-3	70°36.28'N	67°53.63'W	315	902
GeoB22358-3	70°41.69'N	67°41.83'W	261	500
GeoB22359-3	70°46.06'N	67°27.96'W	196	166

205

206 X-ray fluorescence (XRF) scanning were used to characterize elemental  
207 properties of the sediments by using a XRF Core Scanner II (AVAATECH Serial No. 2)  
208 at MARUM, University of Bremen. Data were collected every 2 cm down-core over a 15  
209 mm<sup>2</sup> area with down-core slit size of 10 mm using generator settings of 10 kV, a current  
210 of 0.2 mA and a sampling time of 10 seconds directly at the split core surface of the  
211 archive half. The split core surface was covered with a 4 µm thin SPEXCerti Prep  
212 Ultralene1 foil to avoid contamination of the XRF measurement unit and desiccation of  
213 the sediment. Raw spectra data were processed by the analysis of X-ray spectra in the  
214 Iterative Least square software (WIN AXIL) package from Canberra Eurisys.

215 Accelerator Mass Spectrometry (AMS) radiocarbon dating was carried out on  
216 benthic foraminifera assemblages and shell fragments. Samples were sent to the  
217 MICADAS-laboratory (Alfred Wegener Institute, Bremerhaven), where CO<sub>2</sub> from small  
218 amount of foraminiferal carbonate (~0.5mg) were analysed. Additionally, three dates

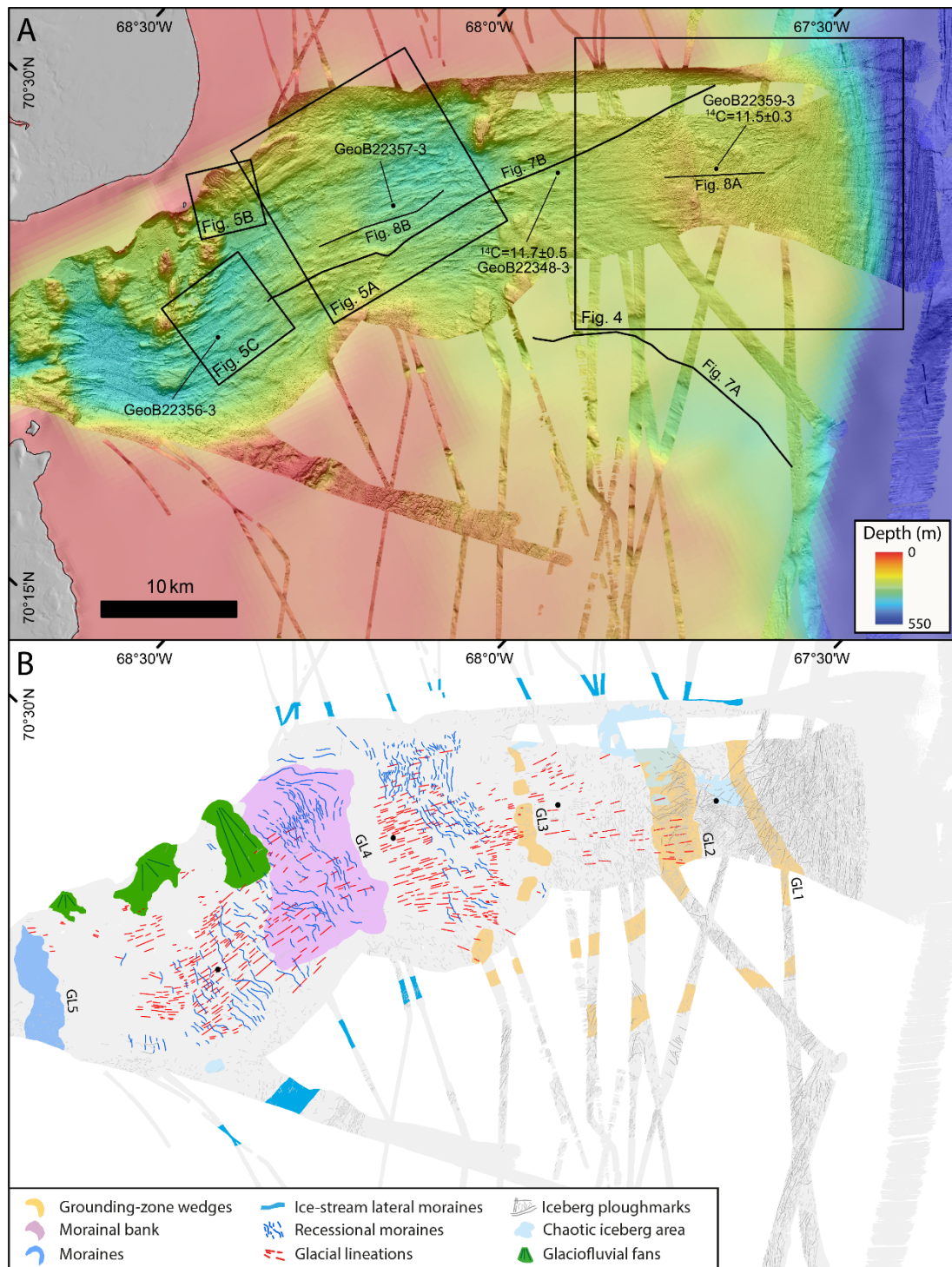
219 based on shell fragments were obtained. The AMS  $^{14}\text{C}$  ages were converted to calendar  
220 years before present (cal. BP) using the online software Calib 8.2 with the Marine20  
221 radiocarbon age calibration curve (Heaton et al., 2020). The Marine20 calibration curve  
222 uses a global marine reservoir age that is not suitable for the polar regions, an issue that  
223 requires the application of additional reservoir correction ( $\Delta\text{R}$ ) values to high-latitudes  
224 samples (Heaton et al., 2020). Therefore, an additional local  $\Delta\text{R}$  of  $81 \pm 18$  was used to  
225 account for the regional offset of the world ocean  $^{14}\text{C}$  age (Pieńkowski et al., 2022). The  
226  $\Delta\text{R}$  value is kept constant for the entire period, although we acknowledge that  
227 oceanographic conditions such as circulation, ventilation and extensive sea-ice cover  
228 during glacial periods may result in the overestimation of the actual calendar ages (Heaton  
229 et al., 2022; Pieńkowski et al., 2022). However, as our ages are from the early Holocene,  
230 they were likely little affected by such changes in oceanographic conditions. We argue  
231 that using only an additional local reservoir correction therefore provides an appropriate  
232 estimate for these ages, with relatively minimal uncertainties.

## 233 **4. Results**

### 234 **4.1. Seafloor geomorphology**

235 Complete multibeam bathymetric coverage of the Clyde fjord-cross-shelf trough  
236 system allowed identifying several submarine glacial landforms (Fig. 2 and 3) that are  
237 here described and interpreted for reconstructing the past configuration, stages and  
238 dynamics of the LIS margin.





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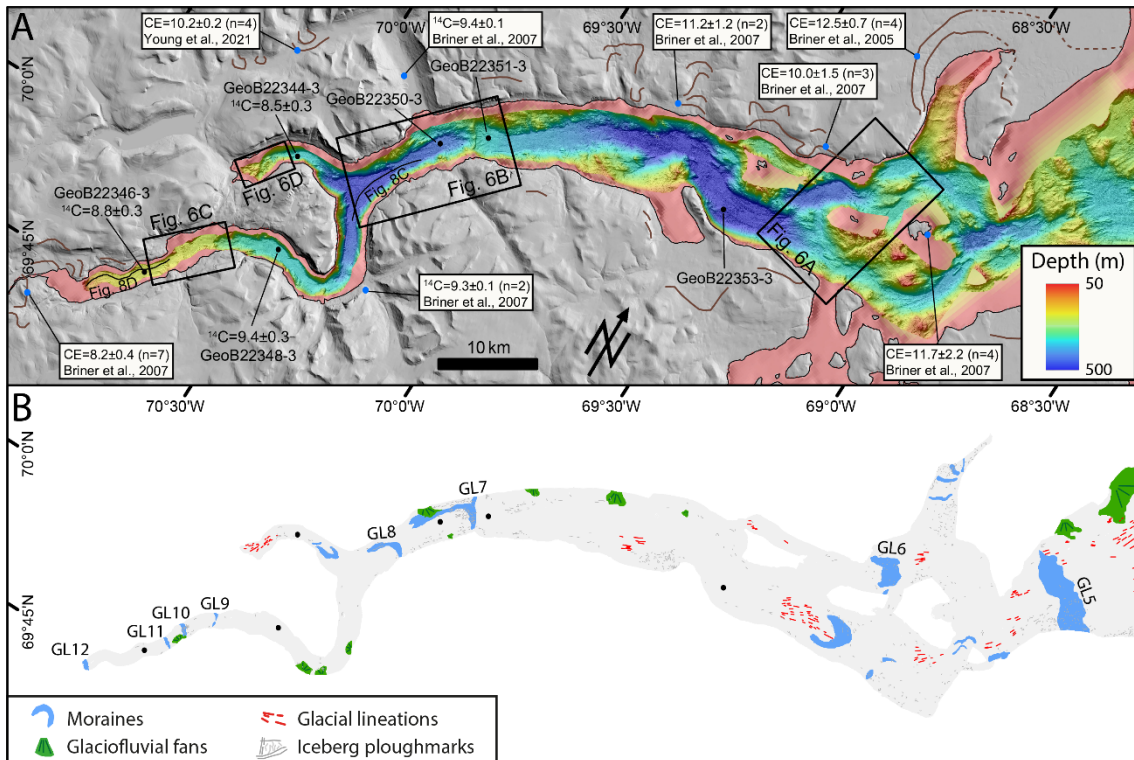
240 **Figure 2** (A) Multibeam bathymetry of Clyde Trough. Boxes show the location of Figs. 4 and 5. Dashed  
 241 brown lines represent moraine ridges from Briner et al. (2007). Ages derived from the sediment cores  
 242 represent minimum age for deglaciation at the core site. (B) The mapped distribution of submarine  
 243 landforms in the trough.

244 *4.1.1. Subglacial landforms*

245 Several streamlined landforms oriented along the fjord and trough axis occur in  
 246 the Clyde fjord-cross-shelf trough system and represent variations in ice-flow direction  
 247 and velocity. They are here grouped under the generic term glacial lineations.

248            *Glacial lineations* - Sets of streamlined and curvilinear ridges oriented parallel to  
249 the trough axis are observed at depths between 100 m and 500 m; they are up to 5 km-  
250 long (Fig. 4, 5A, 5C, 6A and 6D). These streamlined landforms are between 50 m to 500  
251 m-wide and can be as much as 60 m-high compared to the surrounding seafloor. These  
252 ridges are mostly aligned parallel to each others, but are in some cases divergent where  
253 the trough widens. Features with lower length to width ratios (between 1:5 and 1:20) tend  
254 to be asymmetrical, with a gentler seaward slope, and occur mostly beyond bedrock  
255 outcrops. Conversely, ridges with higher length to width ratios (up to 1:50) are more  
256 symmetrical, subtle and tend to have a smoother appearance. These ridges are in some  
257 cases superimposed on grounding line landforms.

258            These streamlined landforms are interpreted as glacial lineations, such as mega-  
259 scale glacial lineations (MSGs), drumlins and crag-and-tails, which provide evidence  
260 for former ice-flow directions as they are parallel to another (Clark, 1993; Stokes and  
261 Clark, 2001; Spagnolo et al., 2014; Dowdeswell et al., 2016; Maclean et al., 2016;  
262 Batchelor et al., 2018; Ottesen et al., 2022). Glacial lineations are differentiated on the  
263 basis of their width to length ratio and their formation processes. MSGs are generally  
264 highly elongated (up to 1:50 length to width ratio) and are produced by the deformation  
265 of soft till beneath a fast-flowing ice stream (e.g., Clark, 1993) or as a product of ice keels  
266 ploughing through sediments (e.g., Tulaczyk et al., 2001; Clark et al., 2003). In turn, crag-  
267 and-tails and drumlins are less elongated (between 1:5 and 1:20 length to width ratio) and  
268 formed by the accumulation of sediments on the seaward side of bedrock obstacles or the  
269 streamlining of bedrock by ice (Stokes and Clark, 2001; Ottesen et al., 2007). The position  
270 of most crag-and-tails beyond bedrock in Clyde Inlet suggests that they were formed by  
271 the streamlining action of warm-based ice (Dowdeswell et al., 2016; Maclean et al.,  
272 2016).



273

274 **Figure 3** (A) Bathymetry of Clyde Inlet. Boxes show the location of Fig. 6A-D. Brown lines represent  
 275 moraine ridges from Briner et al. (2007). Ages derived from the sediment cores represent minimum age for  
 276 deglaciation at the core site. Blue dots indicate cosmogenic exposure (CE) and radiocarbon ( $^{14}\text{C}$ ) dating  
 277 discussed in the text, compiled from previous studies along Clyde Inlet (in ka  $\pm$  1 SD uncertainties). (B)  
 278 The mapped distribution of submarine landforms in the fjord.

279 *4.1.2. Ice marginal landforms*

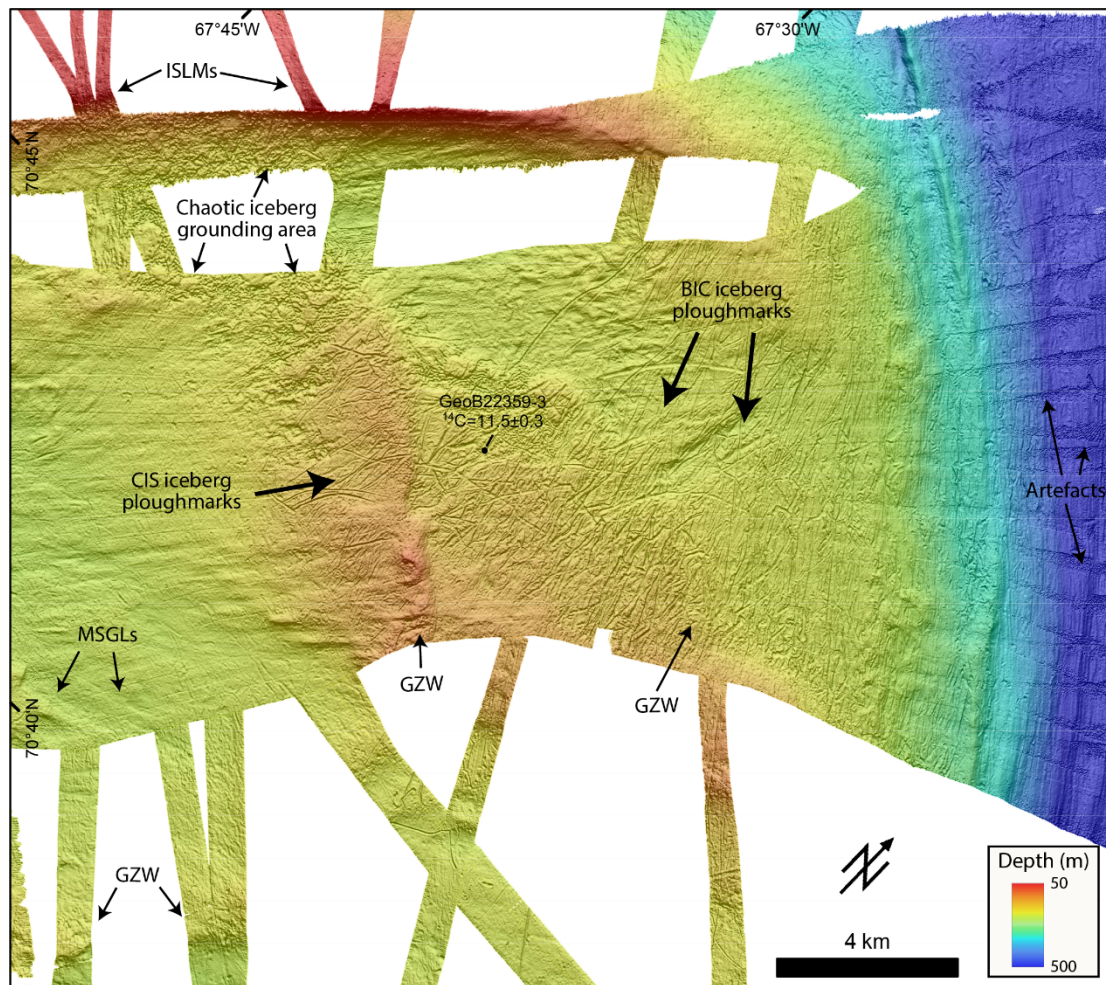
280 Several sediment wedges and ridges are observed on the multibeam bathymetry  
 281 imagery and record the extent and retreat pattern of the LIS during the LGM and the  
 282 deglaciation. GZWs, morainal bank, and end moraines represent major grounding line  
 283 landforms (referred to as GL in Figs. 2-3 and in the discussion) that record former,  
 284 successive position of the ice margin along the Clyde fjord-cross-shelf trough system.

285 *Grounding-zone wedges (GZWs)* - Asymmetric transverse wedges characterized  
 286 by steeper ice-distal slope are identified in the outer and middle part of Clyde Trough  
 287 (Fig. 4 and 7). These wedges are present at water depths ranging from 150 m to 350 m,  
 288 have lengths between 3 and 10 km and rise from 20 to 50 m above the surrounding  
 289 seafloor. On the seismic data, they correspond to wedge-like deposits reaching 15 to 20  
 290 m in thickness and 5 km in length (Fig. 7) with low-reflectivity, transparent and

291 sometimes chaotic acoustic signature. Smooth and gently sloping fan-shaped surfaces  
292 (between 0.5 and 2°) are usually identified on the seafloor in front of those large  
293 transverse wedges.

294         Due to their asymmetric geometry and steep ice-distal slopes, these asymmetric  
295 wedges are interpreted as grounding-zone wedges similar to ice-contact wedges reported  
296 on other deglaciated continental shelves (e.g., Ottesen et al., 2007, 2022; Slabon et al.,  
297 2016; Brouard and Lajeunesse, 2017). Large GZWs are formed during relatively long-  
298 term ice margin stabilizations (decadal- to centennial-scale) which enable subglacial  
299 sediment accumulation at the grounding line (Powell and Domack, 1995; Dowdeswell  
300 and Fugelli, 2012; Batchelor and Dowdeswell, 2015). The accumulation of sediment is  
301 interpreted to be vertically limited by the presence of a floating ice shelf seaward from  
302 the grounded ice, thus favoring horizontal progradation of sediments and forming low-  
303 amplitude extensive wedges (Dowdeswell and Fugelli, 2012 ; Batchelor and Dowdeswell,  
304 2015). The fan-shaped surfaces are interpreted as ice-marginal debris flow lobes  
305 deposited by sediment delivery at the grounding line of a marine terminating glacier  
306 (Syvitski and Shaw, 1995; Ó Cofaigh and Dowdeswell, 2001; Bjarnadóttir et al., 2012;  
307 Batchelor and Dowdeswell, 2015; Dowdeswell et al., 2015; Ottesen et al., 2017).





308

309 **Figure 4** Multibeam bathymetry of submarine landforms in the outer trough. CIS – Clyde Ice Stream and  
 310 BIC – Baffin Island Current. For further interpretation and discussion, see Couette et al. (2022).

311 *Morainal banks* – A large asymmetric wedge with a steeper ice-distal side and  
 312 overprinted by transverse zigzag-shaped ridges and recessional moraines is located in the  
 313 center of Clyde Trough at a depth of 325 m (Fig. 5A and 7B). The wedge is 75 m-high, 6  
 314 km-long and 12 km-wide, while the overprinting zigzag-shaped ridges are 5-10 m-high,  
 315 100 to 200 m-long and between 500 and 1000 m-wide (Fig. 5A). The seismic data show  
 316 three distinct morpho-sedimentary units composing the underlying sediment body: (1) a  
 317 bottom unit showing evidence of glaciotectonism including folds and thrust sheets; (2) an  
 318 upper unit showing seaward dipping reflectors; and 3) ridges with a chaotic acoustic  
 319 signature located on the stoss side of the system. Gently sloping fan-shaped surfaces are  
 320 identified in front of this large asymmetric wedge. The associated deposits attenuate  
 321 underlying landforms such as glacial lineations and recessional moraines (Fig. 5A).

322           Based on its morphology and the overprinted zigzag-shaped ridges, this landform  
323 is interpreted as a morainal bank. Morainal banks are generally associated with a quasi-  
324 stagnant ice margin position during overall deglaciation, coherent with the presence of  
325 recessional moraines (Powell, 1981; Christofferson and Tulaczyk, 2003; Laberg et al.,  
326 2009; Dowdeswell et al., 2015). Zigzag-shaped ridges are, in turn, associated with  
327 sediment deformation related to push and thrust of ice proximal deposits by glacier  
328 readvances (Powell, 1981; Christofferson and Tulaczyk, 2003; Laberg et al., 2009). They  
329 could also represent crevasse-squeeze ridges, produced subglacially by basal till being  
330 squeezed into crevasses at the glacier bed during an ice advance (Evan and Rea, 1999;  
331 Ottesen et al., 2022). The complex acoustic signature with folds and thrust sheets  
332 observed on the seismic profile (Fig. 7B) further support the interpretation of a readvance  
333 of the ice margin that lead to the formation of the morainal bank (Laberg et al., 2009).  
334 The fan-shaped surfaces observed in front of the morainal bank are interpreted as ice-  
335 marginal debris flow lobes (Syvitski and Shaw, 1995; Ottesen et al., 2017).

336           *Major end moraines* - Arcuate asymmetrical ridges with a steeper ice-proximal  
337 slope are observed at several locations along Clyde Inlet (Figs. 6B-C). These landforms  
338 range between 10 m and 150 m in height, and between 1 km and 6 km in width. They are  
339 500 m to 3 km-long and are observed at depths reaching 450 m. On the distal flank, fan-  
340 shaped surfaces occasionally emanate from the ridges over distances of several  
341 kilometers. Ridges located in Patricia Bay have a more subtle and smoother appearance  
342 (Fig. 3). Partially buried transverse ridges with steep-scarps are also observed at a few  
343 locations in the fjord, where they bound step-like basins (Figs. 6B-C). Several linear  
344 ridges running parallel to the fjord axis and specifically located on the flanks of the fjord  
345 are also observed in Clyde Inlet (Fig. 6B). These ridges are 500 m to 3 km-long, 100 m

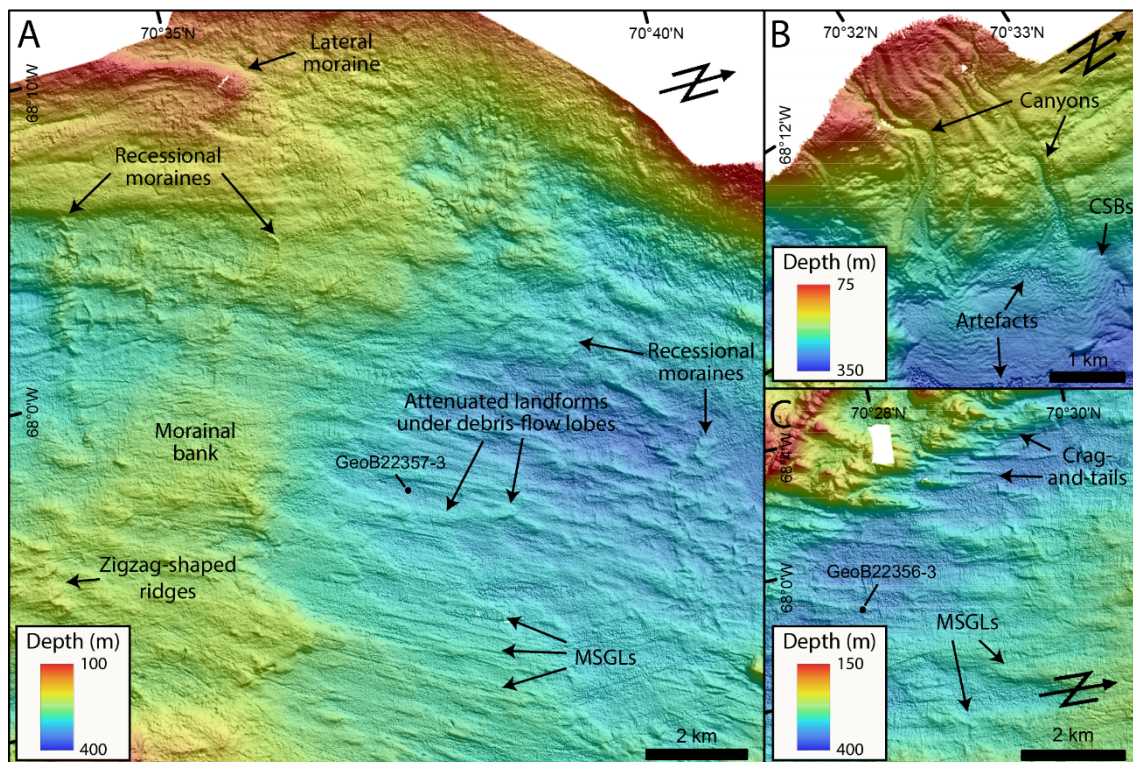
346 to 500 m-wide, up to 100 m-high and appear to be, in some cases, the lateral extension of  
347 the arcuate ridges (Figs. 5A and 6B).

348 Due to their geometry and their position transverse to the fjord axis, the arcuate  
349 ridges are interpreted as major end moraines, whereas the linear and parallel-to-the-fjord  
350 morphologies are, in turn, interpreted as lateral moraines (Dowdeswell and Vásquez,  
351 2013; Dowdeswell et al., 2014; Hodgson et al., 2014). Major end moraines are formed by  
352 the deposition and pushing of sediment during long-term ice margin stabilizations of at  
353 least decades to centuries (Powell and Alley, 1997; Dowdeswell et al., 2014; Batchelor et  
354 al., 2018). In contrast to GZWs, moraine ridges usually develop at the margin of tidewater  
355 glacier that are vertically unrestricted accommodation space at the grounding line (Powell  
356 and Alley, 1997). Similarly to observations in front of GZWs and the morainal bank, the  
357 fan-shaped surfaces are inferred to be ice-marginal debris flow lobes (Syvitski and Shaw,  
358 1995; Ottesen et al., 2017). The moraine ridges in Patricia Bay probably represent  
359 deposition by a secondary ice flow across the bay and onto the Clyde forelands (Briner et  
360 al., 2005). The step-like basins observed in the fjord are interpreted to be the result of  
361 high sediment deposition from transverse deltas fed by nearby rivers and overflowing some  
362 of the basins confined by a succession of receding moraines (Hodgson et al., 2014;  
363 Brouard and Lajeunesse, 2019a).

364 *Recessionnal moraines* - Small ridges transverse to the former ice flow are  
365 specifically identified on the multibeam bathymetric data of the trough (Fig. 5A). They  
366 are observed in the mid- to inner section of the trough in water depths ranging from 200  
367 m to 300 m; their size varies between 2 m and 12 m-high and 100 m to 500 m-wide. Some  
368 of them can be traced for lengths of >5 km, although the majority do not exceed 1 km.  
369 Most of the ridges are arcuate and symmetric. They occur in two distinct clusters of  
370 parallel to sub-parallel features spaced a few hundred meters apart.



371            These small parallel ridges are interpreted as recessional moraines, formed by the  
 372 delivery and pushing of subglacial sediments by minor stillstands or readvances of the  
 373 grounded glacier front in subaqueous conditions (Boulton, 1986; Lindén and Möller,  
 374 2005; Dowdeswell et al., 2008; Ó Cofaigh et al., 2008; Arndt et al., 2017; Batchelor et  
 375 al., 2018; Howe et al., 2019; Ottesen et al., 2022). The presence of recessional moraines  
 376 in marine environments is usually associated with a relatively slow retreat of the ice  
 377 margin in water typically shallower than 350 m (Lindén and Möller, 2005; Dowdeswell  
 378 et al., 2008; Ó Cofaigh et al., 2008; Howe et al., 2019).

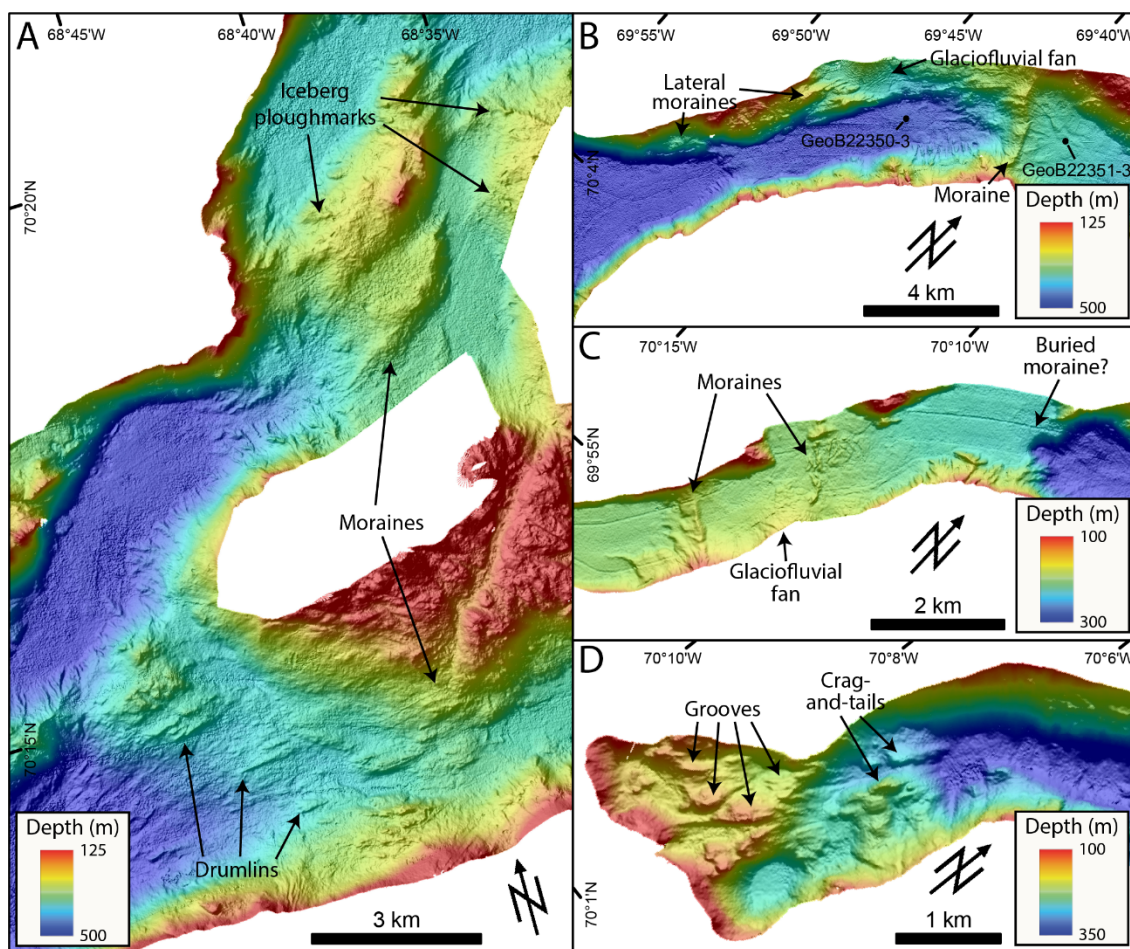


379  
 380 **Figure 5** (A) Multibeam bathymetry of submarine landforms in the middle trough. (B) Example of a  
 381 glaciofluvial fan, with associated canyons and CSBs, in the middle trough. (C) Example of mega-scale  
 382 glacial lineations (MSGLs) in the inner trough.

383            *Ice stream lateral moraines* - Elongated ridges orientated parallel to the former  
 384 ice flow are observed at the lateral boundaries of Clyde Trough (Fig. 4). They are up to  
 385 25 m-high, 1 to 3 km-wide and occur at depths between 50 m to 175 m. These ridges have  
 386 a steeper trough-proximal slope and are observed on both sides of the trough.



387            These extensive elongated ridges are interpreted as ice-stream lateral moraines  
 388 based on their shape, dimensions and location at the margins of the trough (Ottesen et al.,  
 389 2007; Rydningen et al., 2013; Batchelor and Dowdeswell, 2016; Brouard and Lajeunesse,  
 390 2019b). They have been suggested to be formed mainly from the accumulation of  
 391 subglacial till at the shear zone between fast-flowing ice stream and slow-flowing  
 392 portions of an ice sheet or ice free terrain (Dyke and Morris, 1988; Stokes and Clark,  
 393 1999, 2002; Batchelor and Dowdeswell, 2016).



394  
 395 **Figure 6** (A) Multibeam bathymetry of submarine landforms in the outer fjord. (B) Multibeam bathymetry  
 396 of submarine landforms in the middle fjord. (C) Example of glaciofluvial fan and moraines in the inner  
 397 fjord. (D) Example of the crag-and-tails and grooves in Cormack Arm.

398 *4.1.3. Proglacial landforms*

399            Two other sets of landforms are identified in the Clyde fjord-cross shelf  
 400 system and characterise the glacial dynamics in front and beyond the receding margin of  
 401 the ice sheet: iceberg ploughmarks and glaciofluvial fans.

402           *Iceberg ploughmarks* - Straight and sinuous V- or U-shaped furrows are observed  
403 at varying depths along the fjord-cross-shelf trough system (Figs. 4, 5A and 6A). They  
404 are more common on the continental shelf, whereas in the fjord they mostly occur on the  
405 sill. They are usually associated with berms or levees on either side and often cross-cut  
406 each other. They are 2 m to 10 m-deep, up to 200 m-wide and in some cases can be traced  
407 for >5 km. Two predominant orientations can be distinguished for the larger furrows in  
408 the outer trough area: 1) near the shelf break, iceberg ploughmarks superimpose the  
409 outermost GZW, are observed at depths ranging between 175 and 225 m and have a NNW  
410 to SSE orientation; 2) on the stoss side of the second GZW, ploughmarks are observed at  
411 depths between 150 and 200 m and have a nearly orthogonal (SW-NE) along-trough  
412 orientation (Fig. 4). Some large zones of chaotic scouring patterns are also identified in  
413 the outer trough, regardless of shallower surrounding seafloor area (Fig. 4). They are  
414 between 1 to 5 km-wide and present both depressions and mound-like morphologies with  
415 irregular furrows. The depressions are generally between 2 to 5 m-deep, while the mounds  
416 are <2 m-high. The SE flank of the trough is also densely incised by randomly oriented  
417 semi-circular pockmarks (Fig. 2A). They are generally <5 m-deep and rarely exceed 100  
418 m-wide.

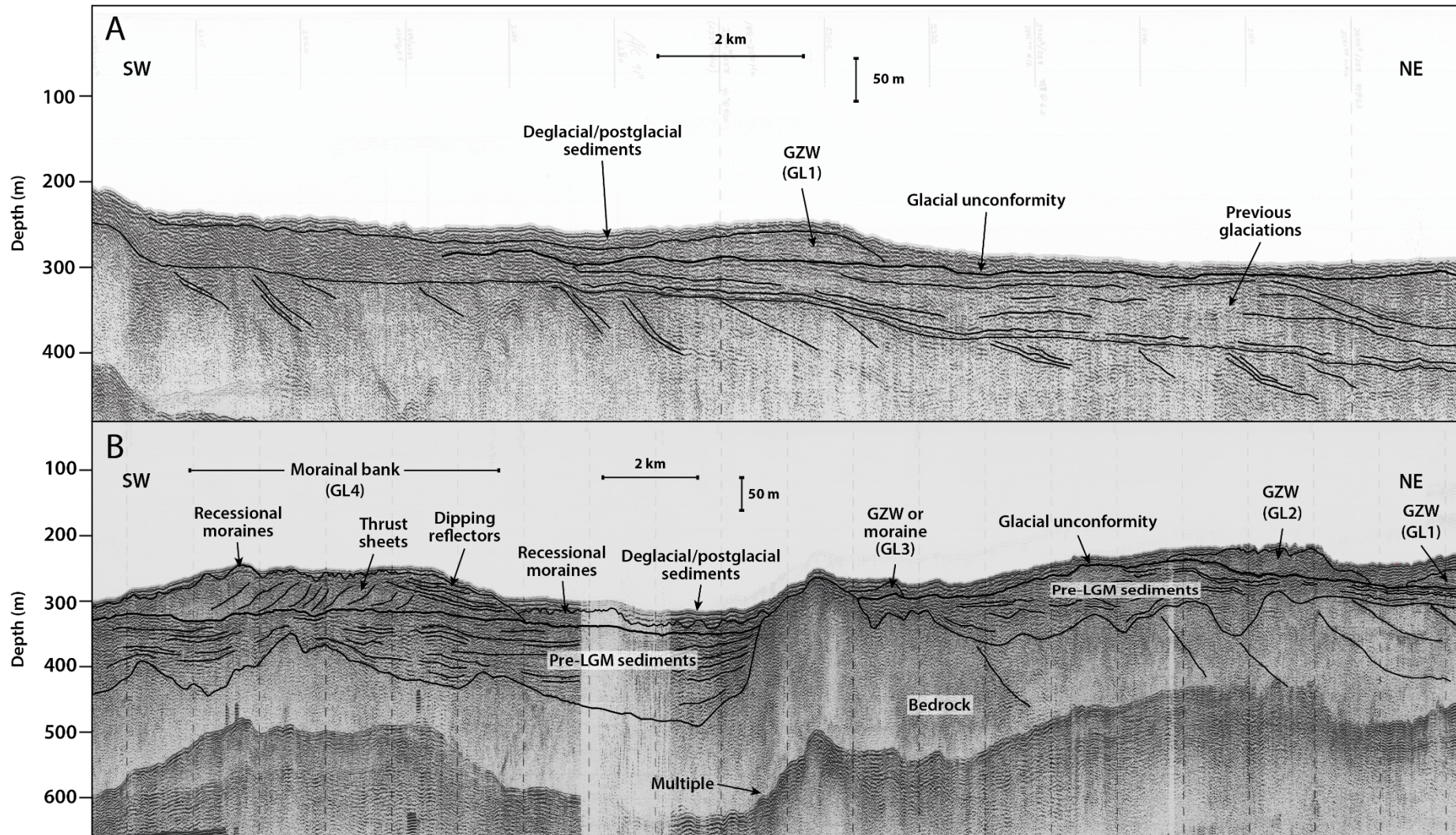
419           Due to their straight and sinuous character, V- and U-shaped cross-profiles and  
420 chaotic patterns, these furrows are interpreted as ploughmarks – also termed iceberg  
421 scours – from iceberg keels associated with calving outlet glaciers during deglaciation  
422 (King, 1976; Vorren et al., 1989; Jakobsson et al., 2011; Dowdeswell et al., 2014; Lewis  
423 et al., 2016; Brouard and Lajeunesse, 2019a, 2019b). The SSE oriented ploughmarks  
424 suggest influence of the Baffin Island Current (BIC), which delivered large icebergs from  
425 northern Baffin Bay in the early phases of deglaciation (Andrews et al., 1998; Jennings  
426 et al., 2011), whereas the NE orientated iceberg ploughmarks is probably related to ice

427 calving from the Clyde Ice Stream (CIS) margin. The larger chaotic zones possibly  
428 correspond to large iceberg (>5 km<sup>2</sup>) grounding areas and may have been produced by  
429 the ploughing of the seafloor by the keel of larger ice masses that were stuck for long  
430 periods of time. Tides and meltwater currents may have caused these iceberg shifts,  
431 creating the chaotic pattern. To our knowledge, such features have rarely been identified  
432 on continental margins and the genesis presented here remains speculative. The semi-  
433 circular pockmarks are interpreted as pits formed by the short-term grounding of icebergs  
434 that were semi-buoyant or that turned over, as described by Woodworth-Lynas et al.  
435 (1991).

436 *Glaciofluvial fans* - Several smooth and gently sloping fan-shaped morphologies  
437 are observed on the flanks of the fjord-cross-shelf trough system, at the mouth of tributary  
438 valleys (Fig. 5B, 6B-C). They are generally 1 to 5 km-long and observed at depths  
439 between 200 m and 400 m. On their upper reaches, they are dissected by series of parallel  
440 15 m-deep channels. On their lower reaches, they are superimposed by transverse 5 m-  
441 high and 200 m to 1 km-wide curvilinear ridges.

442 These fan-shaped landforms are interpreted as glaciofluvial fans formed by  
443 sediment delivery by glacial meltwaters from adjacent alpine valleys (Powell, 1990;  
444 Normandeau et al., 2019). Similar landforms were previously identified from several  
445 glaciated locations (Powell, 1990; Lønne, 1995; Dowdeswell and Vásquez, 2013;  
446 Normandeau et al., 2019; Brouard and Lajeunesse, 2019a, 2019b). Parallel channels are  
447 more common on glacially-fed fans suggesting higher turbidity currents activity used for  
448 delivering sediments downslope (Dowdeswell and Vásquez, 2013; Batchelor et al., 2018;  
449 Normandeau et al., 2019). Transverse curvilinear ridges observed on foreset beds are  
450 crescent-shaped landforms formed by turbidity currents (Normandeau et al., 2019).





451  
452  
453

**Figure 7** (A) Airgun profile 78029\_AG\_275\_0130 in the outer Clyde Trough showing GL1 and deeply buried GDFs. (B) Airgun profile 80028\_AG\_RAYT\_257\_0200 along Clyde Trough showing the sedimentary assemblages. The depth is based on two-way travel time (TWTT) of 1500 m s<sup>-1</sup>.

## 454 **4.2. Acoustic stratigraphy**

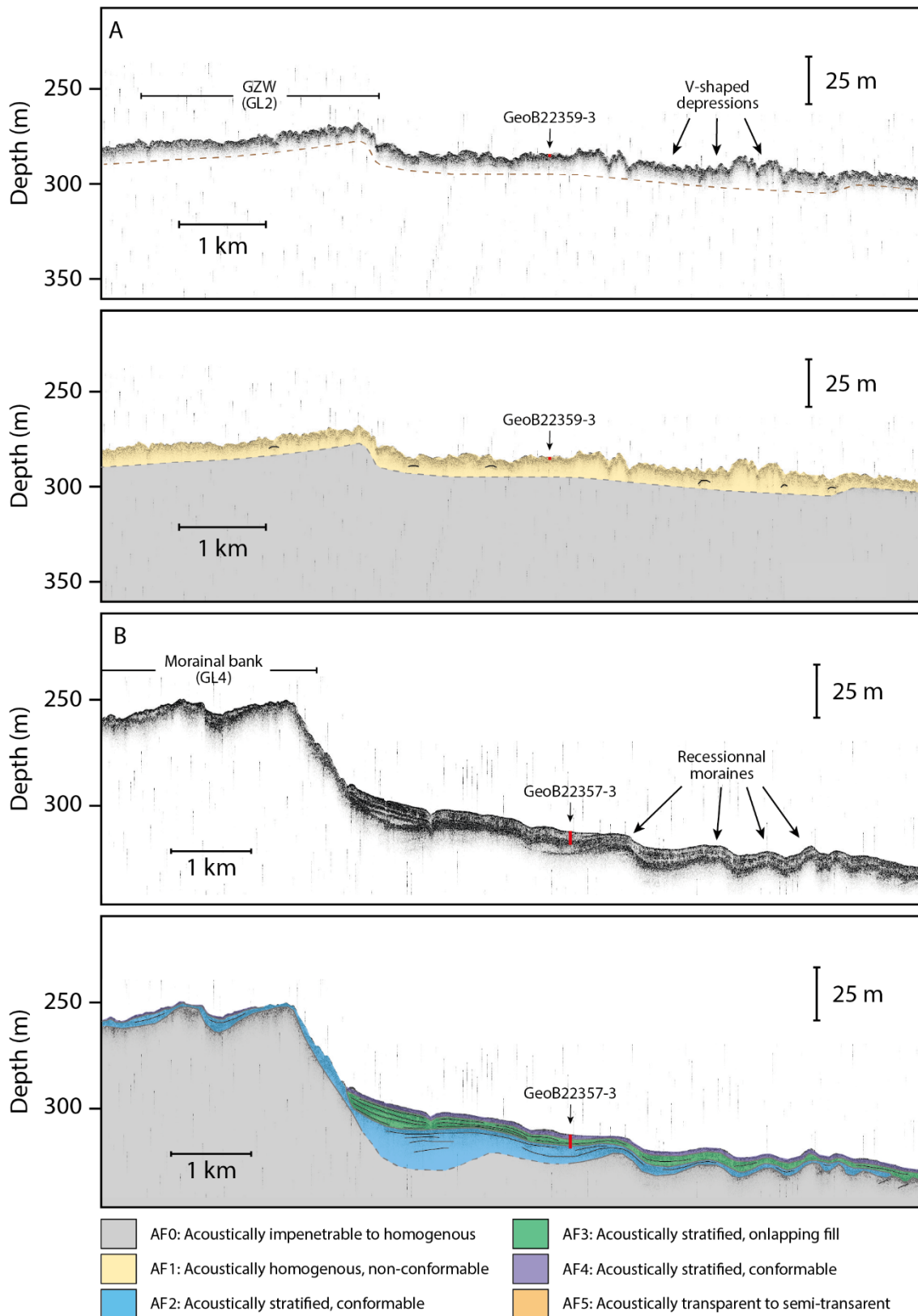
455 Six acoustic facies (AF) were distinguished from >2,500 km of Parasound profiles  
456 in the Clyde fjord-cross-shelf trough system. They were differentiated on the basis of their  
457 acoustic signatures, bounding reflectors and internal geometries. Additionally, airgun  
458 profiles allow identifying the general stratigraphy in the middle to outer trough and in  
459 places the limit between the seaward-dipping strata of the bedrock and Quaternary  
460 sediments (Fig. 7). They also allow delineating an erosional surface, which represent an  
461 unconformity in the sedimentation sequence. This surface is overlain by an acoustically  
462 chaotic to semi-transparent unit, which is in turn overlain by a thin unit (<10 m) of high  
463 amplitude parallel reflectors. The erosional surface and overlying sediments identified on  
464 airgun data likely represent the complete glaciation/deglaciation cycle, with an ice sheet  
465 advance across the continental shelf during the LGM followed by deglacial to postglacial  
466 sedimentation. The Parasound profiles however allow for a more in-depth analysis of the  
467 different facies represented along the system during deglaciation.

### 468 *4.2.1. AF0: Acoustically impenetrable to homogenous facies*

469 AF0 is an impenetrable and homogenous acoustic facies and forms the acoustic  
470 basement in most profiles (Figs. 8A-D). This facies is not visible in some basins where  
471 the attenuation of the acoustic signal in thick sediment prevents penetration. It is  
472 internally structureless and is characterised by a weak, high-amplitude rugged upper  
473 reflector; in most cases it shows an irregular to hyperbolic geometry.

### 474 *4.2.2. AF1: Acoustically homogenous, non-conformable facies*

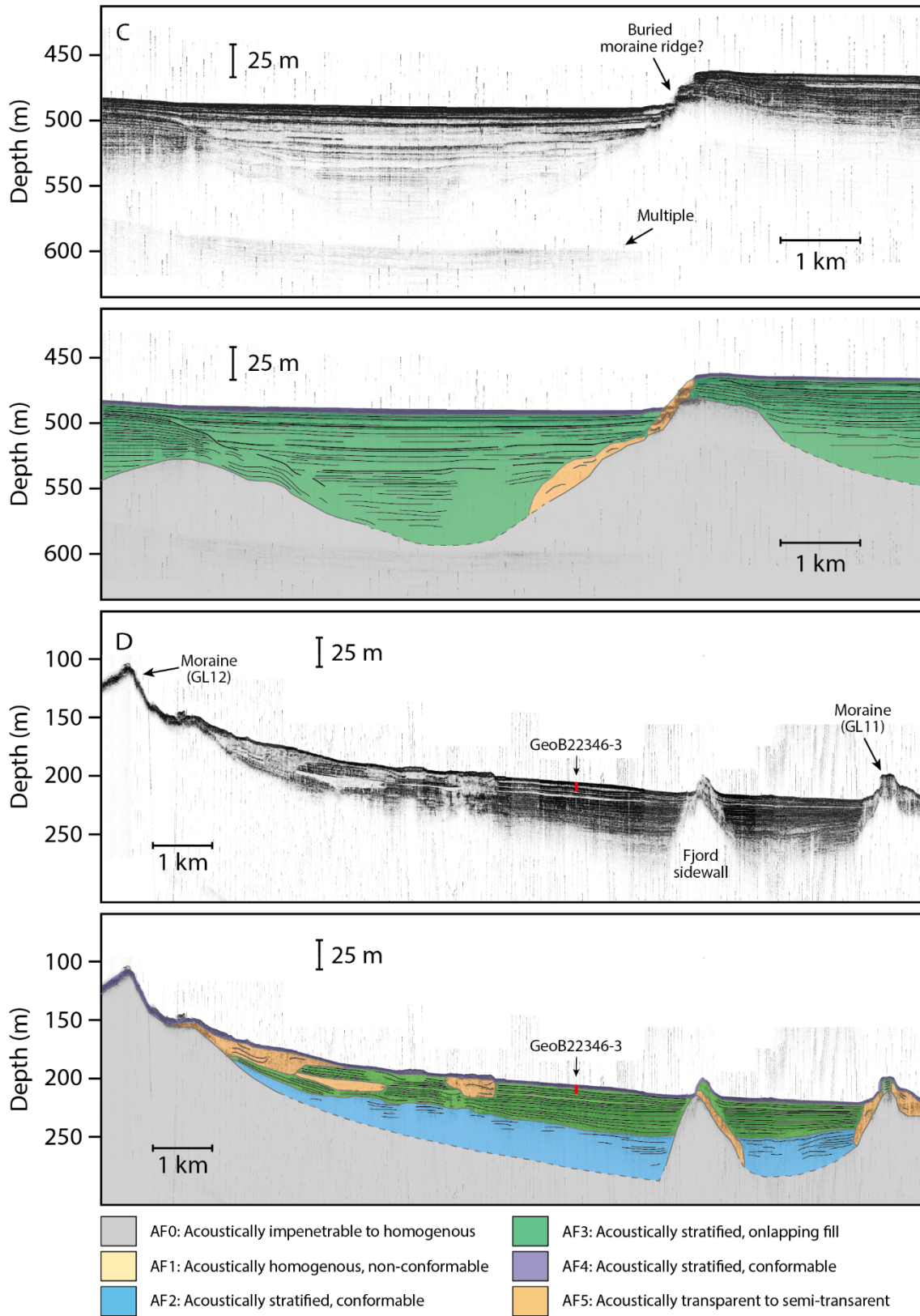
475 AF1 rarely exceeds 5 m thickness and is restricted to the outer trough (Fig. 8A).  
476 It is characterised by chaotic and discontinuous acoustic reflectors with occasional  
477 hyperbolae near the surface. This facies is acoustically homogenous and exhibits weak  
478 lower reflectors. It shows in-filled small-scale (~5 m-deep, ~100 m-wide) V-shaped  
479 depressions and a non-conformable configuration.



480

481 **Figure 8** Parasound profiles along the Clyde Inlet fjord-cross-shelf trough system showing the different  
 482 acoustic facies (Table 1). See Figs. 2 and 3 for location of the profiles. Color coding for the different  
 483 acoustic facies is the same as the one used for the lithological facies in Fig. 9.





484

485

Figure 8 Continued.

486 4.2.3. *AF2: Acoustically stratified, conformable facies*

487 AF2 is an acoustically stratified and conformable facies with a sharp upper  
488 reflector and low to medium amplitude parallel to sub-parallel irregularly-spaced internal  
489 reflectors (Figs. 8B-D). AF2 drapes the underlying unit and in sometimes exhibits discrete  
490 wedge-shape geometries at its lateral boundaries. It is observed over distances of several  
491 kilometers in the deeper parts of basins located within the fjord, but is also present at  
492 some locations on the inner trough and possibly on the upper slope of the shelf edge (Fig.  
493 7B). AF2 is generally between 5 and 20 m-thick, but reaches >30 m in the inner fjord.  
494 This facies is usually observed immediately seaward from grounding line landforms (i.e.,  
495 GZWs, morainal bank and moraines), although their connection remains elusive.

496 4.2.4. *AF3: Acoustically stratified, onlapping fill facies*

497 AF3 is an acoustically stratified facies with parallel medium to high amplitude  
498 reflectors (Figs. 8B-D). It has an onlapping or ponded basin-fill configuration and is  
499 interbedded with occasional thicker transparent units. This facies is usually between 10  
500 and 40 m-thick, but can reach >75 m in the fjord. Acoustically chaotic and transparent  
501 lenticular sediment bodies are in many cases observed within AF3.

502 4.2.5. *AF4: Acoustically stratified, conformable facies*

503 AF4 is an acoustically stratified facies with parallel, closely spaced high  
504 amplitude and opaque reflectors (Figs. 8B-D). It forms a conformable drape of <5 m on  
505 the underlying AF3. It characterizes the uppermost sediment bodies deposited in the  
506 Clyde fjord-cross-shelf trough system.

507 4.2.6. *AF5: Acoustically transparent to semi-transparent facies*

508 AF5 is an acoustically transparent to semi-transparent chaotic facies with rare or  
509 poorly defined internal reflectors (Figs. 8C-D). Sediment bodies showing AF5 usually  
510 have an erosive base, an hummocky surface and are either lenticular or taper on slopes.



511 It is often observed interfingering within the stratified AF2 or AF3. It is generally a few  
512 meters-thick, but can exceed 10 m-thick locally. AF5 is mostly observed inside the fjord,  
513 near slopes and tributary valleys.

#### 514 **4.3. Lithological facies**

515 Glacial lithological facies (LF) were identified from the gravity cores collected in  
516 the Clyde Inlet fjord-cross-shelf trough system by combining visual core descriptions,  
517 photographs of split cores and XRF data. Five facies and three subfacies were determined  
518 based on colour, texture, sedimentary structures (e.g., lamination, bioturbation) and Ca/Ti  
519 ratio changes (Fig. 9). Ratios of calcium (Ca) to titanium (Ti) were used for correlation  
520 between the cores along the Clyde Inlet transect.

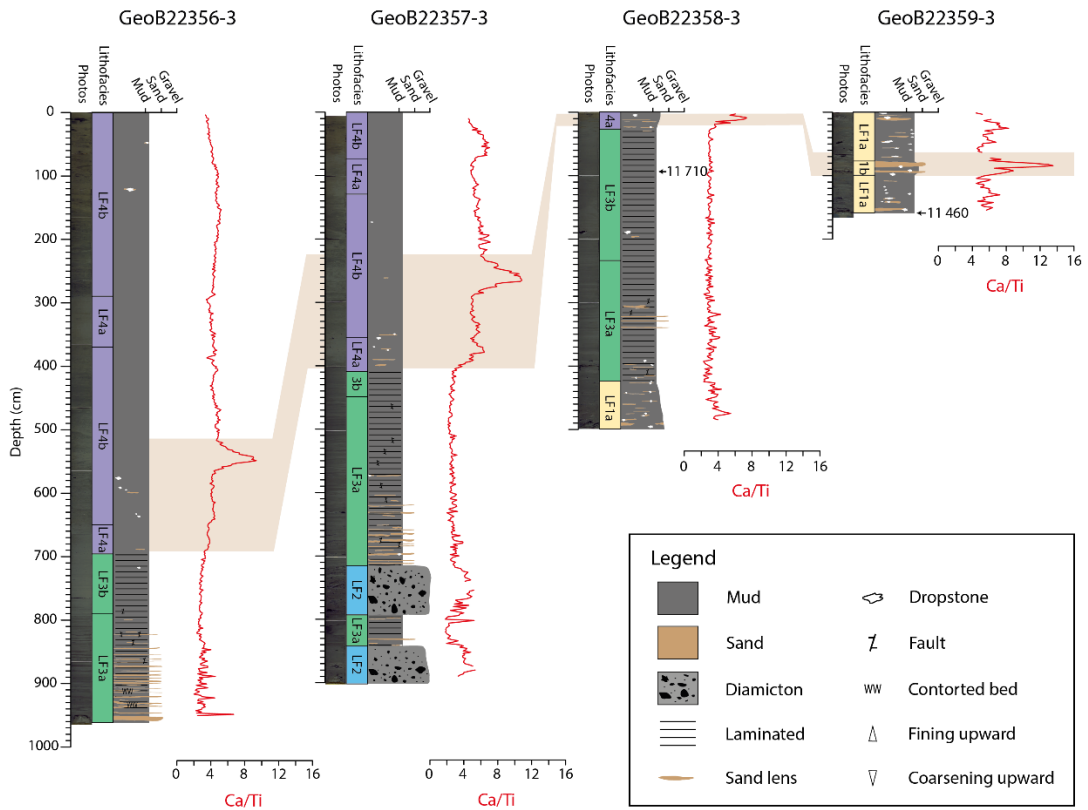
##### 521 *4.3.1. LF1: Unstratified sandy mud with dropstones*

522 LF1 consists of unstratified olive dark gray sandy mud with dropstones and  
523 dispersed pebble-sized clasts. This lithofacies shows no or few traces of bioturbation and  
524 occurs in the cores of the outer trough. LF1a consists of a gray coarse sandy mud, IRD-  
525 rich facies with occasional sand lenses. LF1b has a more reddish brown color with a finer  
526 sandy mud matrix and fewer apparent clast. LF1b is only found in core GeoB22359-3 and  
527 is characterized by a high peak in Ca/Ti ratio.

##### 528 *4.3.2. LF2: Poorly sorted diamicton*

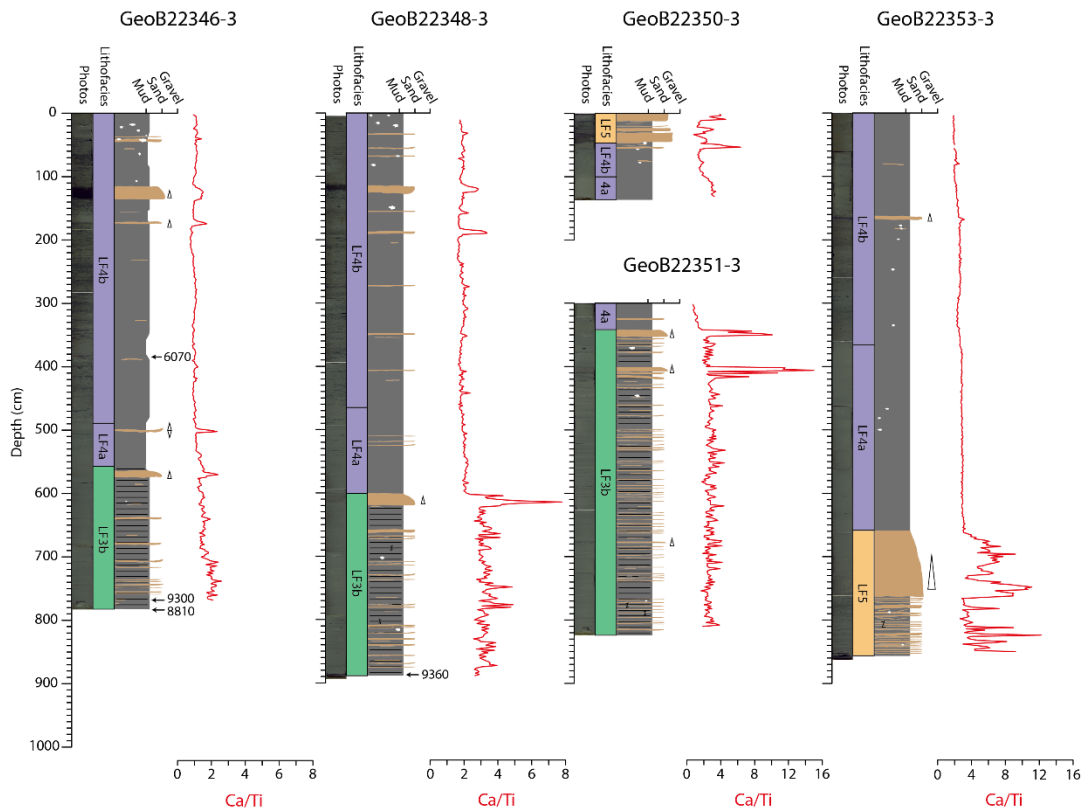
529 LF2 only occurs at the base of core GeoB22357-3 and consists of dark gray  
530 weakly stratified, poorly sorted muddy-sand matrix diamicton. It is characterized by thick  
531 horizons (50-75 cm) of abundant sub-angular to sub-rounded clasts ranging from granules  
532 to pebbles, resulting in relatively higher and more chaotic peaks of Ca/Ti ratio. Diamicton  
533 beds of LF2 have conformable bedding contact and are interbedded with massive to  
534 laminated and faulted silty mud beds of LF3.

### Clyde Trough



535

### Clyde Inlet



536

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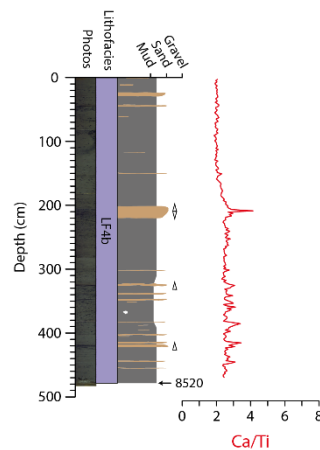
539

540

**Figure 9** Simplified lithological logs, Ca/Ti ratios and calibrated radiocarbon ages of sediment cores collected along the Clyde Inlet fjord-cross-shelf trough system. Note that Ca/Ti ratios have a different scale for GeoB222344-3, GeoB222346-3 and GeoB222348-3. Color coding for the different lithological facies is the same as the one used for the acoustic facies in Fig. 8.

## Cormack Arm

GeoB22344-3



541

542

**Figure 9** Continued.

### 543 *4.3.3. LF3: Laminated mud with sandy layers*

544 LF3 is a laminated gray to dark gray silty mud characterised by a lack of  
545 bioturbation and the presence of scattered angular to sub-angular clasts. It also contains  
546 individual thin sand laminae that has sharp basal contacts. The sand layers are coarser and  
547 thicker at the base of the unit, defining an overall fining upward trend. LF3a is defined  
548 by the presence of faulting, as well as more irregular and prominent laminae. LF3b shows  
549 no faulting but contains faint thinner beds of coarser sediments. Curves of Ca/Ti ratios  
550 from LF3b are chaotic with numerous peaks generally corresponding to coarser laminae.  
551 The laminae become less prominent and more spaced upward at the transition from LF3a  
552 to LF3b.

### 553 *4.3.4. LF4: Massive bioturbated silty mud*

554 LF4 consists of massive olive gray, bioturbated silty mud and occur at the top of  
555 most cores, except GeoB22350-3 and GeoB22359-3. This lithofacies generally contains  
556 black mottles and few clasts. The transition with the underlying unit is gradational; LF4a  
557 is weakly bioturbated, while LF4b contains high concentrations of bioturbation and black  
558 mottles. Ca/Ti ratios in the fjords are relatively low with distinct peaks representing sandy

559 layers. In the trough, these ratios are higher and more irregular with peaks representing  
560 longer events or of bigger magnitude.

#### 561 *4.3.5. LF5: Unstratified sand*

562 LF5 is characterized by unstratified gray coarse sand with lenses of olive gray  
563 silty mud and occur only at the top of core GeoB22350-3 and at the base of core  
564 GeoB22353-3. This unit rest on LF4 and shows traces of bioturbation in the silty mud  
565 lenses. Ratios of Ca/Ti in LF5 are characterized by chaotic and irregular peaks.

### 566 **4.4. Correlation of acoustic stratigraphy and lithological facies**

567 Combining the acoustic stratigraphy and lithological facies allows drawing the  
568 sedimentary architecture of the Clyde fjord-cross-shelf trough system. Table 2 presents a  
569 summary of the acoustic stratigraphy and lithofacies descriptions in their context of unit  
570 correlations.

#### 571 *4.4.1. Unit 0: Bedrock or ice-contact sediments*

572 Based on its stratigraphic position, acoustic appearance and rugged upper  
573 reflector, Unit 0 represents bedrock and/or ice-contact sediments (Syvitski and Shaw,  
574 1995; Streuff et al., 2018). Due to the absence of penetration in coarse sediments, it is  
575 usually difficult to differentiate between the two types on acoustic profiles (Streuff et al.,  
576 2018; Hogan et al., 2020). Nevertheless, Mesozoic to Tertiary sedimentary bedrock is  
577 recognizable on airgun profiles by seaward-dipping strata, whereas ice-contact deposits  
578 have a chaotic internal signature (Fig. 7). These profiles also reveal an unconformity that  
579 lies above >50 m of undifferentiated Quaternary sediments.

#### 580 *4.4.2. Unit 1: Iceberg-influenced sedimentation*

581 Unit 1, comprising AF1 and LF1, is interpreted as iceberg-influenced  
582 sedimentation, with ploughing and subsequent infilling by ice rafted sediments  
583 (Woodworth-Lynas et al., 1991; Arosio et al., 2018; Callard et al., 2018; Streuff et al.,

584 2018; Olsen et al., 2022). Hyperbolic signals present in this facies are characteristic of  
585 point-source diffractions from dispersed cobbles and boulders (Arosio et al., 2018;  
586 Callard et al., 2018). The unstratified structure and high pebble-sized clasts content  
587 suggest that iceberg calving was an important sediment source (Dowdeswell et al., 1994,  
588 2000; Ó Cofaigh and Dowdeswell, 2001; Ó Cofaigh et al., 2013a; Hogan et al., 2016;  
589 Sheldon et al., 2016). Its occurrence in heavily iceberg-disturbed area, as shown by  
590 multibeam bathymetry (Fig. 4), is consistent with this interpretation. The reddish brown  
591 color and the high Ca/Ti ratio of LF1b might correspond to a predominantly detrital  
592 carbonate input from northern Baffin Bay – possibly BBDC-0 (Simon et al., 2014;  
593 Jackson et al., 2017; Jenner et al., 2018; Lévesque et al., 2020).

#### 594 4.4.3. Unit 2: *Glacigenic debris-flows*

595 Based on its conformable geometry, stratified acoustic signature and sandy  
596 matrix-supported diamicton facies, AF2 and LF2 are interpreted as glacigenic debris-  
597 flows likely sourced during stillstands of the ice margin (Powell, 1990; Powell and Alley,  
598 1997; King et al., 1998; Ó Cofaigh and Dowdeswell, 2001; Flink and Noormets, 2018).  
599 The alternating nature of LF2 with the laminated/faulted mud of LF3 is consistent with  
600 an origin of distinct pulses of glacigenic debris-flows interbedded with meltwater plumes  
601 (Ó Cofaigh et al., 2013a; Callard et al., 2018; Jenner et al., 2018; Prothro et al., 2018).  
602 High sand and gravel content often indicate an ice-proximal sedimentation located within  
603 a few kilometers of the grounding line, which requires a stable ice margin over a period  
604 of years to decades in case of a thick accumulation (Ó Cofaigh et al., 2008; Dowdeswell  
605 et al., 2015; Callard et al., 2018; Prothro et al., 2018). Comparable facies have commonly  
606 been identified near glacier-influenced submarine fans and are inferred to be related to  
607 the remobilization of glacigenic debris at the grounding line by abundant subglacial  
608 meltwater discharge (Syvitski, 1991; King et al., 1998; Ó Cofaigh et al., 2013b;

609 Dowdeswell et al., 2015). The occurrence of such facies at a relatively short distance (<2  
610 km) from former major grounding line positions in the Clyde fjord-cross-shelf trough  
611 system therefore support this interpretation.

#### 612 4.4.4. Unit 3: Ice-proximal glaciomarine sedimentation

613 Unit 3, comprising AF3 and LF3, is interpreted as stratified ice-proximal  
614 glaciomarine sedimentation (Ó Cofaigh and Dowdeswell, 2001; Hodgson et al., 2014;  
615 Normandeau et al., 2017; Brouard and Lajeunesse, 2019a; Trottier et al., 2020; Olsen et  
616 al., 2022). The well-preserved laminations suggest fallout sediment deposition from  
617 meltwater plumes (Syvitski, 1991; Ó Cofaigh and Dowdeswell, 2001; Sheldon et al.,  
618 2016; Jenner et al., 2018; Callard et al. 2020; Olsen et al., 2022). Laminae probably  
619 correspond to seasonal changes in sedimentation, where finer layers result from reduced  
620 subglacial meltwater input during winter (Flink and Noormets, 2018; Prothro et al., 2018).  
621 The sand layers with sharp basal contacts correspond to deposition by turbidity currents  
622 or turbid meltwaters sourced from the ice margin (Sheldon et al., 2016; Olsen et al., 2020).  
623 Dispersed clasts, indifferentially found in fine- or coarse laminae are interpreted as ice-  
624 rafted debris (IRD). The occurrence of randomly dispersed IRDs indicates that rain-out  
625 from icebergs was a minor sedimentation process contributing to LF3 (Syvitski, 1991;  
626 Sheldon et al., 2016) and/or accumulation rates were high, with dilution of the IRD signal  
627 (Syvitski and Shaw, 1995; Dowdeswell et al., 2000; Olsen et al., 2022). The latter  
628 interpretation is in agreement with the lack of bioturbation (Ó Cofaigh and Dowdeswell,  
629 2001; Sheldon et al., 2016; Callard et al. 2020). Rapid sediment loading causing minor  
630 submarine slope failures may explain the abundant occurrence of faults in LF3a (Callard  
631 et al., 2018; Allaart et al., 2020). The decrease of the thickness and grain size of the  
632 laminae and the disappearance of faulting in LF3b together reflect a progressively distant  
633 marine terminating ice margin, representing ice-proximal to distal glaciomarine

634 sedimentation (Ó Cofaigh and Dowdeswell, 2001; Dowdeswell et al., 2015; Streuff et al.,  
635 2017; Callard et al., 2018, 2020; Jenner et al., 2018; Allaart et al., 2020).

#### 636 *4.4.5. Unit 4: Ice-distal hemipelagic sedimentation*

637 The acoustic appearance of AF4 and LF4 corresponds to typical ice-distal  
638 hemipelagic sedimentary systems. Comparable acoustic facies have been identified in  
639 polar regions and are usually deposited by meltwater run off, tidal processes and, in a  
640 lesser extent, ice rafting sedimentation (Syvitski, 1991; Syvitski and Shaw, 1995; Hogan  
641 et al., 2016, 2020; Normandeau et al., 2017; Callard et al., 2018; Arosio et al., 2018;  
642 Streuff et al., 2018). The sediments show little to no ice-rafting debris (IRDs), suggesting  
643 a distal or terrestrial ice margins (Syvitski, 1991; Syvitski and Shaw, 1995; Sheldon et  
644 al., 2016; Callard et al., 2018; Olsen et al., 2020, 2022; Syvitski et al., 2022). The  
645 transition from weakly (LF4a) to intensively (LF4b) bioturbated mud essentially  
646 represents decreasing sedimentation rates with landward retreating glacier fronts (Callard  
647 et al., 2018; Jenner et al., 2018). The heavy bioturbation indicates hemipelagic  
648 sedimentation, similar to that of today.

#### 649 *4.4.6. Unit 5: Turbidites or mass-movement deposits*

650 Based on the transparent acoustic signature of AF5 and the unstratified sandy  
651 structure of LF5, unit 5 is interpreted as a coarse-grained unit from various high-energy  
652 sources such as turbidites or mass-movement deposits, which is supported by its  
653 occurrence at the foot of slopes or in front of glacially-fed tributary valleys (Syvitski and  
654 Shaw, 1995; Streuff et al., 2017; Hogan et al., 2020; Olsen et al., 2022). The presence of  
655 bioturbated mud lenses in between beds of coarse sand suggest that Unit 5 is not the  
656 continuity of Unit 4, but they are rather intermittent. The position of the coring sites in  
657 front of major valley deltas may indicate distinct cohesionless mass-movement deposits  
658 from remobilized coastal sediments (Gilbert et al., 1990).

659  
660  
661

**Table 2** Description of sedimentary units identified in the sub-bottom profiles data and sediment cores from the Clyde fjord-cross-shelf trough system. Color coding for the different units is the same as the one used for the acoustic facies in Fig. 8 and the lithological facies in Fig. 9.

Unit	Acoustic facies	Example	Description	Lithological facies	Example	Description	Interpretation
U0	AF0		Acoustically impenetrable. Homogenous and structureless. High-amplitude upper reflector. Occasional hyperbolae.				Acoustic basement (bedrock/ice-contact)
U1	AF1		Semi-transparent and chaotic. Discontinuous and homogenous reflectors. Non conformable geometry.	LF1		Unstratified olive dark gray sandy mud. Randomly disseminated pebble-sized clasts. Few traces of bioturbation.	Iceberg-influenced sedimentation
U2	AF2		Acoustically stratified. Low to medium amplitude parallel to sub-parallel internal reflectors. Conformable geometry.	LF2		Dark gray muddy-sand matrix diamicton. Weakly stratified and poorly sorted. Dispersed sub-angular to sub-rounded clasts. Interbedded with LF3.	Glacial debris-flows
U3	AF3		Acoustically stratified. Medium to high amplitude parallel internal reflectors. Basin fill (ponded) or onlapping geometry.	LF3		Laminated gray to dark gray silty mud. Sand laminae with sharp basal contact. Lack of bioturbation. Occasional clasts. Presence of faults at the base.	Ice-proximal glaciomarine sedimentation
U4	AF4		Acoustically stratified. High amplitude parallel and closely-spaced opaque reflectors. Conformable geometry.	LF4		Massive olive gray silty mud. Moderately to heavily bioturbated. Presence of black mottles and rare clasts. Gradational transition from LF3.	Ice-distal hemipelagic sedimentation
U5	AF5		Acoustically transparent to semi-transparent. Chaotic with rare internal reflectors. Hummocky surface and lenticular shaped geometry.	LF5		Unstratified gray coarse sand with bioturbated lenses of olive gray silty mud.	Turbidites and mass-movement deposits

662



663 **4.5. Radiochronology**

664 Seven ages were obtained from radiocarbon dated material collected from the  
665 sediments cores (Table 3).

666 **Table 3** Radiocarbon and calibrated radiocarbon ages from material collected in sediment cores.

Core number	Depth in core (cm)	Dated material	Laboratory ID	<sup>14</sup> C age yr BP	Calibrated age yr BP (2σ)
GeoB22344-3	472-483	Mixed benthic foraminifera	AWI-2620.1.1	8280 ± 90	8520 (8280-8840)
GeoB22346-3	387	Shell fragment	AWI-1726.1.1	5930 ± 50	6070 (5900-6250)
GeoB22346-3	765	Shell fragment	AWI-1727.1.1	8900 ± 190	9300 (8770-9850)
GeoB22346-3	766-783	Mixed benthic foraminifera	AWI-2619.1.1	8510 ± 100	8810 (8500-9120)
GeoB22348-3	888-896	Mixed benthic foraminifera	AWI-2618.1.1	8940 ± 100	9360 (9050-9600)
GeoB22358-3	92	Shell fragment	AWI-1728.1.1	10670 ± 177	11710 (11190-12290)
GeoB22359-3	156-166	Mixed benthic foraminifera	AWI-2617.1.1	10500 ± 100	11460 (11170-11800)

667 The AMS <sup>14</sup>C ages were calibrated within the age-depth modelling process, using the online software Calib 8.2 with the Marine20  
668 radiocarbon age calibration curve (Heaton et al., 2020). A local reservoir correction (ΔR) of 81±18 was used to account for the regional  
669 offset of the world ocean <sup>14</sup>C age, as determined by Pieńkowski et al. (2022).  
670

671 In Clyde Trough, at the base of core GeoB22359-3, mixed benthic foraminifera  
672 in iceberg-influenced sediments (LF1) provided an age of 11.5 ka cal. BP. A shell  
673 fragment from ice-proximal glaciomarine sediments (LF3), at 92 cm from the top of core  
674 GeoB22358-3, yielded an age of 11.7 ka cal. BP. These ages in deglacial sediments  
675 provide a minimum age for the deglaciation of the outer trough (Fig. 2). However, the  
676 age in GeoB22358-3 shows that it is most likely an underestimation of the deglaciation  
677 as the sample was collected near the top of the core.

678 In the inner Clyde Inlet (Fig. 3), three ages were provided from samples collected  
679 in ice-proximal glaciomarine sediments (LF3) from the base of two sediment cores. The  
680 sample at the base of sediment core GeoB22348-3, taken approximately 25 km from the  
681 fjord head, revealed an age of 9.4 ka cal. BP. The base of sediment core GeoB22346-3,  
682 taken 10 km from the fjord head, yields an age of 8.8 ka cal. BP. A shell fragment near  
683 the base of GeoB22346-3, yielding an age of 9.3 ka cal. ka BP, was rejected because it  
684 was inconsistently older than the benthic foraminifera sample taken a few centimeters  
685 below. The shell fragments could have been reworked, as the age range is consistent with

686 a reworked shell found in a delta at the fjord head and another one found in a nearby  
687 tributary valley (Briner et al., 2007). Therefore, the benthic foraminifera sample is  
688 favoured as being the most reliable age available for the base of that core. An age of 6.1  
689 ka cal. BP was also obtained from a shell fragment at 387 cm downcore, in ice-distal  
690 glaciomarine sediments (LF4).

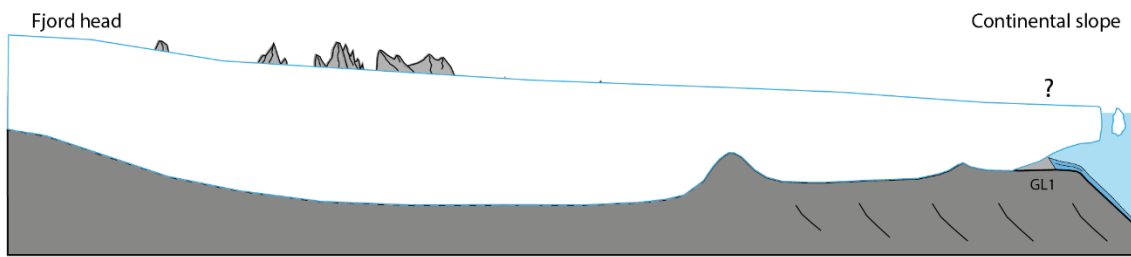
691 Similarly, a foraminifera sample collected in ice-distal glaciomarine sediments  
692 (LF4) from the base of sediment core GeoB22344-3 yielded an age of 8.5 ka cal. BP. It  
693 thus provides a minimum age for full deglaciation of Cormack Arm.

## 694 **5. Discussion**

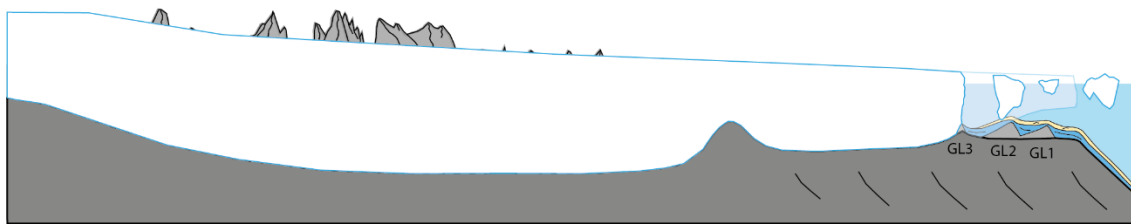
### 695 **5.1. Extent and retreat of the LIS margin**

696 The analysis of the multibeam bathymetry, acoustic profiles and lithological  
697 facies allowed the identification of landforms and sedimentary assemblages that are  
698 typical of high-latitude fjord-cross-shelf trough systems (e.g., Praeg et al., 2007;  
699 Winsborrow et al., 2010; Ó Cofaigh et al., 2013b; Sheldon et al., 2016; Slabon et al., 2016;  
700 Brouard and Lajeunesse, 2017; Ottesen et al., 2022). They provide valuable information  
701 on the maximal extent of the LIS margin and its retreat patterns along the entire Clyde  
702 fjord-cross-shelf trough system. However, deglacial ages on the shelf and nearby fjords  
703 of Baffin Island are scarce and correlations with other systems is therefore tentative. The  
704 radiocarbon dated horizons of the cores are stratigraphically too shallow to construct  
705 reliable age extrapolations to the base of the deglacial units. Nevertheless, they represent  
706 minimum-limiting ages at the coring sites and thus provide some constraints on ages of  
707 the deglaciation patterns we present. The compatibility with land-based studies provides  
708 independent corroboration that allow us to draw a more accurate chronology of the  
709 deglaciation for the Clyde fjord-cross-shelf trough system (Fig. 10).

a) Last Glacial Maximum (ca. 24 000 BP) - Full-glacial conditions

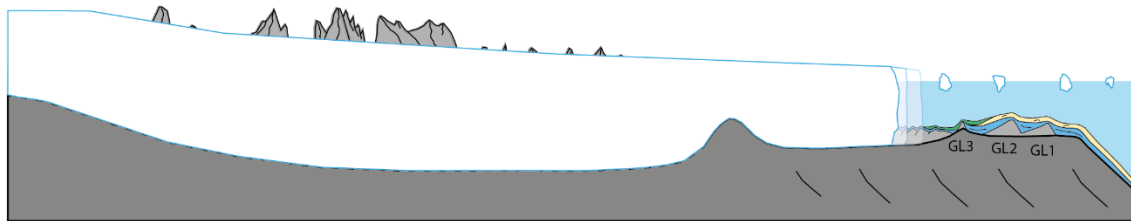


b) Early deglaciation (ca. 14 600 BP) - Ice shelf collapse

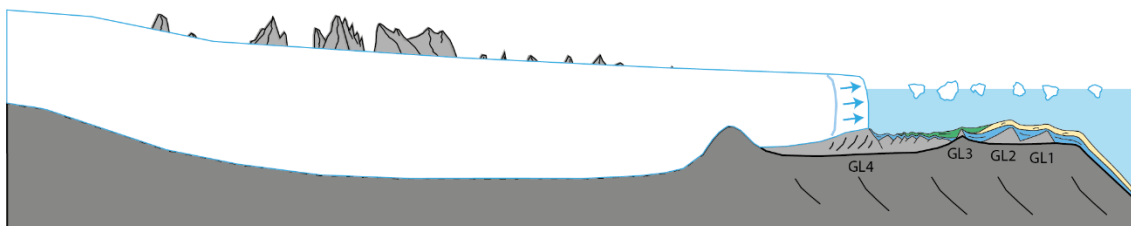


c) Late glacial (ca. 14 600-11 700 BP) - Deglacial stillstands and readvances

i) Recessional moraines



ii) Morainal bank



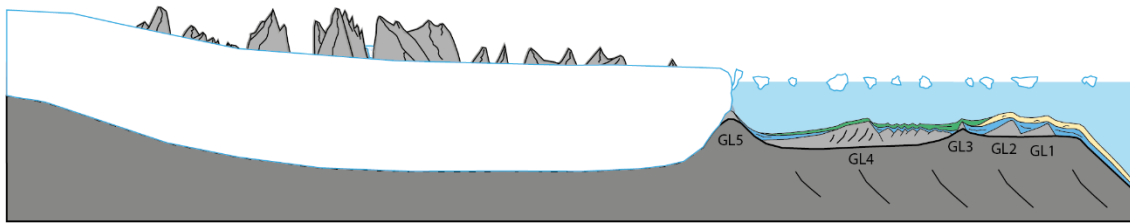
- |  |   |
|--|---|
|  U0: Bedrock/ice-contact deposits     |  U3: Ice-proximal glaciomarine sedimentation |
|  U1: Iceberg-influenced sedimentation |  U4: Ice-distal hemipelagic sedimentation    |
|  U2: Glacial debris-flows             |  U5: Turbidites and mass-movement deposits   |

710

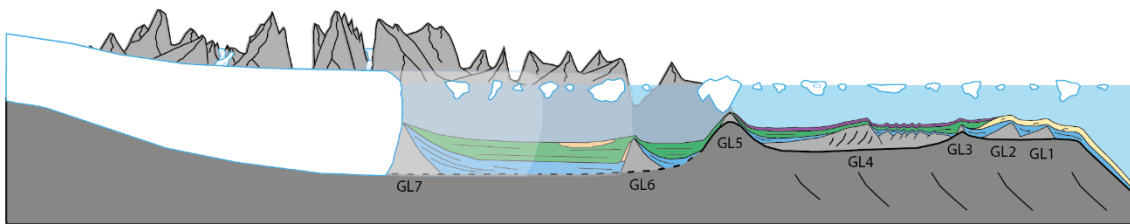
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**Figure 10** Schematic representation of the ice dynamics on the shelf corresponding to the different stages of ice retreat during deglaciation in Clyde fjord-cross-shelf trough system.

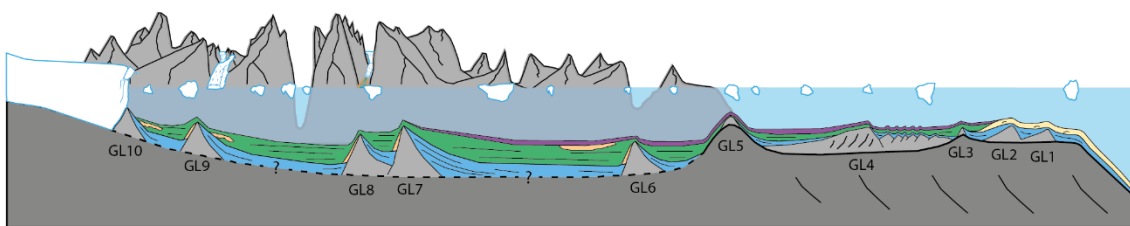
d) End of Younger Dryas (11 700 BP) - Fjord mouth sill stillstand



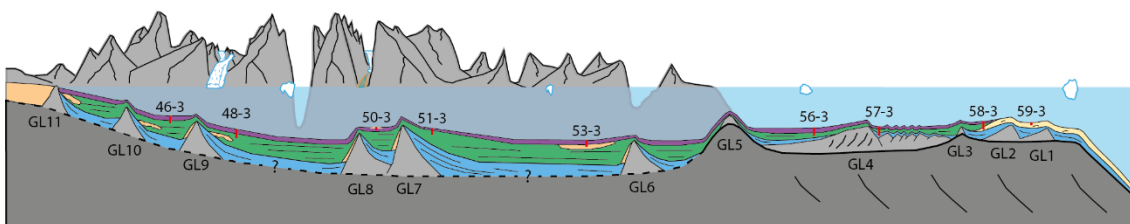
e) Early Holocene (11 700-9500 BP) - Fast retreat into the fjord



f) Cockburn substage (9500-8000 BP) - Fjord head stillstands or readvances



g) Late Holocene (8000 BP - Now) - Paraglacial conditions



713

714

715

Figure 10 Continued.

716 *5.1.1. LGM extent and collapse of the Clyde ice shelf*

717 Couette et al. (2022) asserted that the LIS did not reach the shelf break in Clyde  
718 Trough on the basis of geomorphological, sedimentological and glaciological evidence.  
719 They argued that the absence of gullies on the trough mouth fan and glacial lineations in  
720 the outer trough – which are both diagnostic of a grounded ice margin extending at the  
721 shelf break – suggest that the LIS had a receded position in Clyde Trough. It is also  
722 possible that the extensive iceberg ploughmarks observed in the outer trough have  
723 obscured any landforms that provide evidence for the LIS margin reaching shelf break  
724 during the LGM. In the absence of such evidence, the GZWs observed near the shelf edge  
725 are therefore identified as the most probable maximal position of the LIS in Clyde Trough  
726 during the LGM (Fig. 10a). The onset of deglaciation around Baffin Bay appears to have  
727 occurred between 16 and 14.6 ka cal. BP on eastern Baffin Island (Jennings, 1993; Briner  
728 et al., 2007; Margreth et al., 2017) and western Greenland (Sheldon et al., 2016; Jennings  
729 et al., 2017). This timing for the deglaciation of the shelves is in agreement with  
730 cosmogenic exposure dating from the Clyde Foreland, where deglaciation is interpreted  
731 to have started at ~15 ka BP (Briner et al., 2005). As no direct dating has yet yielded  
732 absolute ages on these landforms, it can not be asserted with certitude which of the two  
733 GZWs, if any, represent the LGM maximal extent. The change in orientation and density  
734 of iceberg ploughmarks suggest that the CIS margin was possibly grounded at GL2 during  
735 the early phases of deglaciation, when numerous icebergs were released into Baffin Bay  
736 and transported via the Baffin Island Current (Andrews et al., 1998). The large iceberg  
737 ploughmarks parallel to the trough on the stoss side of GL2 and the IRD-rich facies in the  
738 outer trough are, in turn, evidence of a period of extensive iceberg release provoked by  
739 the collapse of the Clyde ice shelf and rapid retreat of the CIS (Fig. 10b). Therefore, this  
740 period of rapid ice decay might correspond to Meltwater Pulse 1a (MWP-1a – 14.6-14.0  
741 ka cal. BP; Carlson, 2009; Harrison et al., 2018; Lin et al., 2021), which coincides with

742 the release of icebergs and detrital carbonate-rich sediments into Baffin Bay (BBDC layer  
743 1 – ~14.2-13.7 ka cal. BP) from northwestern Greenland and the eastern CAA (Andrews  
744 et al., 1998; Simon et al., 2012, 2014; Jackson et al., 2017).

745         It is unclear why the LIS margin did not reach the shelf break in Clyde Trough, as  
746 ice extended across the continental shelf in Lancaster Sound as well as in the Buchan and  
747 Scott trough systems of northeastern Baffin Island (Li et al., 2011; Brouard and  
748 Lajeunesse, 2017). Recent studies (i.e., Miller et al., 2002; Margreth et al., 2017) also  
749 suggest that the ice margin did not reach the shelf break on Cumberland Peninsula  
750 (eastern Baffin Island). The position of the ice margin at the LGM along the coast of  
751 eastern Baffin Island was most probably variable. Clyde Trough might represent a zone  
752 of transition between the “Big ice” model of northeastern Baffin Island and the “Just-  
753 Right ice” model observed on Cumberland Peninsula (see Miller et al., 2002). This  
754 intermediate position was also reported in Sam Ford Trough – just north of Clyde Trough  
755 –, although this system is considered to have been occupied by slow flowing ice during  
756 the LGM (Brouard and Lajeunesse, 2017). Slower flowing ice could also explain the  
757 receded position of the ice margin in Clyde Trough, as sparse and undefined glacial  
758 lineations in the outer trough suggests limited ice streaming activity at the LGM.  
759 However, the lack of glacial lineations on the outer trough possibly reflects the significant  
760 disturbance of the seafloor by iceberg ploughmarks. Nonetheless, it is worth noting that  
761 a slow ice flow regime was observed for the adjacent Clyde Lowlands, where ice flowing  
762 through Ayr Lake was non-erosive and slow flowing in its outermost part (Briner et al.,  
763 2005). The opening/diffluent configuration of the outer trough could have favored  
764 reduced ice velocity, thus limiting the formation of glacial lineations at the base of the ice  
765 stream. A reduced ice velocity could result from a limited catchment size due to the  
766 presence of more competent ice stream system on either side of the Clyde fjord-cross-



767 shelf trough system. Similar observations have been made on the shelf of northeastern  
768 Baffin Island, where piracy of ice drainage basins controlled the volume of ice flow into  
769 Sam Ford Trough system (Brouard and Lajeunesse, 2019c). Additionally, the CIS was  
770 sustained by cold-based ice on the banks in front of the Clyde and Aston lowlands (Briner  
771 et al., 2005). GZWs geometry indicates that cold-based ice also had a receded position on  
772 the inter-trough on both sides of Clyde Trough. Sediment assemblages on the Baffin slope  
773 further support that position, at least for the later part of the LGM (Jenner et al., 2018).  
774 Stacked tills underneath GL1 indicate that some earlier glaciations were, however, more  
775 extensive, as previously proposed by various authors for Arctic Canada (i.e., Miller et al.,  
776 1977; England et al., 2009).

#### 777 *5.1.2. Slow late glacial retreat on the shelf*

778         The absence of large GZWs in the middle and inner trough suggests that the ice-  
779 retreat following the initial breakup of the CIS occurred in a steady fashion as shown by  
780 sets of recessional moraines (Fig. 10c i). This slow retreat pattern differs from  
781 observations from other trough systems of northeastern Baffin Island, where deglaciation  
782 occurred in a more stepwise pattern with prolonged stillstands indicated by GZWs  
783 (Brouard and Lajeunesse, 2017). This deglaciation pattern, situated on the retrograde  
784 sloping part of the trough, is uncommon compared to other global occurrences which  
785 generally present few stillstand indicators and are usually attributed to a rapid ice margin  
786 retreat (Weertman, 1974; Favier et al., 2014; Wise et al., 2017). Similar variations in style  
787 and rate of ice retreat along a continental shelf has also been observed on the Antarctic  
788 Peninsula (Dowdeswell et al., 2008; Ó Cofaigh et al., 2014 and references therein) and  
789 may thus reflect the influence of local controls (i.e., bathymetry and topography) on ice  
790 dynamics of northeastern Baffin Island (see section 5.2).

791           The morainal bank (GL4) in the middle trough indicates a stage of ice margin  
792 readvance and short-term stabilization during the generally slow deglaciation of the  
793 trough (Fig. 10c ii). It aligns roughly with lateral moraines dated by Briner et al. (2005)  
794 near Patricia Bay, indicating that this event occurred at  $\sim 12.5 \pm 0.7$  ka. It is therefore  
795 probable that an ice margin readvance during the colder Younger Dryas ( $\sim 12.9 - 11.7$  ka  
796 BP) favoured the formation of the morainal bank. Regardless of their genesis as  
797 glaciotectonic push-and-thrust or crevasse-squeezed ridges, the transverse and zigzag-  
798 shaped ridges on the morainal bank suggest readvance of the LIS margin. These  
799 landforms may reflect warm-based fast flow conditions changing to cold-based freeze-on  
800 conditions near the ice margin (Christofferson and Tulaczyk, 2003; Laberg et al., 2009).  
801 Cold-based conditions at the ice-margin can be caused by ice thinning, fast downward  
802 advection of cold surface ice or basal freezing of the ice stream termination  
803 (Christofferson and Tulaczyk, 2003). Zigzag-shaped ridges have also been associated to  
804 ice readvances on the Værøy and Røst morainal banks, in northern Norway (Laberg et  
805 al., 2009). The Røst morainal bank is also seismically similar to the middle trough  
806 morainal bank with its folded reflectors at the front and the irregular thrust sheets in its  
807 core (Laberg et al., 2009). Folding and thrusting of sediment sheets by glaciotectonism  
808 (Fig. 7B), which is caused by movements at the front of the glacier (i.e., a readvance of  
809 the LIS margin), have been speculated to increase the height of ice-contact deposits  
810 (Powell, 1990; Lønne, 1995; Lyså and Vorren, 1997).

811           The moraine system on top of a bedrock sill at the fjord mouth (GL5) indicates  
812 that the ice margin stabilized here for a short period (Fig. 10d). In the cores collected  
813 along Clyde Trough, ages of 11.7 and 11.5 ka cal. BP below the transition from ice-  
814 proximal glaciomarine sediments to ice-distal hemipelagic sediments provide a minimal  
815 age for the ice margin retreat into the fjord. This transition is also marked by a peak in

816 Ca/Ti in the shelf cores, corresponding to an increase in detrital carbonate-rich sediments  
817 (BBDC layers) from northern Baffin Bay. This increase was also recorded along West  
818 Greenland and the onset of this event was dated at  $\sim 11.6$  ka cal. BP (Jennings et al., 2014,  
819 2017). This timing for ice retreat from the fjord-mouth at the onset of the Holocene is  
820 further supported by data from Scott Trough suggesting ice-distal sedimentation from  $\sim 12$   
821 ka cal. BP (Osterman and Nelson, 1989). Cosmogenic exposure dating from the  
822 continental domain indicates that the LIS margin was located at the Clyde Inlet mouth  
823 until  $\sim 11.7 \pm 2.2$  ka (Briner et al., 2005). These results suggest that the ice margin  
824 retreated into Clyde Inlet from the bedrock sill at the end of the Younger Dryas.  
825 Stabilizations commonly occur at fjord mouths due to the narrowing of the ice stream  
826 width, which reduces the ice flux across the grounding line (Åkesson et al., 2018). This  
827 stabilization could also be the result of a forced equilibrium in glacial mass balance, where  
828 accumulation is too high for the ice sheet margin to retreat into the deeper and narrower  
829 fjord but ablation is too high to allow a readvance onto the shelf and the open sea (Syvitski  
830 and Shaw, 1995).

### 831 *5.1.3. Episodic early Holocene retreat in the fjord*

832 The multibeam bathymetry imagery and acoustic sub-bottom profiles show series  
833 of moraine systems along the fjord, indicating a step-wise retreat of the LIS margin in the  
834 early Holocene, punctuated by intervening periods of fast retreat (Fig. 10e). Two  
835 continental moraine ridges located 10 km and 25 km up-fjord from Ailsa Island yielded  
836 cosmogenic exposure ages of  $11.2 \pm 1.2$  and  $10.0 \pm 1.5$  ka, respectively (Briner et al.,  
837 2007). These ages represent minimum ages for the deglaciation and are probably coeval  
838 or younger than the outer fjord moraine (GL6).

839 Few ages are available in the middle section of Clyde Inlet. However, basal ages  
840 collected in three lakes near the adjacent Inugsuin Fjord constrain the deglaciation of the

841 middle section to  $\sim 10.5$  ka cal. BP (Thomas et al., 2010). Ice retreat was probably more  
842 or less synchronous in Clyde Inlet, as the continental moraine ridges in the outer fjord  
843 yield slightly older ages and a bedrock sample from an unnamed island yielded a  
844 cosmogenic exposure age of  $10.2 \pm 2.2$  ka (Briner et al., 2007). Furthermore, cosmogenic  
845 dating in Naqsaq Valley indicate alpine glacier stabilization and moraine deposition at  
846  $\sim 10.2 \pm 0.2$  ka (Young et al., 2021). It is therefore possible that the middle fjord moraines  
847 (GL7 and GL8) were deposited around that time.

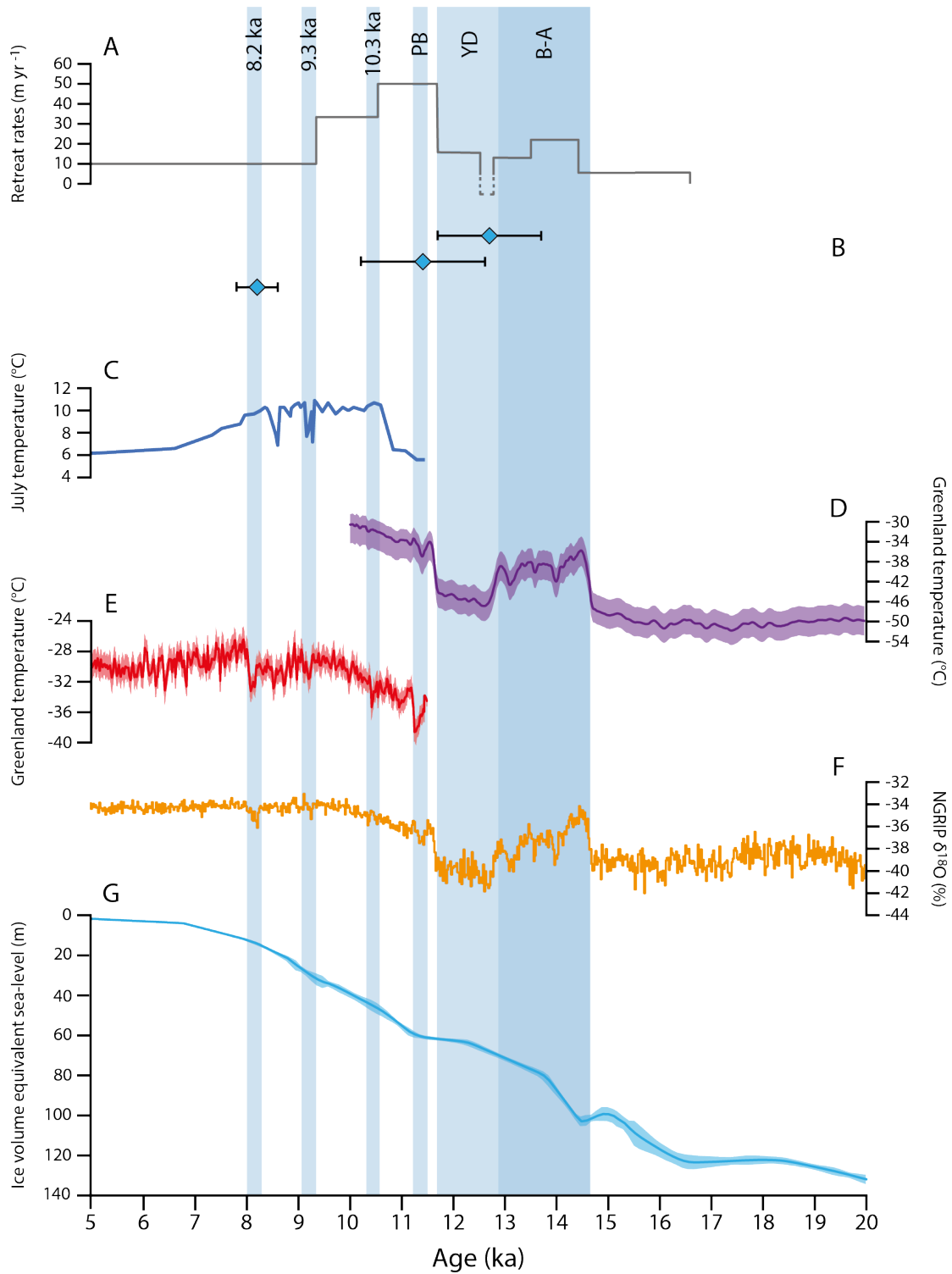
848         During the Cockburn substage, multiple moraines were deposited in the inner  
849 fjord and at the fjord head (Fig. 10f). Wood and shell samples collected in a tributary  
850 valley 40 km upstream from the fjord head by Briner et al. (2007) yielded identical ages  
851 of  $\sim 9.3$  ka cal. BP for deglaciation of the inner fjord. These ages are in agreement with  
852 new basal ages from sediment cores GeoB22348-3 and GeoB22346-3 (Fig. 9), which in  
853 turn are constraining the timing of the inner fjord moraines (GL9-GL11) between 9.4 and  
854 8.8 ka cal. BP. Similarly, radiocarbon ages from shells collected in deltas at the head of  
855 Clyde Inlet indicate that the ice margin retreated beyond the fjord head between 9.1 and  
856 8.6 ka cal. BP (Briner et al., 2007). Similar ages from adjacent Inugsuin Fjord and Sam  
857 Ford Fjord corroborate these observations (Andrews and Drapier, 1967; Briner et al.,  
858 2009a; Syvitski et al., 2022). An ice-contact delta located at the fjord head yielded  
859 cosmogenic exposure age of  $\sim 8.3$  ka  $\pm 0.3$  ka (Briner et al., 2007; Young et al., 2013).  
860 Radiocarbon ages collected 4 km upstream of the ice-contact delta yielded an age of  $\sim 7.9$   
861 ka cal. BP for the deglaciation of the fjord head (Briner et al., 2007). Deglaciation of the  
862 fjord head is marked by a steep decline in Ca/Ti, corresponding to a decrease in coarser  
863 sediment input from the ice margin directly into the fjord. The age of  $\sim 7.9$  ka cal. BP is  
864 also in agreement with the shell sample from core GeoB22346-3 indicating that ice had  
865 retreated from the fjord head before 6.2 ka cal. BP.

866 The results from Clyde Inlet notably contrast with models previously proposed  
867 from eastern Baffin Island, where the LIS was believed to have retreated in a catastrophic  
868 pattern along the fjords (Briner et al., 2007, 2009a). Alternatively, the results support a  
869 more episodic deglaciation model with multiple ice margin stabilizations and moraine  
870 formation proposed for fjords of northeastern Baffin Islands (Brouard and Lajeunesse,  
871 2019a).

872 Subsequently to the withdrawal of its ice margin from Clyde Inlet (Fig. 10g), the  
873 LIS began a slow retreat towards the south until the Barnes Ice Cap became isolated  
874 (Miller et al., 2005; Briner et al., 2009). The LIS and local glaciers receded beyond their  
875 current position until the onset of the Neoglacial ~4.5 ka BP (Miller et al., 2005; Young  
876 et al., 2015). During that period, cooler climatic conditions prevailed and local glaciers  
877 readvanced into the fjord, as marked by lateral moraines from tributary valleys and  
878 glaciofluvial fans (Fig. 6A-B). Rapidly deposited layers are identified in cores of the inner  
879 fjord and are marked by sharp peaks in Ca/Ti ratios and could relate to these readvances.  
880 These coarse layers could also either represent increased sediment input by glacial  
881 meltwater or deposits by glacier-lake outburst floods caused by the oscillation of the  
882 Barnes Ice Cap repeatedly blocking the Clyde River outlet (Barnett and Holdsworth,  
883 1974; Andrews and Barnett, 1979).

## 884 **5.2. Controls on stabilizations and variability of ice retreat**

885 Several external (e.g., atmospheric temperatures, changes in sea level and ocean-  
886 driven melting) and local factors (e.g., topography and bathymetry) have possibly  
887 asserted a certain control on the deglaciation patterns in the Clyde fjord-cross-shelf trough  
888 system. Comparing the chronology presented in the previous section to a compilation of  
889 climatic records allows establishing correlations between the deglaciation and potential  
890 factors that may have influenced retreat of the ice margin across the system (Fig 11).



891

892 **Figure 11** (A) Average retreat rates between stabilizations in the Clyde fjord-cross-shelf trough system.  
 893 The dashed line represents the uncertainties regarding the length and magnitude of the ice margin readvance  
 894 during the Younger Dryas. (B) Exposure ages of moraine ridges along Clyde Inlet from Briner et al. (2007),  
 895 recalculated using the Baffin Bay production rate (Young et al., 2013). (C) Chironomid-derived July  
 896 temperature reconstruction from Lake CF8, eastern Baffin Island (dark blue; Axford et al., 2009). (D)  
 897 Greenland mean-annual temperatures reconstructed using gas-phase  $\delta^{15}\text{N-N}_2$  measurements (purple -  $\pm 1\sigma$ ;  
 898 Buizert et al., 2014). (E) Greenland mean-annual temperatures reconstructed using gas-phase  $\delta\text{Ar-N}_2$   
 899 measurements (red -  $\pm 2\sigma$ ; Kobashi et al., 2017). (F)  $\delta^{18}\text{O}$  record from NGRIP project (orange; Rasmussen  
 900 et al., 2014). (G) Ice volume equivalent sea-level (blue -  $\pm 1\sigma$ ; Lambeck et al., 2014). Vertical bars represent  
 901 cold and warm intervals discussed in the text. B-A: Bølling–Allerød; YD: Younger Dryas; PB: Preboreal.



902 *5.2.1. External forcing as the principal driver of ice retreat patterns*

903         The temporal framework presented in the previous section suggest that climate  
904 played a dominant role in driving first-order deglaciation patterns in Clyde Inlet, as ice  
905 retreat coincides with periods of warming climate and stabilizations are synchronous with  
906 cooling events (Fig. 11). In turn, sea level changes and oceanic forcing might also have  
907 influenced the retreat rates of the ice margin along the Clyde fjord-cross-shelf trough  
908 system (Fig. 11).

909         Rapidly rising sea-level (>40 mm/year; Lambeck et al., 2014) probably favored  
910 the initial collapse of the Clyde ice shelf and rapid retreat of the CIS in the early phases  
911 of the deglaciation (Couette et al., 2022). Changes in the eustatic level may have provoked  
912 the lift-off of the local ice shelf, increasing the area of the ice margin in contact with the  
913 water column, making it more vulnerable to ocean forcings (Joughin et al., 2012;  
914 Jamieson et al., 2014; Jennings et al., 2018). The global sea-level rise initiated at ~16.5  
915 ka eventually led to the ice front destabilization, which induced rapid retreat of the ice  
916 margin (~25 m/year). Rapidly rising global sea level has been speculated to have triggered  
917 the collapse and/or rapid retreat of marine-based ice sheets in the Northern Hemisphere  
918 following the LGM (e.g., Winsborrow et al., 2010; Jakobsson et al., 2011; Rydningen et  
919 al., 2013; Hogan et al., 2016; Arndt et al., 2017; Callard et al., 2018). In turn, the timing  
920 of ice-margin retreat from outer Clyde Trough is synchronous with the onset of the  
921 Bølling–Allerød interstadial (~14.5 – 12.9 ka BP), a period of globally warmer  
922 temperatures (Rasmussen et al., 2014). It also coincides with extensive ice mass loss  
923 around Baffin Bay, as ice streams were retreating from the outer shelf in Western  
924 Greenland (i.e., Sheldon et al., 2016; Jennings et al., 2017) and large numbers of icebergs  
925 were released from northern Baffin Bay (i.e., Andrews et al., 1998; Simon et al., 2012;  
926 Jackson et al., 2017). Reduced global sea level rise during the second half of the Bølling–

927 Allerød (~12 mm/year – Lambeck et al., 2014) probably contributed to a slow-down of  
928 the ice-margin retreat in Clyde Trough (~12.5 m/year). However, lower rates of sea level  
929 rise alone do not explain the slower deglaciation pattern, as it was not observed in  
930 neighbouring troughs (Brouard and Lajeunesse, 2017).

931         The beginning of the Younger Dryas (~12.9 – 11.7 ka BP) was regionally marked  
932 by an abrupt lowering of the temperature by 2°C (Rasmussen et al., 2014). This sudden  
933 decrease in temperature likely favoured a readvance of the LIS margin, before resuming  
934 its slow retreat as the temperature gradually increased again. The presence of the moraine  
935 indicating stagnation at the fjord mouth is consistent with the relatively cold conditions  
936 across Baffin Bay at the end of the Younger Dryas (Buizert et al., 2014; Rasmussen et al.,  
937 2014) that promoted a positive mass balance and possibly counteracted ice loss by calving  
938 at the ice margin.

939         Alternatively, the early Holocene was marked by warmer atmospheric  
940 temperature around the Baffin Bay region, which favoured extensive ice margin retreat  
941 (Pendleton et al., 2019; Lesnek et al., 2020). While temperatures were generally warm  
942 during the Holocene, cold spells were recorded in different proxies (i.e., Axford et al.,  
943 2009; Rasmussen et al., 2014) and appear to have favored ice margin stabilizations in  
944 Clyde Inlet. This deglaciation model is in agreement with observations around Baffin Bay  
945 where widespread moraine deposition have been associated with cold-climate oscillations  
946 at 11.3, 10.4, 9.3 and 8.2 ka (Young et al., 2020; Lesnek et al., 2020).

947         Inflow of subsurface water might have been an additional contributor to retreat  
948 rates in our study area through deglaciation. The intrusion of subsurface warm water have  
949 provoked accelerated melting and enhanced ice-margin retreat in modern-day marine-  
950 terminating glaciers (e.g., Straneo et al., 2011; Jeong et al., 2016; Howe et al., 2019), as

951 well as on the shelves and fjords of many formerly glaciated regions (e.g., Sheldon et al.,  
952 2016; Arndt et al., 2017; Batchelor et al., 2019a; Allaart et al., 2020). However, the Baffin  
953 Island Current incorporates colder Arctic water from the CAA and Nares Strait (Tang et  
954 al., 2004; Münchow et al., 2015), therefore lowering the layer of warmer subsurface  
955 water. A reduction in the depth of warmer currents would weaken its influence on ice  
956 retreat along the Baffin shelf, in particular for the shallower Clyde Trough. Moreover, the  
957 shallow fjord-mouth sill (< 200 m) probably prevented subsurface warm water from  
958 entering the fjord and triggering a catastrophic retreat. It is therefore unlikely that inflow  
959 of warmer currents had major influence on the retreat of the CIS.

960         Although external forcing played an unequivocal role in the deglaciation of Clyde  
961 Inlet, the variability in retreat patterns when compared to other neighbouring fjord-cross-  
962 shelf trough systems indicates a local control on LIS margin oscillations.

### 963 *5.2.2. Fjord/trough geometry controlling ice margin stabilization*

964         Local-scale topography appears to have been a key factor in controlling the  
965 location of stabilizations during the retreat of the CIS. Depositional wedges and related  
966 stabilizations have been noted where topographic constrictions, highs and/or bends in the  
967 trough orientation occur. Although most topographic controls in the trough are bedrock-  
968 influenced, older GZWs might have produced a pinning point adequate enough to help  
969 the ice margin stabilize in the outer trough at the LGM (Fig. 7). Acting as a topographic  
970 high, these older GZWs restrained the LIS from flowing farther seaward regardless of the  
971 wider diffluent bed morphology by increasing the basal drag exerted on the ice stream  
972 (Dowdeswell and Fugelli, 2012; Batchelor and Dowdeswell, 2015; Hogan et al., 2016;  
973 Bart et al., 2017; Danielson and Bart, 2019; Greenwood et al., 2021; Ottesen et al., 2022).  
974 A bend in the trough orientation, corresponding to a change in bedrock lithology, possibly  
975 further contributed to the formation of the morainal bank by enhancing lateral drag on the

976 side of the glacier (Syvitski and Shaw, 1995; Lyså and Vorren, 1997; Laberg et al., 2009;  
977 Jamieson et al., 2012, 2014; Ó Cofaigh et al., 2014; Bradwell et al., 2019). The relatively  
978 shallower depths compared to other cross-shelf trough systems of northeastern Baffin  
979 Island probably restricted the area of the ice front in contact with the ocean, resulting in  
980 a slower deglaciation in the middle trough as observed in similar settings (Arndt et al.,  
981 2017; Jakobsen et al., 2020). The shallower depths likely also contributed to the slower  
982 deglaciation pattern observed across the retrograde slope of the mid- and inner trough  
983 following the YD readvance.

984         In contrast, deeper water in the fjord may have favored the acceleration in ice  
985 retreat (>50 m/year). Loss of contact with vertical pinning point reduces drag which, in  
986 turn, increases mass flow and iceberg calving rates (Syvitski and Shaw, 1995; Jamieson  
987 et al., 2014; Batchelor et al., 2019a). Coupled with the generally warmer atmosphere  
988 temperature of the early Holocene, it created an ideal setting for an enhanced retreat rate  
989 of the LIS in Clyde Inlet. However, cold events probably provoked intervening  
990 stabilizations of the ice margin in Clyde Inlet, while the fjord geometry influenced the  
991 location of most stabilizations. The outer fjord moraine is located at a pinning point  
992 created by multiple islands at the confluence of Clyde Inlet and Inugsuin Fjord. This  
993 pinning-point allowed the ice margin to anchor and stabilize on the topographic high  
994 between islands. The location of the middle fjord moraine at the confluence of Clyde Inlet  
995 and Cormack Arm suggests, however, the influence of a funnel-shaped constriction of the  
996 ice (i.e., Syvitski and Shaw, 1995; Brouard and Lajeunesse, 2019a). Other ice-margin  
997 stabilizations in the inner fjord occurred at bends and lateral constrictions of the fjord  
998 width because of enhanced lateral-drag (Jamieson et al., 2012; Åkesson et al., 2018;  
999 Batchelor et al., 2019a; Brouard and Lajeunesse, 2019a). It must be emphasized here that

1000 not all constrictions or bends are associated with a stabilization in the fjord, as climate is  
1001 likely the main driver for initiating a slow-down of the ice retreat.

## 1002 **6. Conclusions**

1003 The combination of multibeam bathymetry imagery, seismostratigraphic profiles  
1004 and sediment cores collected in the Clyde Inlet fjord-cross-shelf trough system (NE  
1005 Baffin Island) provide new insights into the extent and retreat patterns of the Laurentide  
1006 Ice Sheet margin in the region during the Last Glacial cycle. Key results of this analysis  
1007 are:

- 1008 • The LIS margin probably did not extend all the way across the continental shelf in  
1009 Clyde Trough during the LGM; its maximal extent was rather located some 10 km  
1010 from the shelf break. However, as the absence of direct dating on the upper continental  
1011 slope restrains us from providing confirmation, the question of the maximal extent of  
1012 the LIS in Clyde Trough remains open.
- 1013 • Deglaciation on the shelf was temporally constrained to the late glacial (16-11.7 ka).  
1014 It was marked by an initial collapse of the Clyde ice shelf and rapid LIS margin retreat,  
1015 followed by a slow retreat of the ice margin with intervening stabilizations interrupted  
1016 by a readvance during the Younger Dryas. This deglaciation pattern differs from  
1017 observations made in other troughs of northeastern Baffin Island shelf, where it  
1018 appears to have been more rapid and episodic with wider spaced GZWs.
- 1019 • Similarly to other fjords of northeastern Baffin Island, the ice margin retreated into  
1020 Clyde Inlet in a less catastrophic pattern than previously proposed for the early  
1021 Holocene (11.7-8 ka). Our age constraints support earlier works that suggest  
1022 numerous ice margin stabilizations during the early Holocene, which could be linked

1023 to cold climate events at ~10.3, ~9.3 and ~8.2 ka. The retreat was, however, rapid  
1024 between successive stabilizations due to greater water depths in the fjord.

- 1025 • Climate was the main driver of deglaciation in the Clyde area, as the available  
1026 chronology suggests that most stabilizations coincided with regional-wide cooling  
1027 events. Deglaciation patterns in the Clyde area were strongly influenced by  
1028 topography, as ice margin stabilizations occurred at pinning points in both the trough  
1029 and fjord. Oceanic forcing, such as global sea level fluctuations and ocean  
1030 temperatures, appears to only have a secondary influence on rates of ice sheet retreat  
1031 in the Clyde fjord-cross-shelf trough system.

1032 These results highlight the variability of ice-sheet retreat patterns and controls  
1033 along a single high Arctic fjord-cross-shelf trough system, and from one system to  
1034 another. However, uncertainties remain concerning the timing of ice margin stabilizations  
1035 on northeastern Baffin Island, especially on the continental shelf. Future work on Baffin  
1036 Island fjord-cross-shelf trough systems should therefore focus on establishing a robust  
1037 deglaciation chronology combining both marine- and terrestrial-based investigations,  
1038 which would improve knowledge on factors controlling glacier behaviour and provide  
1039 key information for testing numerical simulations on climate and predicting future ice  
1040 mass loss in a warming world.

#### 1041 **Data availability**

1042 The multibeam bathymetric data collected in 2017 during expedition MSM66 have been  
1043 deposited in the PANGAEA repository (<https://doi.org/10.1594/PANGAEA.902341>).

1044 The multibeam bathymetric data from ArcticNet cruises can be visualized on the  
1045 Université Laval Géoindex + website (<http://geoindex-plus.bibl.ulaval.ca>). The  
1046 Parasound profiles collected in 2017 during expedition MSM66 have been deposited in

1047 the PANGAEA repository (<https://doi.pangaea.de/10.1594/PANGAEA.944843>). The  
1048 seismic reflection data along with the acquisition specifics are available on the Marine  
1049 Data Holding public repository of National Resources Canada (<http://geogratis.gc.ca/>).

## 1050 **Author contributions**

1051 POC conceived the study and initiated this project in cooperation with PL and  
1052 JFG. POC interpreted the geophysical data sets, conducted the sediment core analysis,  
1053 wrote the paper and prepared the figures. PL, JFG, BD, CG, DH, and EB were involved  
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