

2020-10-03

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<http://hdl.handle.net/10026.1/16486>

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10.1016/j.geomorph.2020.107453

Geomorphology

Elsevier BV

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1 **Role of dense shelf water in the development of Antarctic submarine canyon morphology**

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15 **Keywords:** slope process, continental slope, submarine gully, meltwater

16

17 **1. Abstract**

18 Increased ocean heat supply to the Antarctic continental shelves is projected to cause accelerated ice sheet loss  
19 and contribute significantly to global sea-level rise over coming decades. Changes in temperature or salinity of  
20 dense shelf waters around Antarctica, resulting from increased glacial meltwater input, have the potential to  
21 significantly impact the location and structure of the global Meridional Overturning Circulation, with seabed  
22 irregularities such as submarine canyons, driving these flows toward the abyss. Submarine canyons also  
23 influence the location of intruding warm water currents by acting as preferential routes for rising Circumpolar  
24 Deep Water. These global changes have implications for large-scale effects to atmospheric and oceanic  
25 circulation. The ability for numerical modellers to predict these future behaviours is dependent upon our ability  
26 to understand both modern and past oceanic, sedimentological and glaciological processes. This knowledge  
27 allows ocean models to better predict the flux and pathways of Circumpolar Deep Water delivery to the shelf,  
28 and consequently to ice shelf cavities where melt is concentrated. Here we seek to understand how dense shelf  
29 water and other continental slope processes influence submarine canyon morphology by analysing newly

30 collected geophysical and oceanographic data from a region of significant and prolonged dense shelf water  
31 export, the Hillary Canyon in the Ross Sea. We find that cascading flows of dense shelf water do not contribute  
32 to significant gully incision at the shelf edge during interglacial periods, however, are strong enough to prevent  
33 gully infilling and contribute to canyon-levee aggradation down-slope. We find buried paleo-gullies beneath  
34 gullies incising the modern seafloor. Paleo-gullies occur as single gullies and in complexes indicating that gully  
35 activity was continuous over multiple glacial cycles and formed an important role in the development of the  
36 shelf edge and upper slope. Glacial cycles likely drive large-scale shifts in canyon head processes with periods  
37 of intense seafloor erosion and significant gully incision likely occurring when ice grounded near to the shelf  
38 edge, during glacial and deglacial periods, when sediment-laden subglacial meltwater was released at the shelf  
39 edge. We put slope morphology observed at the Hillary Canyon head into global perspective to show that  
40 cascading flows of dense shelf water do not exert consistent patterns of erosion on high-latitude continental  
41 margins.

## 42 **2. Introduction**

43 Increased ocean heat supply to Antarctic continental shelves is projected to cause accelerated ice sheet loss and  
44 contribute significantly to global sea-level rise over coming decades to centuries (De Conto et al., 2016; IPCC,  
45 2019). Currently, numerical modelling studies lack necessary resolution and spatial coverage of seafloor  
46 morphology data to sufficiently constrain past and future sub-ice shelf melting, ice-sheet collapse, and sea level  
47 change estimates (Petrini et al., 2018; Colleoni et al., 2018). This is because one of the major causes of current  
48 Antarctic ice sheet retreat stems from increased ocean heat supply to the continental shelves surrounding  
49 Antarctica, with atmospheric temperature rise contributing to a lesser extent (Rignot et al., 2019; IPCC, 2019).  
50 Recent studies show that heat and volume transport around Antarctica are substantially enhanced where seafloor  
51 irregularities, such as submarine canyons, allow dense shelf waters to descend down-slope (Morrison et al.,  
52 2020). Warm water incursions onto the shelf can be intermittent and highly localised and can vary depending on  
53 the geometry of the ice shelf and seafloor bathymetry (Padman et al., 2018). Lack of necessary resolution and  
54 data availability to image these irregularities therefore makes predictions and future estimates of ice sheet and  
55 oceanic changes difficult. Thinning of ice shelves, due to increasing ocean temperatures and warm water  
56 incursions can lead to rapid ice retreat. This is especially true where marine based ice-sheets occur in  
57 conjunction with a landward sloping seabed, seen around the West and East Antarctic Ice Sheet (WAIS and

58 EAIS, respectively) (Joughin et al., 2011). This fast ice-shelf retreat can lead to an overall disintegration of the  
59 marine-based section of ice-sheets.

60 Dense shelf water around Antarctica is produced predominantly in the Weddell and Ross Sea polynyas where  
61 seabed irregularities, such as canyons and basins, can drive these flows from the shelf to the deep ocean where  
62 mixing with ambient water drives the global Meridional Overturning Circulation (Jacobs et al., 2002; Purkey  
63 and Johnson, 2010). Changes in temperature or salinity of these waters, such as due to increased meltwater input  
64 from ice sheets and seasonal sea ice melt, lead to significant changes to the Meridional Overturning Circulation  
65 (Jacobs et al., 2004; Seidov et al., 2001; Weaver et al., 2003; Seidov et al., 2005). Recent studies show that  
66 freshening of bottom waters is already occurring in the Ross and Weddell seas due to factors such as increased  
67 ice mass loss in the Antarctic Peninsula and Amundsen Sea (Silvano et al., 2018). This has significant global  
68 implications for large-scale effects to ocean and atmospheric circulation, with changes to the strength of the  
69 Meridional Overturning Circulation potentially leading to abrupt and global climate changes (Ramhstaf, 1994;  
70 Manabe and Stouffer, 1995). Understanding the influence that dense shelf water has on seafloor morphology  
71 and vice versa has important implications for determining how these processes changed in the past, in response  
72 to different climatic conditions and ice sheet configurations on the continental shelf, and how ice sheet  
73 configuration may change in the future. Submarine canyons influence the location and dynamics of intruding  
74 warm waters that enhance melting and ice shelf retreat with dense shelf water cascading driving net onshore  
75 heat and volume transport of warm currents around Antarctica (Dinniman et al., 2003; Morrison et al., 2020).  
76 Thus, understanding how canyon morphology and isobath curvature evolved through time is crucial in  
77 understanding factors, processes and feedbacks contributing to past and future ice sheet retreat.

78 Ice-ocean interactions are arguably the most poorly constrained aspect of ice sheet, ocean and climate models.  
79 Extensive Antarctic paleo-climate records are recovered from the flanks of canyons or drifts associated with  
80 sediment delivery down-canyon (e.g. Rebesco et al., 2007; Barker and Camerlenghi, 2002), yet very little is  
81 known about canyon process across modern, as well as glacial-interglacial timescales. Understanding large-  
82 scale oceanic feedbacks as well as the differences in amplitude and frequency of Antarctic continental slope  
83 processes under different climatic conditions, noted in far-field paleo-oceanographic records, is essential for  
84 constraining how ice sheet and oceanic interactions may change in the future (e.g. Zachos et al., 2001).

85 Understanding the dynamics of processes operating in submarine canyons more widely is of global importance.  
86 This can improve understanding of hazards such as turbidity currents which transport the greatest sediment

87 volumes on earth (Talling et al., 2014). These flows have significant influence on the global carbon flux (Galy et  
88 al., 2007) and can influence shelf and deep-sea ecosystems through supplying nutrients (Canals et al., 2006).  
89 They contribute to continental margin and fan construction and aid transportation of pollutants to the deep-sea  
90 (Nilsen et al., 2008). On glaciated margins, canyons are not ubiquitous as on low latitude margins and are  
91 particularly rare on certain margins (Rui et al., 2019). Their study contributes to a better understanding of the  
92 dynamics of ice buildup and retreat and associated glacigenic sediment transport which is crucial in  
93 understanding the spatial and temporal variability of glacimarine and oceanographic processes operating on  
94 high-latitude margins.

95 We investigate how dense shelf water influences submarine canyon morphology by analysing new geophysical  
96 and oceanographic data from a region of significant and prolonged dense shelf water export, the Hillary Canyon  
97 in the Eastern Ross Sea, Antarctica (Fig. 1) (Bergamasco et al., 2002; Orsi et al., 2009; Morrison et al., 2020).  
98 We present a quantitative analysis of the main morphological features at the Hillary Canyon head, including  
99 gullies incising the modern seafloor and buried paleo-gullies and discuss processes influencing their distribution  
100 and formation. We discuss the ability for dense shelf water to influence canyon morphology and the modern and  
101 past implications of this before discussing the effects of dense shelf water around Antarctica and other high-  
102 latitude continental margins more widely.

### 103 **3. Study Area**

104 The Hillary Canyon is a 180 km long, non-shelf incising canyon on the Eastern Ross Sea continental margin,  
105 Antarctica (Fig. 1). The morphology of the canyon is largely unknown due to a paucity of seafloor hydrographic  
106 data with much of the seafloor lacking sufficient resolution data. The continental shelf landward of the Hillary  
107 Canyon spans ~300 km to the modern front of the Ross Ice Shelf and is dissected by numerous deep, glacially  
108 carved troughs. The canyon lies at the mouths of two of those glacial troughs, the Pennell Trough and Glomar  
109 Challenger Basin, that were occupied by ice streams over multiple glacial cycles (Cooper et al., 1991; De Santis  
110 et al., 1995; Anderson et al., 2014; Halberstadt et al., 2016; Bart and Owolana, 2012). Grounded ice from the  
111 WAIS reached the shelf edge in the Eastern Ross Sea periodically since the Miocene and is thought to have last  
112 retreated during the Holocene (e.g. Anderson and Bartek, 1990; De Santis et al., 1995; McKay et al., 2016;  
113 Anderson et al., 2019).

114 The Ross Sea lies on the Western Antarctic Rift System and consists of three major sedimentary basins that  
115 formed during the break-up of Gondwana during the Late Cretaceous due to crustal extension and thinning

116 (Cooper et al., 1991). The basins are filled with eight major sedimentary packages following subsidence and  
117 deposition of prograding subglacial and glacial marine sedimentary packages with the oldest dated package  
118 (upper RSS-1) dating to the upper Eocene – Oligocene period (Traube and Zayatz, 1993; Brancolini et al., 1995;  
119 De Santis et al., 1995). To the east, the eastern Ross Sea Trough Mouth Fan forms a major sediment depocenter  
120 at the shelf edge seaward of the Glomar Challenger and the Whales Deep Basins likely dating to the Pliocene-  
121 Pleistocene (Alonso et al., 1992; Bart et al., 2000). The fan is characterised by outward-bulging contours with  
122 seismic data showing progradation of the margin. To the west of the Hillary Canyon, the Iselin bank forms a  
123 distinct topographic barrier with steep flanks.

124 The Hillary Canyon is a major conduit for Ross Sea Bottom Water (Bergamasco et al., 2002; Orsi and  
125 Wiederwohl., 2009; Morrison et al., 2020) contributing to Antarctic Bottom Water exported from the Ross Sea.  
126 Dense shelf water forms through sea ice formation and subsequent brine rejection and heat loss. High Salinity  
127 Shelf Water (characterised by salinity values of 34.75 to 34.85 psu; Jacobs et al., 1985) and Ice Shelf Water  
128 (characterised by temperatures below freezing due to supercooling beneath the Ross Ice Shelf) mix with ambient  
129 water (predominantly modified Circumpolar Deep Water; mCDW) forming dense shelf water (Bergamasco et  
130 al., 2002). Circumpolar Deep Water (CDW) can intrude onto the Ross Sea shelf and mix with shelf waters,  
131 where it forms mCDW (Budillon et al., 2011). These shelfward intrusions are enhanced, amongst other factors,  
132 by the presence of seafloor depressions, such as the Hillary Canyon (Dinniman et al., 2003). Oceanographic and  
133 modelling studies show that dense shelf water moves toward the shelf break along seafloor depressions and  
134 flows from the shelf to the abyss (Bergamasco et al., 2002; Budillon et al., 2011; Morrison et al., 2020).

#### 135 **4. Methodology**

136 Geophysical and oceanographic data were collected in austral summer 2017 by R/V OGS Explora during the  
137 EUROFLEETS-funded ANTSSS expedition. The geophysical data covers 1575 km<sup>2</sup> at the Hillary Canyon head.  
138 Data were collected using a Reason SeaBat 7150 multibeam echosounder with an operating frequency of 12  
139 kHz and swath width of 150°. Data were processed using PDS2000 to 50 m grid size. A Benthos Chirp III  
140 collected acoustic sub-bottom profiler data (frequency range 2-7 kHz; maximum ping rate 15 Hz). Single-  
141 channel seismic data were acquired over 453 km. The system was configured with a linear array of two 210  
142 cu.in GI guns spaced 2 m apart and towed 4 m below the surface. Both guns were shot in harmonic mode (105  
143 G + 105 I) with a shotpoint interval of 7-8 seconds (~13-15 m at 3.5-3.8 knots). The receiver was a 10 m solid  
144 state streamer with 10 channels towed 1-1.5 m beneath the surface. The near offset distance was 40 m. Data

145 were processed using Schlumberger Vista v.7 software following conventional seismic processing sequences,  
146 including quality control checks, amplitude recovery, bandpass filtering, 10 trace common shot domain  
147 stacking, signal enhancement using FX deconvolution (Gales et al., 2017).

148 Oceanographic data were collected by an Acoustic Doppler Current Profiler (ADCP), lowered (L-) ADCP,  
149 expendable BathyThermograph (XBT) and Conductivity-Temperature-Depth (CTD) SBE911plus probe which  
150 was equipped with optical backscattering and fluorescence sensors. An RDI Ocean Surveyor 75 kHz vessel  
151 mounted ADCP was configured with a 30° beam angle (4 beam phased array) with a maximum ping rate of 75  
152 kHz and maximum operational range of 550 m to record horizontal water velocity. 20 XBT (Sippican  
153 Lockheed) probes were launched at the Hillary Canyon head. The XBTs had a sampling rate of 10 Hz, vertical  
154 resolution of 60 cm and temperature resolution of 0.01°C. Eight CTD casts recorded at a sampling frequency of  
155 24 Hz and were averaged into 1 dbar bins. A Workhorse Sentinel Lowered-ADCP with a frequency of 300 kHz  
156 was attached to the water sampler frame.

157 Gully parameters were extracted from along-slope profiles 50 m below the shelf edge. Measured parameters  
158 included gully length (distance that gully can be traced down-slope), gully width (distance between points of  
159 maximum curvature of gully flanks), gully relief (vertical distance from maximum gully incision to line defining  
160 gully width), gully steepness (ratio between gully depth and width), gully sinuosity (measure of ratio of channel  
161 length vs straight-line distance of gully start to end) and cross-sectional shape to determine whether the profile is  
162 U-shape (parabolic) or V-shaped, measured using the General Power Law programme (SM. Fig.1; Pattyn and  
163 Van Huele, 1998). Paleo-gully parameters including width and relief were extracted from seismic line  
164 IT17RS305B. Relief estimates were based on a constant velocity estimation of 1650 m/s taken from  
165 compressional P-wave velocity measurements from borehole U1523 located on the upper Ross Sea slope  
166 (McKay et al., 2019).

167 Grain size analysis was used to assess surficial sediment on the upper slope of the Eastern Ross Sea. Grain size  
168 measurements were carried out on box core RS14-BC3 (40 cm recovery) and gravity core RS14-C3 (275 cm  
169 recovery) collected in 2014 on R/V *Italica* in the frame of the RNRA ROSSLOPE-2 Project on the upper slope  
170 of the Eastern Ross Sea in 1214 m and 1215 m water depth respectively (Fig. 1). Samples were prepared in 1 cm  
171 thick intervals and treated with hydrogen peroxide to remove organic matter. Wet sieving separated gravel (>  
172 2mm) and a Malvern Particle Size analyser was used with the addition of sodium polyphosphate (for dispersion)  
173 for the fine grain size analysis to obtain volume measurements of sand-silt-clay where sand (63 µm - 2 mm), silt

174 (2  $\mu\text{m}$  - 63  $\mu\text{m}$ ) and clay (< 2  $\mu\text{m}$ ) grain size boundaries are used. The maximum grain size able to be measured  
175 by the Malvern particle size analyser is 2 mm, therefore a clast count was also conducted to assess the coarse  
176 fraction (>2 mm) using wet-separation of the gravel (>2mm) component and X-rays of the cores.

## 177 **5. Results**

### 178 **5.1. Shelf-edge and canyon morphology**

179 The Hillary canyon head spans ~85 km along the shelf edge of the Eastern Ross Sea. The outer shelf is  
180 predominantly landward-deepening at the mouth of the Glomar Challenger Basin (from ~175°5'W to  
181 ~176°45'W). In the Pennell Basin, most of the shelf deepens landward, with the very outer shelf sloping  
182 seaward due to the presence of a sill at the mouth of the trough (west of ~176°45'W). The continental slope at  
183 the Glomar Challenger Basin mouth is characterised by an average slope gradient of ~3.5° with concave-upward  
184 profile to ~1500 m depth (d-d' in Fig. 1). Below this the canyon thalweg is relatively linear. At the mouth of the  
185 Pennell Trough, slope gradient increases to ~4.5°. To the east, the Eastern Ross Sea Fan forms a convex-  
186 outward depocenter seaward of the Glomar Challenger Basin with slopes <3°. Although data is limited here,  
187 widely spaced gullies occur on the outer-shelf and upper-slope of the fan. On the western fan flank, a small  
188 slide-scar (~18 km<sup>2</sup>) occurs that is ~11 km long, with a relief of ~30 m. Small blocks, ranging between 0.6 and 6  
189 km in length, occur within the slide scar (Fig. 2). Oblique seismic profiles crossing the upper Hillary canyon  
190 (Fig. 2) and the lower part of the Ross Sea Fan western flank, show chaotic seismic units, with discontinuous,  
191 sub-horizontal reflectors onlapping the paleo-continental slope (Fig. 2C). Convex-shaped and opaque lenses lie  
192 at the base of the continental slope (Fig. 2C).

193 At the Glomar Challenger Basin mouth (~175°47'W), the shelf edge is incised by deep gully systems, covering  
194 ~55 km along-slope. The 49 gullies that incise the shelf edge are relatively straight, narrow and mostly non-  
195 branching, with average reliefs of 40 m, widths of 0.85 km and lengths of 4.5 km (Fig. 3, 4). The gullies have an  
196 average sinuosity of 1.3 (where 1 = straight and >1 indicates increasing sinuosity). The gullies have a cross-  
197 section shape of 0.9, indicating a V-shaped cross section (where values <1 indicate convex upward gully flanks  
198 and values >2 indicate a parabolic or box-shaped morphology; Pattyn and Van Huele, 1998). The gullies are  
199 predominantly wider, longer and steeper and are more widely spaced toward the Eastern Ross Sea Fan  
200 (175°30'W) where slope gradients are less steep (~3.5°). The gullies increase in frequency, density and steepness  
201 and decrease in width and length further westward where slope gradient increases to ~4.5°. A distinct lack of

202 gullies is observed at the mouth of the Pennell Trough (from 177°43'W) over ~30 km where the seafloor is  
203 relatively smooth and homogenous.

204 Sub-bottom profiler and single-channel seismic data collected parallel to the shelf edge show that the gullies  
205 incise the seafloor rather than being formed by aggrading gully levees (Fig. 3D, E). The gullies incise several  
206 parallel reflectors; many terminating above a strong parallel reflector we term R1 that shoals to sub-crop at the  
207 seabed toward the Pennell Trough (Fig. 3B). Beneath R1, numerous paleo-gullies are observed in the region of  
208 surface gullies, with no expression of paleo-gullies further westward at the Pennell Trough mouth (Fig. 3C).  
209 Most of the paleo-gullies have U-shaped cross-sections and occur in vertically stacked complexes, or as isolated  
210 occurrences. There are multiple generations of paleo-gullies that cut down from different horizons. Although  
211 some paleo-gullies follow the same pathway to surface gullies, incising in similar spatial locations but at greater  
212 depths, there are multiple occurrences of paleo-gullies initiating in regions absent of surface gullies. The seismic  
213 character of the paleo-gully fill is slightly brighter than the surrounding sediments, with a less chaotic nature. A  
214 series of smaller paleo-gullies occur along reflector R1, compared to the gullies incising the modern seafloor,  
215 with average reliefs of 28 m and widths of 0.24 km. Below R1, the paleo-gullies are significantly greater in  
216 relief compared to gullies incising the seafloor, with average reliefs of 154 m and widths of 0.65 km. No levees  
217 are observed in the seismic data indicating that the paleo-gullies are incisional. Further west, the subsurface is  
218 characterised by lens-shaped semi-transparent packages that pinch out toward the axis of the canyon (Fig. 3B).

219 The gullies converge with shallow channels around 1000 m water depth (Fig. 3A). The channels are U-shaped in  
220 cross-section and are wider (average 2.3 km wide) and shallower (average 13 m relief) than the shelf-edge  
221 gullies. The channels converge towards the main canyon axis which narrows and deepens down-slope, with a  
222 relief of ~400 m and width of ~45 km at 2400 m water depth (Fig. 1C). Seismic data show that the canyon here  
223 is deeply entrenched with large levees either side of the canyon thalweg (Fig. 5). The levees are characterised by  
224 continuous horizontal reflectors and are asymmetric with the eastern levee significantly larger with onlapping  
225 and depositional strata. The flanks of the main canyon thalweg are characterised by gullies and small-scale  
226 mass-wasting (Fig. 5).

## 227 **5.2. Oceanographic observations**

228 Oceanographic observations at the Hillary Canyon head, collected in austral summer 2017 over a 3-week  
229 interval, show sustained high-velocity flows of cold, dense, bottom currents with velocities reaching ~1.0 m/s  
230 (Fig. 6). Over this interval, the highest velocity (~1.1 m/s) and coldest flows (~-1.5°C) were observed to the

231 west of the Hillary Canyon head, at the mouth of the Pennell Trough, in a region of homogenous seafloor absent  
232 of gullies. Although the flow may be quite variable in time and space, the predominant direction of the currents  
233 during the observed interval was off the continental shelf, toward the canyon axis (Fig. 6C), in agreement with  
234 previous measurements showing energetic bottom currents present all year round (Bergamasco et al., 2002;  
235 Budillon et al., 2011). Further east, toward the mouth of the Glomar Challenger Basin, shipboard ADCP data  
236 show a decrease in current velocity, although direction remains predominantly constant off the continental shelf  
237 (Fig. 6B, C). Here, XBT profiles show temperature is slightly higher ( $\sim 0.5^{\circ}\text{C}$ ) near seafloor compared to the  
238 Pennell Trough mouth (Fig. 6A).

239 The CTD probe data (Fig. 7A) show the vertical structure of the water column west of the Hillary Canyon head.  
240 The profiles show the presence of cold ( $\sim -0.8^{\circ}\text{C}$ ), low salinity ( $\sim 34.1$  PSU) water near the surface with high  
241 levels of turbidity ( $\sim 0.14$  NTU) and fluorescence ( $\sim 1.5$   $\text{mg}/\text{m}^3$ ). Warm and saline waters lay beneath, in the  
242 intermediate layers. Near the seafloor, salinity decreases slightly, and temperature decreases significantly to  
243  $\sim 34.6$  PSU and  $\sim -1.3^{\circ}\text{C}$  showing the signature of Ice Shelf Water. Fluorescence remains very low at the seafloor  
244 ( $< 0.5$   $\text{mg}/\text{m}^3$ ). In all CTD casts, turbidity values increase toward the seafloor to  $\sim 0.12 - 0.13$  NTU. L-ADCP  
245 transects show a significant increase in current velocity toward the seafloor at stations 2, 5 and 6, with velocity  
246 nearing  $1.1$  m/s (Fig. 7B).

### 247 **5.3. Grain size measurements**

248 Grain size measurements from box core RS14-BC3 and gravity core RS14-C3 from the Ross Sea mid slope at  
249  $\sim 1214$  m water depth (Fig. 1) show surficial (upper 6 cm RS14-BC3; upper 25cm RS14-C3) sediments of sandy  
250 silt with gravel clasts and silty clay dominating beneath (7-37 cm RS14-BC3; 29-271 cm RS14-C3; SM.Fig.2).  
251 A sandy interval occurs between 44-45 cm of RS14-C3 (43% sand). Ice rafted debris, as indicated by gravel  
252 grains ( $> 2\text{mm}$ ), decrease in the top of core RS14-BC3 to 4 cm before increasing to peak at 8 cm (11 counts), 17  
253 cm (20 counts) and 37 cm (20 counts) core depth.

## 254 **6. Discussion**

255 In this section we discuss new geomorphic, sub-surface and oceanographic results from the Hillary canyon in  
256 relation to the glacial, oceanographic and sedimentary processes influencing canyon morphology. We discuss  
257 the influence of dense shelf water on gully morphology more widely and compare characteristics of submarine  
258 gullies observed on other high-latitude continental margins.

259 **6.1. Influence of grounded ice extent on continental slope morphology**

260 The distribution of gullies incising the seafloor along the Eastern Ross Sea margin correspond with the  
261 maximum extent of grounded ice during the Last Glacial Maximum (Fig. 8). Geomorphological and  
262 sedimentological reconstructions show that ice advanced across the Eastern Ross Sea shelf during glacial  
263 periods, although timing and extent was not homogenous (e.g. Shipp et al., 1999; Bart and Owolana, 2012;  
264 Anderson et al., 2014; Halberstadt et al., 2016; Prothro et al., 2020). At the mouth of the Glomar Challenger  
265 Basin, large, V-shaped and deeply incised gullies occur. Here, multibeam echosounder data show the presence  
266 of well-defined megascale glacial lineations occurring up to near the shelf edge indicating that ice streams  
267 reached the shelf edge (Shipp et al., 1999; Halberstadt et al., 2016). By contrast, a distinct lack of gullies is  
268 observed at the mouth of the Pennell Trough over ~30 km along-slope. The presence of a large grounding zone  
269 wedge occurs 120 km landward of shelf edge and north of this there are no well-defined megascale glacial  
270 lineations suggesting that ice did not extend to the shelf edge here since the Last Glacial Maximum (although  
271 multibeam echosounder data here are very sparse compared to the Glomar Challenger Basin) (Fig. 8;  
272 Halberstadt et al., 2016). This configuration is also suggested by age and sedimentology of available sediment  
273 cores (Prothro et al., 2018; 2020). A single multibeam swath shows megascale glacial lineations on the  
274 northwestern flank of the bank dividing the Pennell Trough from the Glomar Challenger Basin indicating  
275 grounded ice flow parallel to the Pennell Trough axis (Prothro et al., 2018; 2020). These lineations do not reach  
276 the shelf edge but terminate against the morainal ridge lying at the shelf edge and closing the Pennell Trough  
277 mouth, suggesting that grounded ice did not reach the shelf edge or that glacial lineations have been completely  
278 reworked by iceberg keels after ice retreat (Fig. 8).

279 The extent of grounded ice across the continental shelf impacts slope processes and thus gully incision.  
280 Grounded ice transports large volumes of unsorted sediments to its margins particularly through ice streams  
281 (Golledge et al., 2011). At the Glomar Challenger Basin mouth, seismic data show a prograded continental  
282 margin (Fig. 2B, C) providing evidence for a prolonged history of ice advance and sediment delivery to the shelf  
283 edge over much of the Neogene and Quaternary (Cooper et al., 1991; De Santis et al., 1995, Anderson et al.,  
284 2019). Sediment failures are observed in recent and older strata along the continental slope (Fig. 2B, C). Along  
285 the margin at the mouth of the Glomar Challenger Basin, chaotic seismic units are observed with discontinuous,  
286 subhorizontal reflectors onlapping the paleo-continental slope, suggesting that sliding processes occurred during  
287 past margin progradation, or between periods of margin build out (Fig. 2C). Convex-shaped, opaque, lenses at

288 the base of the continental slope indicate repeated episodes of mass transfer gravity flows and channel-levee  
289 turbiditic construction (Laberg and Vorren, 1995) (Fig. 2C). Small slide scars also occur on the present day  
290 seabed on the western flank of the Eastern Ross Sea Fan, with small blocks interpreted as slide blocks or relict  
291 seafloor within the slide scar (Fig. 2). Although these slides may have been triggered following LGM and  
292 previous glacial maxima, the slides were likely influenced by ice loading. Grounded ice can lead to rapid  
293 deposition of sediment causing under-compaction of the slope and increase in pore pressure; formation of weak  
294 layers, where failure occurs within weaker interglacial substrate sediments (e.g. contouritic, plumitic,  
295 hemipelagic, diatomaceous layers) with excess pore water pressure caused by loading by stiffer glacial  
296 sediments rapidly deposited during glacial conditions and loading / unloading of ice at the shelf break (e.g.  
297 Melles and Kuhn, 1993; Dugan and Flemings, 2000; Laberg & Vorren 2000; Long et al., 2003; Volpi et al.,  
298 2003; Maslin et al., 2004; Donda et al., 2008). These slope failures may result in sediment and water mixing to  
299 generate turbidity currents explaining erosion further down-slope e.g. where debris flows transition to erosive  
300 turbidity currents (Fig. 9B) Piper et al., 1999; Talling et al., 2014).

301 Ice sheet grounding affects the structure and physical properties of the shelf edge and controls sediment  
302 distribution and associated margin architectural changes that influence the occurrence of mass-wasting  
303 processes (Vorren et al., 1998; Long et al., 2003). The high-volume of sediment delivery associated with  
304 advancing grounded ice can precondition the slope to down-slope processes such as turbidity currents which can  
305 cause seafloor erosion (Damuth et al., 1978; Vorren et al., 1998; Laberg and Vorren, 2000). Moreover, sediment  
306 supplied at the mouth of glacial troughs tends to have lower shear strength as a consequence of shear  
307 remoulding, under-compaction and sorting produced by meltwater, thus facilitating mass movements and  
308 subsequently a lower gradient of the continental slope (Rebesco et al., 1998). In addition, not only the amount of  
309 sediment delivery, but also its distribution and subsequent margin architectural changes (associated with  
310 changes in the glacial thermal regimes) control the occurrence of failure episodes and mass-wasting processes  
311 (Diviacco et al., 2006; Rebesco and Camerlenghi, 2008). The importance of slope preconditioning has been  
312 observed in other glacial and non-glacial environments where sediment availability has been shown to be a  
313 crucial factor in processes such as turbidity current generation where dilute flows can mobilise accumulations of  
314 fine sediments (e.g. Vorren et al., 1998; Hage et al., 2019; Normandeou et al., 2019; Pope et al., 2019).

315 Grounded ice can influence the availability of subglacial and ice front meltwater (discussed in section 6.3).

316 These effects are modified by specific characteristics of the study area, such as the underlying geology which

317 affects seabed erodibility, deformation and thus ice flow, slope geometry and gradient, glacial history, drainage  
318 basin size and climate history (e.g. Rebesco et al., 1998; Ó Cofaigh et al., 2003; Wynn et al., 2012).

## 319 **6.2. Can dense shelf water cause gully erosion?**

320 Oceanographic observations show that over a 3-week period, cold, dense waters move toward the canyon head,  
321 reaching speeds of ~1 m/s near to the seafloor. Gullies are absent from the region of most energetic cascading  
322 flows. This follows previous in-situ observations (e.g. CLIMA project; Bergamasco et al., 2002; Petrelli et al.,  
323 2008) from the shelf break and numerical modelling studies that show the Hillary Canyon to be a major region  
324 of cascading dense shelf water with waters driven to the west of the canyon by Coriolis (Fig. 6; Budillon et al.,  
325 2011; Morrison et al., 2020). We use bed shear stress ( $\tau\theta$ ; equation [1]) to assess the current's ability to erode  
326 the seafloor, where  $C_d$  is the friction factor,  $\rho$  is water density and  $u$  is depth averaged flow speed (SM.Table 1).

$$327 \quad \tau\theta = C_d\rho u^2 \quad (1)$$

328 Under the observed modern austral summer conditions, bottom currents are great enough to initiate sediment  
329 motion. Calculations of bed shear stresses associated with observed currents ( $\tau\theta = 2.57$  Pa) show that these  
330 currents are able to erode marine muds (0.05-2 Pa) and sandy muds (0.1-1.5 Pa) (McCave, 1984; Jacobs et al.,  
331 2011) which dominate interglacial sediment supplied to the slope via iceberg rafting, along-slope currents,  
332 glaciomarine and hemipelagic, pelagic and biogenic settling (e.g. Vanneste and Larter, 1995; Rebesco et al.,  
333 1996). Sediment trap deployments on the outer Ross Sea shelf show that the modern vertical flux is dominated  
334 by faecal pellets and aggregate particles with equivalent stokes sphere sizes of 59 $\mu$ m-96 $\mu$ m (Jaeger et al., 1996).  
335 There is no information about seabed sediment on the steep upper slope where dense water flows reach highest  
336 velocity in present time. However, recent box (RS14-BC3) and gravity core (RS14-C3) grain size measurements  
337 from the Ross Sea mid slope (Fig. 1) show surficial interglacial sediments of sandy silt (SM.Fig. 2), from which  
338 the fine silt fraction component was likely removed and the coarse sediment (IRD) sorted by bottom currents.  
339 Considering that the observed along-slope and down-slope currents are fastest on the upper slope (Fig. 6), they  
340 are less able to erode glaciogenic sediments which have much higher critical shear stresses that form the bulk of  
341 the mid slope cores beneath surficial interglacial sediments (e.g. 2.2-23 Pa; Pike et al., 2018). The glaciogenic  
342 sediments are characterized by increased clay content, low sand content and increased Ice Rafted Debris (Lucchi  
343 et al., 2002; Fig 1; SM.Fig.2). At the shelf edge, the currents can initiate suspension of recent sediment, and thus

344 maintain gully morphology avoiding significant sediment infilling, however, are unlikely to erode glacial  
345 sediments present.

346 Our observations suggest that under modern conditions, erosion of gullies is not likely to be dominated by flows  
347 of cascading dense shelf water as gullies are absent from the region of most energetic and sustained dense shelf  
348 water overflow (Bergamasco et al., 2002; Morrison et al., 2020). Recent in-situ direct monitoring experiments of  
349 very dilute plumes in other ice-marginal settings show that dilute plumes can generate turbidity currents if the  
350 seafloor is sufficiently preconditioned e.g. if sediment is allowed to build up between resuspension events (Hage  
351 et al., 2019). The movement of sediment by dense water cascading differs from turbidity currents in that dense  
352 water cascading is gravity-driven, where the descending movement of dense water entrains sediments, as  
353 opposed to turbidity currents where it is the movement of loose sediment that entrains water movement (Canals  
354 et al., 2006). Although the ability for dense shelf water to entrain sediment and produce gravity flows is poorly  
355 understood (Canals et al., 2006; Talling et al., 2014), sustained accumulations and subsequent resuspension  
356 events of biogenic and fine-grained particles at the Hillary canyon head could condition the slope for potential  
357 sediment gravity flows. The Ross Sea has moderate rates of modern primary productivity in the upper 100 m of  
358 the water column (e.g. Jaeger et al., 1996; Smith et al., 1996). Under modern conditions, if sediments supplied  
359 to the outer shelf / upper slope were able to accumulate over prolonged periods of time, such as between major  
360 dense shelf water cascading events (e.g. via iceberg rafting, along-slope currents, glaciomarine, hemipelagic,  
361 pelagic and biogenic settling), movement of sediment and concentration may become sufficient to generate  
362 sediment gravity flows if triggered by some external mechanism e.g. earthquake or glacial earthquake, glacio-  
363 isostatic adjustment, wave loading, internal waves and tides, energetic tidal jets, landslides, energetic dense  
364 water cascade or release of sediment-laden subglacial meltwater (Palanques et al., 2009; Talling et al., 2014;  
365 Clare et al., 2016; Maier et al., 2019; Nettles and Ekström, 2010).

366 The Eastern Ross Sea is a passive margin, therefore less likely to be affected by tectonic influences such as  
367 earthquakes, although glaciotectonic processes may occur during glacial retreat (Lee and Philips et al., 2013).  
368 Most of the continent is permanently covered in ice and the canyon is located >300 km from the modern ice  
369 front, therefore sediment resuspension cannot be associated with river discharge. The shelf-edge currently lies at  
370 ~570 m water depth therefore is not directly affected by surface waves and storms, although internal tides and  
371 waves are known to occur throughout this region (Robertson et al., 2003) thus may become important if current  
372 velocity is enough to resuspend sediments (e.g. Maier et al., 2019). Intense meteorological events on glaciated

373 margins may control the production of dense shelf water plumes, which can resuspend significant amounts of  
374 sediment (Bensi et al., 2019). Where sediment can build up between resuspension events, exceptionally strong  
375 and persistent episodes of dense shelf water cascading could increase sediment movement and concentration  
376 enough to produce sediment gravity flows down slope. This could result in sediment transport down-canyon,  
377 although these low-density flows are unlikely to cause significant gully erosion (Fig. 9A).

378 Under full-glacial conditions, where grounded ice extended to the shelf edge, the formation of Ice Shelf Water  
379 and High Salinity Shelf Water directly over the continental shelf was likely impeded as no or very little sea ice  
380 was able to form on the continental shelf due to increased grounded ice extent. This limits the delivery of salt  
381 required for dense water formation, minimising dense shelf water developing and cascading over the shelf edge.  
382 Sea ice and brine rejection could, however, continue to form seaward of the ice margin over the continental  
383 slope. Following glacial retreat from the shelf edge (e.g. during deglaciation), dense shelf water cascading likely  
384 resumes due to sea ice reforming over the continental shelf and Ice Shelf Water forming under ice shelf cavities.  
385 The vigour of these cascading flows during glacial / deglacial periods are currently not known.

### 386 **6.3. Can glacial meltwater cause gully erosion?**

387 The pattern of gullying observed along the Ross Sea margin is consistent with ice grounding near to the shelf  
388 edge, as suggested by the occurrence of glacial facies in available sediment cores (Prothro et al., 2018; 2020).  
389 Deeply incised and V-shaped gullies are present at the Glomar Challenger Basin mouth, yet absent from the  
390 Pennell Trough mouth (Fig. 3; Fig. 8). The observation of a V-shaped, deeply incised and sinuous thalweg,  
391 suggests that the formation of gullies is dominated by fluid flow rather than generation by mechanisms such as  
392 slope failure which are likely to form escarpments and lack of well-defined thalweg (e.g. Kenyon, 1987; Simons  
393 and Senturk, 1992). Grounded ice can influence the availability of subglacial and ice front meltwater which can  
394 become sediment-laden and entrained into sediment-laden gravity flows down slope causing erosion of the  
395 seafloor (Anderson, 1999; Noormets et al., 2009; Ó Cofaigh et al., 2003). Subglacial meltwater is formed by  
396 processes such as geothermal heat fluxes, friction, strain heating and surface melting as shown by the presence  
397 of paleo-subglacial drainage systems such as subglacial lakes and meltwater channels on the continental shelf  
398 (e.g. Sugden et al., 1991; Simkins et al., 2017). Though surface melting is minimal in Antarctica at present day,  
399 it may be effective in certain areas and contribute to slope instabilities within glacier terminus systems (Rebesco  
400 et al, 2014).

401 At the mouth of Glomar Challenger Basin, gullies are greater in relief, length, width and sinuosity on the eastern  
402 side of the basin mouth, where upper slope gradients are  $\sim 3.5^\circ$  compared to smaller, straighter and higher  
403 density gullies to the west of the basin mouth where slope gradients increase to  $\sim 4.5^\circ$  (Fig. 3, 4). The small  
404 change in slope gradient across the continental slope likely reflects the proximity of the eastern part of the slope  
405 to the Eastern Ross Sea Trough Mouth Fan, which is characterised by outward bulging contours formed by  
406 margin progradation. The increase in slope gradient to the west is likely to cause higher velocities of down-slope  
407 flows, leading to greater flow confinement, narrow and straight gullies (Wynn et al., 2012). The differences in  
408 gully morphology may also reflect greater availability, or focussing, of sediment-laden subglacial meltwater to  
409 the east of Glomar Challenger Basin mouth, on the flank of the Eastern Ross Sea Fan (Rebesco et al., 1998;  
410 Noormets et al., 2009; Fig. 8). Meltwater may be drawn toward trough margins due to differences in subglacial  
411 water pressure gradients and additional frictional heat at trough margins caused by faster flowing ice streams,  
412 compared to slower moving ice on adjacent banks (Rothlisberger, 1972; Noormets et al., 2009; Boulton et al.,  
413 2007).

414 Meltwater can also form when intrusions of CDW reach the ice sheet front, causing rapid melting of marine  
415 terminating ice shelves and tidewater cliffs (Silvano et al., 2018). Sub-ice-shelf melt caused by intruding warm  
416 currents can cause energetic and turbulent plumes of very cold, fresh water to rise at the ice calving face due to  
417 reversed buoyancy differences (Jenkins et al., 2011; Truffer and Motyka, 2016). Similarly, buoyant plumes that  
418 track the underside of floating ice can also emerge from point-source meltwater channels under ice shelves  
419 (Carter and Fricker, 2012; Jenkins et al., 2011). Turbulent mixing drives convective motion at the ice-ocean  
420 interface which may cause sediment resuspension if the turbulence caused by the buoyancy effects of rising  
421 meltwater jets are great enough to resuspend recently deposited sediments at the shelf edge (Heese and  
422 Khodabaksh, 2006; Jenkins et al., 2011).

423 Where ice is grounded at the shelf edge, slope gradients may be great enough to allow sediment-laden gravity  
424 flows to evolve down slope. Both mechanisms of erosion by sediment-laden gravity flows can explain the  
425 pattern on deep gully incision at the Glomar Challenger Basin mouth, where ice grounded near to the shelf edge,  
426 and absence of gullies from the Pennell Trough, where ice grounded  $>120$  km from the shelf edge (Fig. 8; Fig.  
427 9B).

#### 428 **6.4. Can intruding warm currents erode the shelf edge?**

429 The region of greatest gully size and sinuosity (Fig. 3A, 4) occurs where numerical modelling studies show  
430 CDW intruding onto the shelf to the east of Glomar Challenger Basin (Morrison et al., 2020). Smaller and  
431 higher density gullies occur further west where CDW intrusions are reduced and dense shelf water cascading is  
432 greatest (Bergamasco et al., 2002; Morisson et al., 2020). Recent modelling shows a strong dynamic relationship  
433 between CDW and dense shelf water overflow within the Hillary Canyon with spatial distribution due to  
434 Coriolis driving dense shelf waters toward the west of the canyon (Morrison et al., 2020). As dense shelf water  
435 descends, less dense water is displaced within the canyon causing sea level height to decrease resulting in a  
436 barotropic pressure gradient that drives CDW shelfward (Morrison et al., 2020). Direct observations of CDW  
437 are absent at this location presently, however, simulated CDW velocities are shown to reach  $0.25 \text{ ms}^{-1}$  (Morrison  
438 et al., 2020). We suggest that intruding flows of CDW do not exert significant influence on gully morphology as  
439 comparisons with modelled and direct observations of cascading dense shelf water velocities reaching  $1 \text{ ms}^{-1}$   
440 are shown to be too weak to incise glacial sediments at the shelf edge (Gordon et al., 2009; Morrison et al.,  
441 2020). The motion field in the areas surrounding CDW intrusions may allow recent fine-grained sediments to be  
442 resuspended and allow gully morphology to be maintained at the shelf edge, however are not strong enough to  
443 permit significant seafloor erosion here.

444 The spatial distribution and dynamic relationship between CDW intrusions and dense shelf water also acted  
445 during past glacial and interglacial periods, driven by Coriolis and the equilibrium between water mass density.  
446 When the ice sheet was grounded at the shelf edge, formation of dense shelf water over the continental shelf was  
447 likely reduced due to limited opportunity for sea ice formation, although dense water formation was likely to  
448 have continued seaward of the ice margin over the continental slope. It follows that dense shelf water cascading  
449 also occurred along the western flank of the Glomar Challenger Basin with intrusions of CDW likely across the  
450 eastern flank of the Glomar Challenger Basin. The incursion of CDW may have triggered the initial ice retreat  
451 and caused the release of a large amount of meltwater. This may explain the large size of the gullies compared  
452 to the western sector.

### 453 **6.5. Paleo-gully formation**

454 New seismic data show paleo-gullies beneath gullies incising the modern seafloor (Fig. 3C), indicating that this  
455 region has been influenced by gully-forming processes over sustained periods of time. The surface gullies  
456 terminate above a strong parallel reflector, R1, that shoals toward the seafloor at the Pennell Trough, where both  
457 surface and paleo-gullies are absent. Reflector R1 is characterised by small (average relief of  $\sim 28 \text{ m}$  and width

458 of 0.24 km) gullies occurring in similar frequency across the shelf edge to gullies incising the modern seafloor.  
459 Gullies incising R1 do not follow the same spatial distributions to the surface gullies, with gullies located both  
460 directly beneath, and in regions absent of surface gullies (Fig. 3C). The small size of the gullies incising R1  
461 compared to the modern seafloor implies formation by minimally erosive flows, or flows occurring over shorter  
462 time-scales before being abruptly filled (Shumaker et al., 2017). Reflector R1 may indicate a major climatic  
463 event such as major ice advance across the continental shelf; although, further evidence from sediment cores is  
464 required to constrain the age and magnitude of this transition further.

465 The paleo-gullies beneath R1 occur as both single gullies and in complexes in the same region as the surface  
466 gullies. Similarly to gullies incising R1, the paleo-gullies do not follow the same spatial distribution as gullies  
467 incising the modern seafloor, or those incising R1 (Fig. 3C). The size of the paleo-gullies beneath R1 are  
468 significantly greater in relief than the most recent seafloor-incising gullies indicating that gully-forming events  
469 were likely larger in the past or the processes that formed them were sustained over longer time periods, for  
470 example due to greater volumes of sediment and/or meltwater availability (Hayes et al., 1975). The size and  
471 stacked nature of the gullies implies that the shelf edge underwent periods of inactivity and partial infilling  
472 before gully incision restarted suggesting that a complex period of gully fill and erosion occurred (Shumaker et  
473 al., 2017). This suggests that gully activity was continuous over multiple glacial cycles and formed an important  
474 role in the development of the shelf edge and upper slope, with incision likely related to duration and persistence  
475 of ice-sheet residence.

476 The lack of gully spatial continuity with depth indicates that there is no preferential route or passage that  
477 processes, such as sediment laden subglacial meltwater released from beneath an ice sheet, or meltwater  
478 generated by warm intruding currents, follow on the continental slope. This suggests that ice sheet configuration  
479 varied at the shelf edge over multiple glacial cycles. As paleo-gullies are only observed at the mouth of the  
480 Glomar Challenger Basin, this may provide further evidence for sustained ice advance and retreat compared to a  
481 much more limited extent in the Pennell Trough.

#### 482 **6.6. How do gullies compare to other high-latitude continental margins?**

483 We put slope morphology observed at the Hillary Canyon head into global perspective to understand if  
484 cascading flows of dense shelf water exert consistent patterns of erosion on high-latitude continental margins.  
485 Analysis of gully morphometric parameters over ~2300 km of the Antarctic continental margin, including the  
486 western Antarctic Peninsula, Bellingshausen, Amundsen, Weddell and Ross seas show that gullies at the Glomar

487 Challenger Basin mouth are significantly different in morphology to gullies observed in other areas of dense  
488 shelf water cascading (Fig. 10). The Weddell Sea is a region of significant dense shelf water production,  
489 contributing to ~50% global Antarctic Bottom Water production (Budillon et al., 2011). Gullies at the Filchner  
490 Trough mouth, Weddell Sea, are smaller in relief, length and width, U-shaped in cross-section with low  
491 sinuosity and found on lower slope gradients ( $\sim 2.5^\circ$ ) (Gales et al., 2012). The significant difference in gully  
492 morphology between two globally important regions of dense shelf water cascading (Fig. 10), shows that these  
493 cascading flows do not exert consistent patterns of erosion on high-latitude continental margins under modern  
494 day conditions. It is likely that this pattern of erosion has greater dependence on other factors, such as glacier  
495 drainage basin size, regional oceanography, meltwater availability, glacial extent and polynya activity rather  
496 than the cascading flows themselves causing gully incision during interglacial conditions.

497 Analogous gullies to those observed at the Glomar Challenger Basin mouth form in regions unaffected by  
498 modern cascading flows of dense shelf water e.g. along the western Antarctic Peninsula margin and some  
499 Arctic margins e.g. Kongsfjorden Trough, Bear Island Trough and Adfjorden Trough (Noormets et al., 2009;  
500 Gales et al., 2013; Rui et al., 2019; Post et al., 2020). These gullies are characterised by V-shaped cross-  
501 sections, low gully lengths, medium sinuosities and similar reliefs and are found on medium slope gradients of  
502  $\sim 6-9^\circ$  both within and outside cross-shelf trough confines (Gales et al., 2013). Due to the V-shaped and deeply  
503 incised morphologies, these gullies are suggested to form by processes dominated by suspended sediment load  
504 such as release of sediment-laden subglacial meltwater (e.g. Simons and Senturk, 1992). This is consistent with  
505 analysis of gullies observed along the Glomar Challenger Basin margin which suggest gullies formed by gravity  
506 flows generated by resuspension of sediment by glacial meltwater processes.

## 507 7. Conclusions

508 Shelf-slope processes and climatic variations can have significant influence on seafloor morphology, especially  
509 in Polar regions, where climate, ice sheet and sea level changes play a crucial role (Fig. 9). New geophysical and  
510 oceanographic data show that the Hillary Canyon is the main conduit for cascading flows of dense shelf water to  
511 the abyss with canyon levees likely formed of overbank deposits indicating a prolonged history of down-slope  
512 flows. Incisional gullies occur at the canyon head, which merge into channels and into the main canyon thalweg.  
513 The distribution of gullies along the shelf edge is not homogenous with a distinct absence of gullies in a region  
514 of intense modern dense shelf water export. The gullies correspond with maximum ice extent at the Ross Sea  
515 shelf edge indicating gully incision was likely controlled by glacial advance. At the Pennell Trough mouth,

516 gullies are absent as the maximum extent of grounded ice was >120 km inland of the shelf-edge. Glacial  
517 advance likely preconditioned the slope to down-slope processes such as release of sediment laden subglacial  
518 meltwater from the ice sheet terminus.

519 Our results suggest that cascading flows of dense shelf water are strong enough to prevent gully infilling and  
520 contribute to canyon-levee aggradation down-slope; however, do not contribute to significant gully incision at  
521 the shelf edge. During full-glacial conditions, where ice grounded at the shelf edge, the formation and cascading  
522 of dense shelf water was limited by lack of continental shelf polynya. Sediment laden subglacial meltwater  
523 released from beneath an ice sheet or meltwater generated by warm intruding currents that resuspends sediment  
524 may produce sediment gravity flows where slope gradients are great enough for flow ignition (e.g. ice front  
525 located at the shelf break). Understanding both modern and past processes influencing canyon morphology has  
526 important implications for reconstructing past ice sheet behaviour from sediment cores. Better definition of  
527 factors influencing seafloor morphology will assist future numerical modelling studies in predicting the dynamic  
528 behaviour of processes that influence changes to ice-shelf and sea level variations and applies to canyons around  
529 Antarctica more widely. These findings raise important questions concerning factors controlling ice advance,  
530 duration and persistence of ice-sheet residence at the shelf edge and what previously limited ice advance in  
531 Pennell Trough. Our findings show that gullies were absent in this region over multiple glacial cycles, perhaps  
532 indicating a stronger inflow of CDW here limited ice advance. The factors controlling the preferential locations  
533 of CDW intrusions are of utmost importance as major drivers of Antarctic ice sheet retreat.

534 Further sedimentological information is required to identify the timing and frequency of these slope processes  
535 and their reliance on climatic cycles more widely over previous millennia. This will help constrain links  
536 between paleo-ice sheet dynamics, paleo-oceanographic processes and slope processes which is crucial for  
537 understanding future changes.

## 538 **8. Acknowledgements**

539 This work was supported by the EUROFLEETS Funding Program (ANTSSS project) and Italian National  
540 Antarctic Research Program (PNRA16 00205; ODYSSEA) projects. The authors thank Captain Franco Sedmak  
541 and crew of R/V OGS Explora; party chief Riccardo Codiglia; the technicians and scientific party of the  
542 expedition; We thank Laura de Steur for supplying the L-ADCP during the ANTSSS expedition and for  
543 processing of L-ADCP data. We acknowledge IHS Markit, Paradigm, Schlumberger for Kingdom and Vista

544 academic license to OGS. RM was funded by the Royal Society Te Apārangi Marsden Fund award MFP-  
545 VUW1808.

546 **9. References**

547 Alonso, B., Anderson, J.B., Diaz, J.T., Bartek, L.R. 1992. Pliocene-Pleistocene seismic stratigraphy of the Ross  
548 Sea: evidence of multiple ice sheet grounding episodes. In Contributions to Antarctic Research III, Antarctic  
549 Research Series 57, American Geophysical Union, Washington D.C, 93-103.

550 Anderson, J. 1999. Antarctic Marine Geology. Cambridge University Press.

551 Anderson, J., Bartek, L. 1990. Preliminary results of high-resolution seismic reflection survey of the Ross Sea  
552 continental shelf. Antarctic Journal of the US 25, 61-63.

553 Anderson, J., Conway, H., Bart, P., Witus, A., Greenwood, S., McKay, R., Hall, B., Ackert, R., Licht, K.,  
554 Jakobsson, M., Stone, J. 2014. Ross Sea paleo-ice sheet drainage and deglacial history during and since the  
555 LGM. Quaternary Science Reviews 100, 31-54.

556 Anderson, J., Simkins, L., Bart, P., De Santis, L., Halberstadt, A., Olivo, E., Greenwood, S. 2019. Seismic and  
557 geomorphic records of Antarctic ice sheet evolution in the Ross Sea and controlling factors in its behaviour.  
558 Geological Society, London, Special Publications 475, 223-240.

559 Arndt, J., Schenke, H.W., Jakobsson, M., Nitsche, F.O., Buys, G., Goleby, B., Rebesco, M., Bohoyo, F., Hong,  
560 J.K., Black, J., Greku, R., Udintsev, G., Barrios, F., Reynoso-Peralta, W., Taisei, M., Wigley, R. 2013. The  
561 International Bathymetric Chart of the Southern Ocean (IBCSO) Version 1.0. PANGAEA,  
562 doi.org/10.1594/PANGAEA.805736.

563 Barker, P., Camerlenghi, A. 2002. Glacial history of the Antarctic Peninsula from Pacific margin sediments.  
564 Proceedings of the Ocean Drilling Program: Scientific Results 178, 1-40.

565 Bart, P.J., Anderson, J.B. 2000. Relative stability of the Antarctic ice sheets during the Plio-Pleistocene from the  
566 perspective of the continental shelf. Earth and Planetary Science Letters 182, 259–272.

567 Bart, P.J., Owolana, B. 2012. On the duration of West Antarctic Ice Sheet grounding events in Ross Sea during  
568 the Quaternary. Quaternary Science Reviews 47, 101-115.

569 Bensi, M., Kovacevic, V., Langone, L., Aliani, S., Ursella, L., Goszczko, I., Solwedel, T., Skogseth, R., Nilsen,  
570 F., Deponte, D., Mansutti, P., Laterza, R., Rebesco, M., Rui, L., Lucchi, R.G., Wahlin, A., Viola, A.,  
571 Beszczynska-Moller, A., Rubino, A. 2019. Deep flow variability offshore South-west Svalbard (Fram Strait).  
572 *Water* 11, doi.org/10.3390/w11040683.

573 Bergamasco, A., Defendi, V., Zambianchi, E., Spezie, G. 2002. Evidence of dense water overflow on the Ross  
574 Sea shelf-break. *Antarctic Science* 14, 271-277.

575 Boulton, G., Lunn, R., Vidstrand, P., Zatsepin, S. 2007. Subglacial drainage by groundwater-channel coupling,  
576 and the origin of esker systems: Part 1 – glaciological observations. *Quaternary Science Reviews* 26, 1067-  
577 1090.

578 Brancolini, G., Cooper, A., Coren, F. 1995. Seismic facies and glacial history in the western Ross Sea,  
579 Antarctica. *Geology and Seismic Stratigraphy of the Antarctic Margin, Antarctic Research Series* 68, 209-233.

580 Budillon, G., Castagno, P., Aliani, S., Spezie, G., Padman, L. 2011. Thermohaline variability and Antarctic  
581 bottom water formation at the Ross Sea shelf break. *Deep Sea Research Part I: Oceanographic Research Papers*  
582 58, 1002-1018.

583 Canals, M., Puig, P., de Madron, X.D., Heussner, S., Palanques, A., Fabres, J. 2006. Flushing submarine  
584 canyons. *Nature* 444, 354-357.

585 Carter, S.P., Fricker, H.A. 2012. The supply of subglacial meltwater to the grounding line of the Siple Coast,  
586 West Antarctica. *Annals of Glaciology* 53, 267-280.

587 Clare, M., Hughes Clarke, J., Talling, P., Cartigny, M., Pratomo, D. 2016. Preconditioning and triggering of  
588 offshore slope failures and turbidity currents revealed by most detailed monitoring yet at a fjord-head delta.  
589 *Earth and Planetary Science Letters* 450, 208–220.

590 Colleoni, F., De Santis, L., Siddoway, S., Bergamasco, A., Golledge, N.R., Lohmann, G., Passchier, S., Siegert,  
591 M.J. 2018. Spatio-temporal variability of processes across the Antarctic ice-bed-ocean interfaces. *Nature*  
592 *Communications* 9, doi.org/10.1038/s41467-018-04583-0.

593 Cooper, A.K., Barrett, P.J., Hinz, K., Traube, V., Leitchenkov, G., Stagg, H.M.J. 1991. Cenozoic prograding  
594 sequences of the Antarctic continental margin: a record of glacio-eustatic and tectonic events. *Marine Geology*  
595 102, 175–213.

596 Damuth, J. 1978. Echo character of the Norwegian-Greenland Sea: Relationship to Quaternary sedimentation.  
597 *Marine Geology* 28, 1-36.

598 DeConto, R.M., Pollard, D., 2016. Contribution of Antarctica to past and future sea-level rise. *Nature* 531, 591–  
599 597.

600 De Santis, L., Anderson, J.B., Brancolini, G., Zayatz, I. 1995. Seismic record of late Oligocene through Miocene  
601 glaciation on the Central and Eastern Continental Shelf of the Ross Sea. In: Cooper, A.K., Barker, P.F.,  
602 Brancolini, G. (eds). *Geology and Seismic Stratigraphy of the Antarctic Margin*. Antarctic Research Series 68,  
603 AGU Washington, DC, 235-260.

604 Dinniman, M.S., Klinck, J.M., Smith Jr, W.O. 2003. Cross-shelf exchange in a model of the Ross Sea  
605 circulation and biogeochemistry. *Deep Sea Research Part II: Topical Studies in Oceanography* 50, 3103-3120.

606 Diviacco, P., Rebesco, M., Camerlenghi, A. 2006. Late Pliocene mega debris flow deposit and related fluid  
607 escapes identified on the Antarctic Peninsula continental margin by seismic reflection data analysis. *Marine*  
608 *Geophysical Researches* 27, 1-19.

609 Donda, F., O'Brien, P.E., De Santis, L., Rebesco, M., Brancolini, G., 2008. Mass wasting processes in the  
610 Western Wilkes Land margin: possible implications for East Antarctic glacial history. *Palaeogeography,*  
611 *Palaeoclimatology, Palaeoecology* 260, 77-91.

612 Dugan, B., Flemings, P. W. 2000. Overpressure and fluid flow in the New Jersey continental slope: Implications  
613 for slope failure and cold seeps. *Science* 289, 288-291.

614 Gales, J.A., Forwick, M., Laberg, J.S., Vorren, T.O., Larter, R.D., Graham, A.G.C., Baeten, N., Amundsen,  
615 H.B. 2013. Arctic and Antarctic submarine gullies – a comparison of high latitude continental margins.  
616 *Geomorphology* 201, 449-461.

617 Gales, J.A., Larter, R.D., Mitchell, N.C., Hillenbrand, C.-D., Østerhus, S., Shoosmith, D. 2012. Southern  
618 Weddell Sea shelf edge geomorphology: Implications for cold, high salinity water overflow as an Antarctic

619 gully-forming mechanism. *Journal of Geophysical Research – Earth Surface* 117, F0421,  
620 doi.org/10.1029/2012JF002357.

621 Gales, J., Zgur, F., Accettella, D., Bergamasco, A., Codiglia, R., Colleoni, F., Cuffaro, M., DeSantis, L.,  
622 DiCurzio, E., Filippone, P., Gordini, E., Kim, S., Lopez, C., Kovacevic, V., Mansutti, P., Olivo, E., Rebesco,  
623 M., Sterzai, P., Tomini, I., Visnovic, G., Yanguang, L. 2017. Antarctic Ice Sheet Stability from continental  
624 slope processes investigation (ANTSSS). EUROFLEETS2 Cruise Summary Report, pp 61.

625 Galy, V., France-Lanord, C., Beyssac, O., Faure, P., Kudrass, H., Palhol, F. 2007. Efficient organic carbon  
626 burial in the Bengal fan sustained by the Himalayan erosional system. *Nature* 450, 407–410.

627 Golledge, N.R., Levy, R.H. 2011. Geometry and dynamics of an East Antarctic Ice Sheet outlet glacier, under  
628 past and present climates. *Journal of Geophysical Research - Earth Surface* 116, F3,  
629 doi.org/10.1029/2011JF002028.

630 Gordon, A.L., Orsi, A.H., Muench, R., Huber, B.A., Zambianchi, E., Visbeck, M. 2009. Western Ross Sea  
631 continental slope gravity currents. *Deep-Sea Research II* 56, 796-817.

632 Hage, S., Cartigny M.J.B., Sumner, E.J., Clare, M.A., Hughes Clarke, J.E., Talling, P.J., Lintern, D.G.,  
633 Simmons, S.M., Silva Jacinto, R., Vellinga, A.J., Allin, J.R., Azpiroz-Zabala, M., Gales, J.A., Hizzett, J., Hunt,  
634 J., Mozzato, A., Partson, D., Pope, E., Stacey, C., Symons, W., Vardy, M., Watts, C. 2019. Direct monitoring  
635 reveals initiation of turbidity currents from extremely dilute river plumes. *Geophysical Research Letters*,  
636 doi.org/10.1029/2019GL084526.

637 Halberstadt, A.R., Simkins, L.M., Greenwood, S.L., Anderson, J.B. 2016. Past ice sheet behavior: retreat  
638 scenarios and changing controls in the Ross Sea, Antarctica. *The Cryosphere* 10, 1003–1020.

639 Hayes, D.E., Frakes, L.A. 1975. General synthesis: Deep Sea Drilling Project 28. In: Hayes D.E., Frakes L.A. et  
640 al. (eds). *Initial Reports of the Deep Sea Drilling Project*. US Government Printing Office, Washington, DC 28,  
641 919–942.

642 Heese, R., Khodabakhsh, S. 2006. Significance of fine-grained sediment lofting from meltwater generated  
643 turbidity currents for the timing of glaciomarine sediment transport into the deep sea. *Sedimentary Geology* 186,  
644 1-11.

645 IPCC. 2019. Summary for Policymakers. In: IPCC Special Report on the Ocean and Cryosphere in a Changing  
646 Climate. Pörtner, H.-O., Roberts, D.C., Masson-Delmotte, V., Zhai, P., Tignor, M., Poloczanska, E.,  
647 Mintenbeck, K., Nicolai, M., Okem, A., Petzold, J., Rama, B., Weyer, N. (eds). In press.

648 Jacobs, S.S. 2004. Bottom water production and its links with the thermohaline circulation. *Antarctic Science*  
649 16, 427-437.

650 Jacobs, S.S., Fairbanks, R.G., Horibe, Y. 1985. Origin and evolution of water masses near the Antarctic  
651 continental margin: evidence from  $H_2^{18}O_2/H_2^{16}O_2$  ratios in seawater. *Antarctic Research Series* 43, 59–85.  
652

653 Jacobs, S.S., Giulivi, C.F., Mele, P.A. 2002. Freshening of the Ross Sea during the late 20th Century. *Science*  
654 297, 386–389.

655  
656 Jacobs, W., Le Hir, P., Van Kesteren, W., Can, P. 2011. Erosion threshold for sand-mud mixtures, *Continental*  
657 *Shelf Research* 31, S14– S2.

658 Jaeger, J.M., Nittrouer, C.A., Demaster, D.J., Kelchner, C., Dunbar, R.B. 1996. Lateral transport of settling  
659 particles in the Ross Sea and implications for the fate of biogenic material. *Journal of Geophysical Research*  
660 101, 18479–18488.

661 Jenkins, A. 2011. Convection driven melting near the grounding lines of ice shelves and tidewater glaciers.  
662 *Journal of Physical Oceanography* 41, doi.org/10.1175/JPO-D-11-03.1.

663 Joughin, I., Alley, R. B. 2011. Stability of the West Antarctic ice sheet in a warming world. *Nature Geoscience*,  
664 4, 506-513.

665 Kenyon, N.H. 1987. Mass-wasting features on the continental slope of northwest Europe. *Marine Geology* 74,  
666 57-77.

667 Laberg, J.S., Vorren, T.O. 1995. Late Weichselian submarine debris flows deposits on the Bear Island Trough  
668 Mouth Fan. *Marine Geology* 127, 45

669  
670 Laberg, J.S., Vorren, T.O. 2000. Flow behaviour of the submarine glacial debris flows on the Bear Island  
671 trough mouth fan, western Barents Sea. *Sedimentology* 47, 1105-1117.

672 Lee, J.R., Philips, E. 2013. Glacitectonics – a key approach to examining ice dynamics, substrate rheology and  
673 ice-bed coupling. *Proceedings of the Geologists' Association* 124, 731-737.

674 Long, D., Stevenson, A., Wilson, C., Bulat, J. 2003. Slope failures in the Faroe-Shetland channel. In: Locat, J.,  
675 Miennert, J. (eds). *Submarine Mass Movements and Their Consequences*. Kluwer Academic, The Netherlands.

676 Lucchi, R., Rebesco, M., Camerlenghim A., Buseti, M., Tomadin, L., Villa, G., Persico, D., Morigi, C., Bonci,  
677 M., Giorgetti, G. 2002. Mid-late Pleistocene glacial-marine sedimentary processes of a high-latitude, deep sea  
678 sediment drift (Antarctic Peninsula Pacific margin). *Marine Geology* 189, 343-370.

679 Maier, K.L., Rosenberger, K., Paull, C.K., Gwiazda, R., Gales, J., Lorenson, T., Barry, J., Talling, P.J.,  
680 McGann, M., Xu, J.P., Lundsten, E., Anderson, K., Litvin, S.Y., Parsons, D., Clare, M.A., Simmons, S.M.,  
681 Sumner, E.J., Cartigny, M.J., and the Monterey Coordinated Canyon Experiment Team. 2019. Sediment and  
682 organic carbon transport and deposition driven by internal tides along Monterey Canyon, offshore  
683 California. *Deep-Sea Research I*, doi.org/10.1016/j.dsr.2019.103108.

684 Manabe, S., Stouffer, R.J. 1995. Simulation of abrupt climate change induced by freshwater input to the North  
685 Atlantic Ocean. *Nature* 378, 165-167.

686 Maslin, M., Owen, M., Day, S., Long, D. 2004. Linking continental-slope failures and climate change: Testing  
687 the clathrate gun hypothesis. *Geology* 32, 53-56.

688 McCave, I.N. 1984. Erosion, transport and deposition of fine-grained marine sediments. *Geological Society*  
689 *London Special Publication* 15, 35– 69.

690 McKay, R., Golledge, N., Maas, S., Naish, T., Levy, R., Dunbar, G., Kuhn, G. 2016. Antarctic marine ice-sheet  
691 retreat in the Ross Sea during the early Holocene. *Geology* 44, 7-10.

692 McKay, R., De Santis, L., Kulhanek, D., and the Expedition 374 Scientists. 2019. *Proceedings of the*  
693 *International Ocean Discovery Program Volume 374*. Publications.iodp.org,  
694 doi.org/10.14379/iodp.374.107.2019.

695 Melles, M., Kuhn, G. 1993. Sub-bottom profiling and sedimentological studies in the southern Weddell Sea,  
696 Antarctica: evidence for large scale erosional / depositional processes. *Deep-Sea Research* 40, 739-760.

697 Morrison, A., Hogg, A., England, M., Spence, P. 2020. Warm circumpolar deep water transport toward  
698 Antarctica driven by local dense water exports in canyons. *Science Advances* 6, eaav2516.

699 Nettles, M., Ekström, G. 2010. Glacial earthquakes in Greenland and Antarctica. *Annual Review of Earth and*  
700 *Planetary Sciences* 38, 467-491.

701 Nilsen, T.H., Shew, R.D., Steffens, G.S., Studlick, J.R.J. 2008. Atlas of Deep-water Outcrops. AAPG Studies in  
702 *Geology* 56, 504 pp.

703 Noormets, R., Dowdeswell, J., Larter, R., Ó Cofaigh, C., Evans, J. 2009. Morphology of the upper continental  
704 slope in the Bellingshausen and Amundsen seas – implications for sedimentary processes at the shelf edge of  
705 West Antarctica. *Marine Geology* 258, 100-114.

706 Normandeau, A., Dietrich, P., Hughes Clarke, J.E., Van Wychen, W., Lajeunesse, P., Burgess, D., Ghienne, J.F.  
707 2019. Retreat pattern of glaciers controls the occurrence of turbidity currents on high-latitude fjord deltas  
708 (eastern Baffin Island). *Journal of Geophysical Research - Earth Surface* 124, doi.org/10.1029/2018JF004970.

709 Ó Cofaigh, C., Taylor, J., Dowdeswell, J., Pudsey, C. 2003. Palaeo-ice streams, trough mouth fans and high-  
710 latitude continental slope sedimentation. *Boreas* 32, 37-55.

711 Orsi, A.H., Wiederwohl, C.L. 2009. A recount of Ross Sea waters. *Deep Sea Research, Part II. Topical Studies*  
712 *in Oceanography* 56, 778–795.

713 Padman, L., Siegfried, M., Fricker, H. 2018. Ocean tide influences on the Antarctic and Greenland Ice Sheets.  
714 *Review of Geophysics* 56, 142-184.

715 Palanques, A., Puig, P., Latasa, M., Scharek, R. 2009. Deep sediment transport induced by storms and dense  
716 shelf-water cascading in the northwestern Mediterranean basin. *Deep Sea Research Part I: Oceanographic*  
717 *Research Papers* 56, 425-434.

718 Pattyn, F., Van Huele, W. 1998. Power Law or Power Flow. *Earth Surface Processes and Landforms* 23, 761-  
719 767.

720 Petrelli, P., Bindoff, N., Bergamasco, A. 2008. The sea ice dynamics of Terra Nova Bay and Ross Ice Shelf  
721 polynyas during a spring and winter simulation. *Journal of Geophysical Research* 113, C09003,  
722 doi:10.1029/2006JC004048.

723 Petrini, M., Colleoni, F., Kirchner, N., Hughes, A.L.C., Camerlenghi, A., Rebesco, M., Lucchi, R.G., Forte, E.,  
724 Colucci, R.R., Noormets, R. 2018. Interplay of grounding-line dynamics and sub-shelf melting during retreat of  
725 the Bjornoyrenna Ice Stream. *Scientific Reports* 8, DOI:10.1038/s41598-018-25664-6.

726 Pike, L., Gaskin, S., Ashmore, P. 2018. Flume tests of fluvial erosion mechanisms in till-bed channels. *Earth  
727 Surface Processes and Landforms* 43, 259-270.

728 Piper, D., Cochonat, P., Morrison, M. 1999. Sidescan sonar evidence for progressive evolution of submarine  
729 failure into a turbidity current: the 1929 Grand Banks event. *Sedimentology* 46, 79-97.

730 Pope, E., Normandeau, A., Ó Cofaigh, C., Stokes, C., Talling, P. 2019. Controls on the formation of turbidity  
731 current channels associated with marine-terminating glaciers and ice sheets. *Marine Geology* 415, 105951.

732 Post, A., O'Brien, P., Edwards, S., Carroll, A., Malakoff, K., Armand, L. 2020. Upper slope processes and  
733 seafloor ecosystems on the Sabrina continental slope, East Antarctica. *Marine Geology* 422, 106091.

734 Prothro, L., Majewski, W., Yokoyama, Y., Simkins, L., Anderson, J., Yamane, M., Miyairi, Y., Ohnkouchi, N.  
735 2020. Timing and pathways of East Antarctic Ice Sheet retreat. *Quaternary Science Reviews* 230, 106166.

736 Prothro, L., Simkins, L., Majewski, W., Anderson, J. 2018. Glacial retreat patterns and processes determined  
737 from integrated sedimentology and geomorphology records. *Marine Geology* 395, 104-119.

738 Purkey, S., Johnson, G. 2010. Warming of global abyssal and deep Southern Ocean waters between the 1990s  
739 and 2000s: contributions to global heat and sea level rise budgets. *Journal of Climate* 23, 6336-6351.

740 Rahmstorf, S. 1994. Rapid climate transitions in a coupled ocean-atmosphere model. *Nature* 372, 82-85.

741 Rebesco, M., Camerlenghi, A. 2008. Late Pliocene margin development and mega debris flow deposits on the  
742 Antarctic continental margins: Evidence of the onset of the modern Antarctic Ice Sheet? *Palaeogeography,  
743 Palaeoclimatology, Palaeoecology* 260, 149-167.

744 Rebesco, M., Camerlenghi, A., Volpi, V., Neagu, C., Accettella, D., Lindberg, B., Cova, A., Zgur, F., Magico  
745 Party. 2007. Interaction of processes and importance of contourites: insights from the detailed morphology of  
746 sediment Drift 7, Antarctica. *Geological Society, London, Special Publications* 276, 95-110.

747 Rebesco, M., Hernandez-Molina, J., Van Rooij, D., Wahlin, A. 2014. Contourites and associated sediments  
748 controlled by deep-water circulation processes: State of the art and future considerations. *Marine Geology* 352,  
749 111-154.

750 Rebesco, M., Camerlenghi, A., Zanolla, C. 1998. Bathymetry and morphogenesis of the continental margin west  
751 of the Antarctic Peninsula. *Terra Antarctica* 4, 715-725.

752 Rebesco, M., Larter, R., Camerlenghi, A., Barker, P. 1996. Giant sediment drifts on the continental rise west of  
753 the Antarctic Peninsula. *Geo-Marine Letters* 16, 65-75.

754 Rignot, E., Mouginot, J., Scheuchl, B., Van den Broeke, M., van Wessem, M.J., Morlighem, M. 2019. Four  
755 decades of Antarctic Ice Sheet mass balance from 1979-2017. *Proceedings of the National Academy of*  
756 *Sciences*, doi.org/10.1073/pnas.1812883116.

757 Robertson, R., Beckmann, A., Hellmer, H. 2003. M2 tidal dynamics in the Ross Sea. *Antarctic Science* 15, 41–  
758 46.

759 Rothlisberger, H. 1972. Water pressure in intra- and subglacial channels. *Journal of Glaciology* 11, 177-203.

760 Rui, L., Rebesco, M., Casamor, J.L., Laberg, J.S., Rydningen, T.A., Caburlotto, A., Forwick, M., Urgeles, R.,  
761 Accettella, D., Lucchi, R.G., Barsanti, M., Demarte, M., Ivaldi, R., Delbono, I. 2019. Geomorphology and  
762 development of a high-latitude channel system: the INBIS channel case (NW Barents Sea, Arctic). *Arktos* 5, 15-  
763 29.

764 Shumaker, L.E., Jobe, Z.R., Graham, S.A. 2017. Evolution of submarine gullies on a prograding slope: Insights  
765 from 3D seismic reflection data. *Marine Geology* 393, 35-46.

766 Seidov, D., Barron, E.J., Haupt, B.J. 2001. Meltwater and the global ocean conveyor: northern versus southern  
767 connections. *Global and Planetary Change* 30, 253–266.

768 Seidov, D., Stouffer, R., Haupt, B. 2005. Is there a simple bi-polar ocean seesaw? *Global and Planetary change*  
769 49, 19-27.

770 Shipp, S., Anderson, J.B., Domack, E.W. 1999. Late Pleistocene / Holocene retreat of the west Antarctic ice-  
771 sheet system in the Ross Sea. Part 1. Geophysical results. *Geological Society of America Bulletin* 111, 1486-  
772 1516.

773 Silvano, A., Rintoul, S.R., Pena-Molina, B., Hobb, W.R., Van Wijk, E., Aoki, S., Tamura, T., Williams, G.D.  
774 2018. Freshening by glacial meltwater enhances melting of ice shelves and reduces formation of Antarctic  
775 bottom water. *Science Advances* 4, doi.org/10.1126/sciadv.aap9467.

776 Simkins, L., Anderson, J., Greenwood, S., Gonnermann, H., Prothro, L., Halberstadt, A., Stearns, L., Pollard,  
777 D., De Conto, R. 2017. Anatomy of a meltwater drainage system beneath the ancestral East Antarctic ice sheet.  
778 *Nature Geoscience* 10, 691-697.

779 Simons, D.B., Sentürk, F. 1992. *Sediment Transport Technology: Water and Sediment Dynamics*. Water  
780 Resources Publications, Highlands Ranch, Colorado, 901 pp.

781 Smith Jr, W., Nelson, D., DiTullio, G., Leventer, A. 1996. Temporal and spatial patterns in the Ross Sea:  
782 Phytoplankton biomass, elemental composition, productivity and growth rates. *Journal of Geophysical*  
783 *Research: Oceans* 101, 18455-18465.

784 Sugden, D.E., Denton, D.H., Marchant, D.R. 1991. Subglacial meltwater channel system and ice sheet over  
785 riding of the Asgard range, Antarctica. *Geografiska Annaler* 73A, 109-121.

786 Talling P. 2014. On the triggers, resulting flow types and frequency of subaqueous sediment density flows in  
787 different settings. *Marine Geology* 352, 155–182.

788 Traube, V., Zayatz, I. 1993. Main stages of development of the Eastern Basin, Ross Sea, imprinted in its  
789 structure. In *LIRA workshop on Landscape Evolution – a multidisciplinary approach to the relation of Cenozoic*  
790 *climate change and tectonics in the Ross Sea area, Antarctica*. Wateren, F., van der Verbers, A., Tessensohl, F.  
791 (eds). Rijks Geol Dienst, Haarlem, Netherlands, pp. 91-94.

792 Truffer, M., Motyka, R.J. 2016. Where glaciers meet water: subaqueous melt and its relevance to glaciers in  
793 various settings. *Review of Geophysics* 54, doi.org/10.1002/2015RG000494.

794 Vanneste, L., Larter, R. 1995. Deep-tow boomer survey of the Antarctic Peninsula Pacific margin: an  
795 investigation of the morphology and acoustic characteristics of late Quaternary sedimentary deposits on the  
796 outer continental shelf and upper slope. In: Cooper, A., Barker, B., Brancolini, B. (eds). *Geology and seismic*  
797 *stratigraphy of the Antarctic margin*. Antarctic Research Series 68, American Geophysical Union, Washington,  
798 D.C., 97-121.

799 Volpi, V., Camerlenghi, A., Hillenbrand, C.-D., Rebesco, M., Ivaldi, R. 2003. Effects of biogenic silica on  
800 sediment compaction and slope stability on the Pacific margin of the Antarctic Peninsula. *Basin Research* 15,  
801 339–363.

802 Vorren, T.O., Laberg, J.S., Blaume, F., Dowdeswell, J.A., Kenyon, N.H., Mienert, J., Rumohr, J., Werner, F.  
803 1998. The Norwegian-Greenland Sea continental margins: Morphology and late Quaternary sedimentary  
804 processes and environment. *Quaternary Science Reviews* 17, 273-302.

805 Weaver, A., Saenko, O., Clark, P., Mitrovica, J. 2003. Meltwater Pulse 1A from Antarctica as a trigger of the  
806 Bolling-Allerod Warm Interval. *Science* 299, 1709-1713.

807 Wynn, R., Talling, P., Masson, D., Le Bas, T., Cronin, B., Stevenson, C. 2012. The influence of subtle gradient  
808 changes on deep-water gravity flows: a case study from the Moroccan turbidite system. In: Prather, B.E;  
809 Deptuck, M.E; Mohrig, D; Van Hoorn, B; Wynn, R.B, (eds.). *Application of the Principles of Seismic*  
810 *Geomorphology to Continental-Slope and Base-of-Slope Systems: Case Studies from Seafloor and Near*  
811 *Seafloor Analogues: SEPM, Special Publication 99*, 371-383.

812 Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K. 2001. Trends, rhythms and aberrations in global  
813 climate 65 Ma to present. *Science* 292, 686-693.

814

## 815 **10. Figure captions**

816

817 Figure 1. A. Study area in the Ross Sea. White box shows location of (B). Bathymetric data is International  
818 Bathymetric Chart of the Southern Ocean (IBSCO, Arndt et al., 2013). Yellow triangle is position of cores  
819 RS14-BC3 and RS14-C3. Contour spacing is 200 m. B. Hillary Canyon regional bathymetry taken from IBSCO.  
820 White lines indicate positions of along and down-slope profiles shown in (C). White boxes show locations of  
821 Fig. 2, 3 and 5. Yellow triangle is position of cores RS14-BC3 and RS14-C3. C. Along and down-slope profiles  
822 across the Hillary Canyon derived from bathymetry using IBSCO. Stars highlight position of along-slope  
823 profiles shown in (B).

824

825 Figure 2. Submarine slide morphology and prograded continental margin at the mouth of the Glomar Challenger  
826 Basin. A. Slide morphology. Hillshaded bathymetry is gridded at 50 m. Regional bathymetry is from IBSCO.

827 Contour spacing is 50 m. Black lines mark location of single channel seismic lines IT17RS305 and IT17RS306.

828 B. Seismic line IT17RS306 (near trace). C. Interpretation of seismic line IT17RS306 (near trace).

829 Figure 3. A. Morphology at the head of the Hillary Canyon, Ross Sea. Hillshaded bathymetry gridded at 50 m

830 cell size. Contour spacing is 50 m. Background bathymetric data is International Bathymetric Chart of the

831 Southern Ocean (IBSCO). Black lines mark location of single channel seismic lines IT17RS305 in (B) and

832 IT17RS306. Black dashed line x-x' locates along-slope profile in (A). White dashed lines b-b' and c-c' locate

833 sub-bottom profiles in (D) and (E). B. Single channel seismic line IT17RS305. Black box locates (C). C. Zoom-

834 in of seismic line IT17RS305 located in (B). D. Unmigrated sub-bottom profile located in (A). E. Unmigrated

835 sub-bottom profile located in (A).

836 Figure 4. Gully morphometric parameters along profile X-X' located in Figure 3A. Width of bars in top two

837 panels correspond to gully width. Black diamonds correspond to average slope gradient ( $^{\circ}$ ) of the upper slope

838 with distance parallel to the shelf edge.

839 Figure 5. A. Morphology of the Hillary Canyon thalweg. Hillshaded bathymetry is gridded at 50 m. Contour

840 spacing is 50 m. Location of (A) shown in Figure 1B. B. Single channel seismic line IT17RS303 showing cross-

841 section through the Hillary Canyon thalweg and canyon levees.

842 Figure 6. Oceanographic measurements along profile a-a'. A. Temperature data from 20 eXpendable

843 BathyThermograph (XBT) profiles along profile a-a' located in (C). B. Current velocity data from Acoustic

844 Doppler Current Profiler along profile a-a' located in (C). C. Hillshaded bathymetric data gridded at 50 m at the

845 head of the Hillary Canyon, Ross Sea. Black line locates profile a-a' shown in (A) and (B). Blue lines are

846 current vectors indicating current direction at 400 m water depth along dashed line in (B). Red stars locate eight

847 Conductivity Temperature Depth profiles shown in Fig. 7A.

848 Figure 7. Oceanographic measurements at the head of the Hillary Canyon, Ross Sea. A. Conductivity

849 Temperature Depth (CTD) profiles with fluorescence and turbidity sensor data. Location of CTD profiles are

850 shown in Fig. 6C. B. Lowered-Acoustic Doppler Current Profiler data. Positions are shown in Fig. 6C. Red

851 dashed line marks 0.7 m/s for comparison in all profiles.

852 Figure 8. Inferred ice position at Last Glacial Maximum. Background bathymetric data is International

853 Bathymetric Chart of the Southern Ocean (IBSCO). Contour spacing is 50 m. Hillshaded bathymetry from

854 ANTSSS expedition gridded at 50 m cell size. White dashed line is inferred ice extent at LGM taken from

855 Halberstadt et al. (2016). GZW is Grounding Zone Wedge position taken from Halberstadt et al. (2016). MR is  
 856 morainal ridge. Red lines are gully positions.

857 Figure 9. Schematic of canyon head processes operating in glacial vs interglacial settings. A. Interglacial /  
 858 modern conditions where gravity flows are generated fine grained biogenic / hemipelagic material that is  
 859 resuspended by energetic cascading cold, dense water. B. Glacial conditions where energetic turbidity currents  
 860 may be generated by either: the release of subglacial meltwater; meltwater generated by intruding warm currents  
 861 at the ice front; slope failure. HSSW is High Salinity Shelf Water.

862 Figure 10. Antarctic gully morphometric parameters. A. Gully length vs gully relief. B. Gully width vs gully  
 863 length. C. Gully sinuosity vs cross sectional shape. Data from the Weddell Sea, Bellingshausen Sea, Amundsen  
 864 Sea, Western Antarctic Peninsula (WAP) taken from Gales et al. (2013). Ross Sea data from this study. For  
 865 calculations of gully parameters, see SM.Fig.1.

866

## 867 11. Supplementary Material

868 SM Table 1. Variables used in calculations [1]

Value	Description	Data source
$u$	Depth averaged flow speed ( $\text{m/s}^{-1}$ )	This study; $1 \text{ m/s}^{-1}$
$C_d$	Friction factor	<i>Dimensionless (typically 0.0025)</i>
$P$	Water density	<i>This study; 1028.27 (cold water density)</i>

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870 SM Figure 1. Measured gully morphometric parameters. A. Cross sectional view of gully where  $G^w$  is gully  
 871 width (distance between points of maximum curvature of gully flanks);  $G^l$  is gully relief (vertical distance from  
 872 maximum gully incision to line defining gully width); and  $G^{st}$  is gully steepness (ratio between gully relief and  
 873 width). B. Plan view of gully thalweg where  $G^L$  is gully length,  $G^{SL}$  is straight line distance of gully length and  
 874  $G^s$  is gully sinuosity (ratio of gully length vs straight-line distance of gully length; where 1 = straight and  $>1$   
 875 indicates increasing sinuosity).

876 SM Figure 2. Cumulative grain size data for cores RS14-BC3 and RS13-C3 (locations marked in Fig 1). A.  
 877 Cumulative grain size plot for box core RS14-BC3 (0-5 cm core depth). B. Cumulative grain size plot for core  
 878 RS14-C3 (0-25 cm core depth). Depth of samples chosen to represent interglacial sediments. Outlier sample 24-

879 25 highlighted in key. C. Grain size volume measurements of sand-silt-clay, sand (63  $\mu\text{m}$  - 2 mm), silt (2  $\mu\text{m}$  -  
880 63  $\mu\text{m}$ ) and clay (< 2  $\mu\text{m}$ ) for gravity core RS14-C3. D. Grain size volume measurements of sand-silt-clay,  
881 where sand (63  $\mu\text{m}$  - 2 mm), silt (2  $\mu\text{m}$  - 63  $\mu\text{m}$ ) and clay (< 2  $\mu\text{m}$ ) for box core RS14-BC3. Gravel counts (>  
882 2mm) shown in right column.

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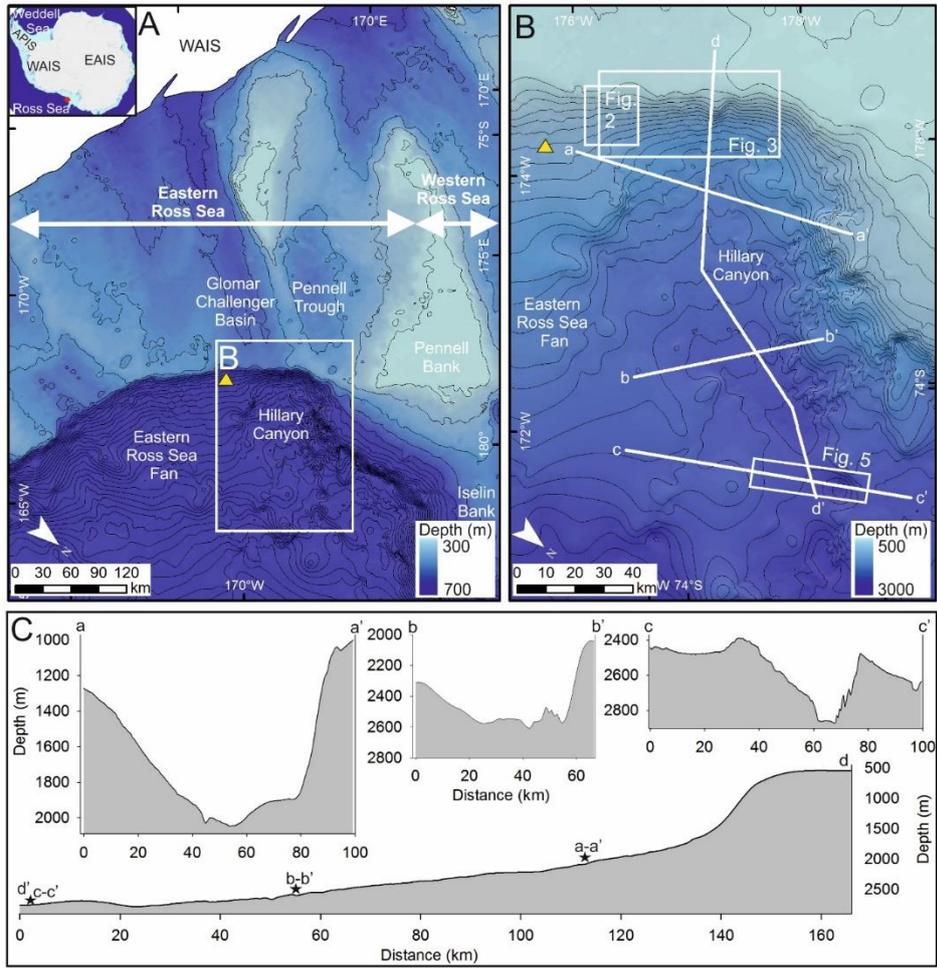
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901 **12. Figures**

902 Figure 1.



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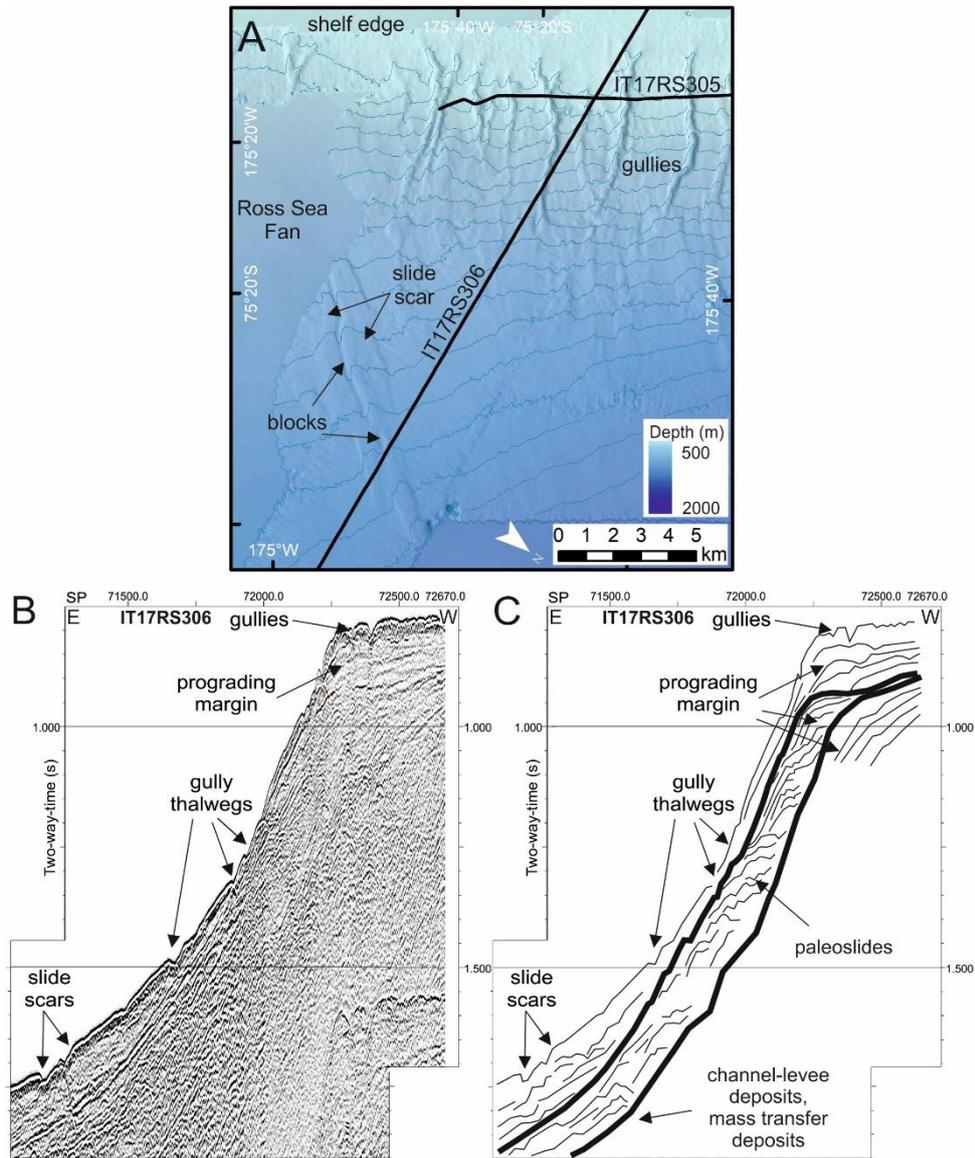
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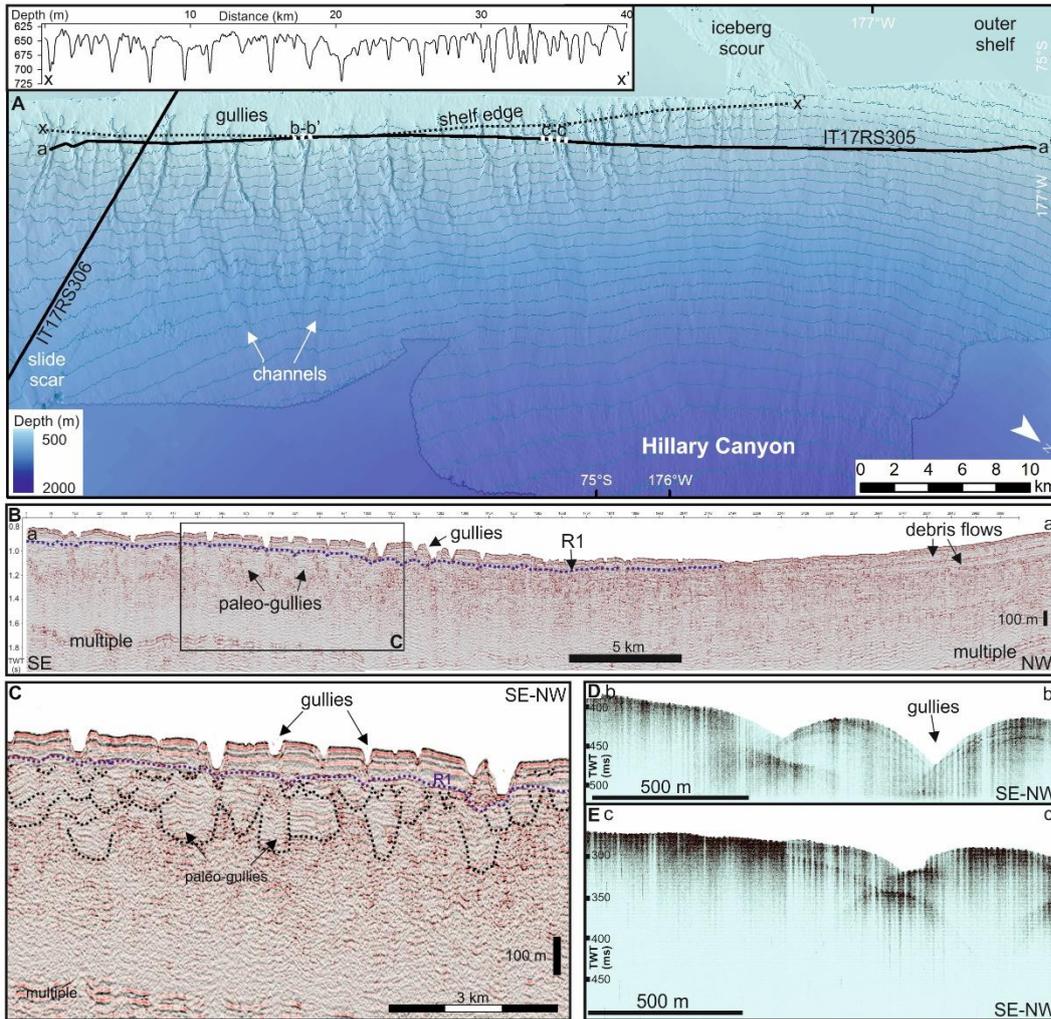
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920 Figure 3



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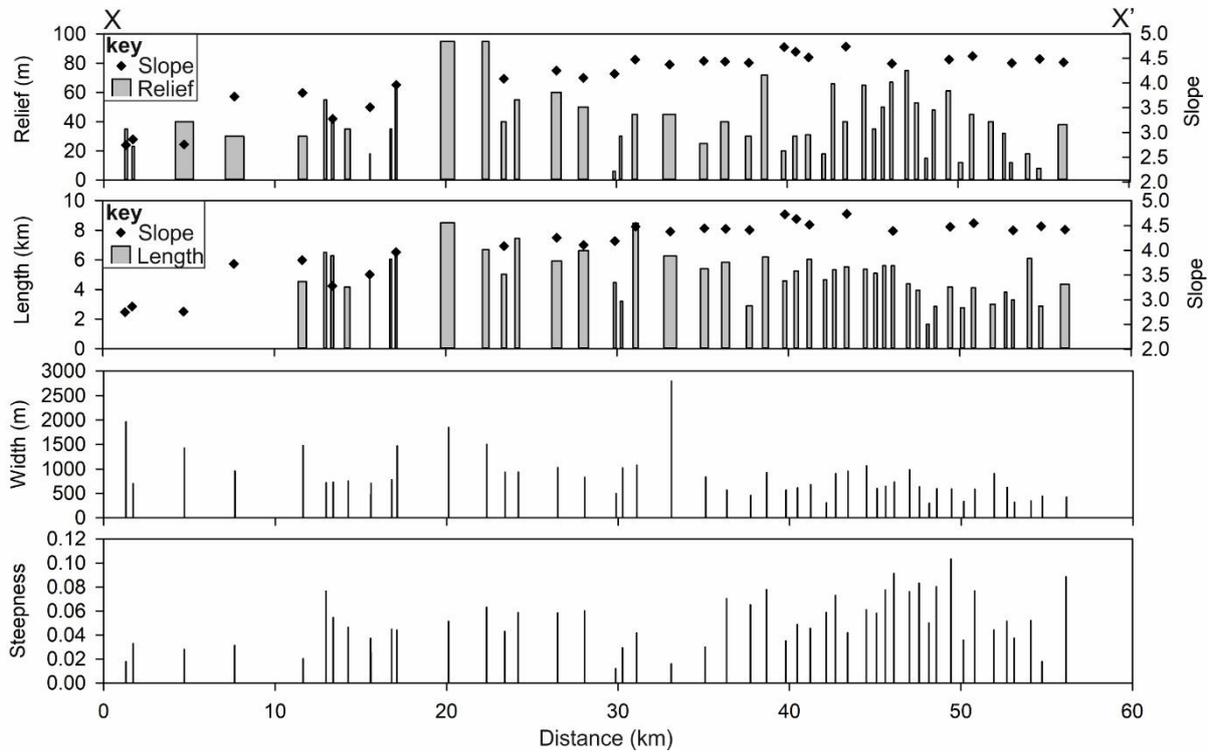
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930 Figure 4



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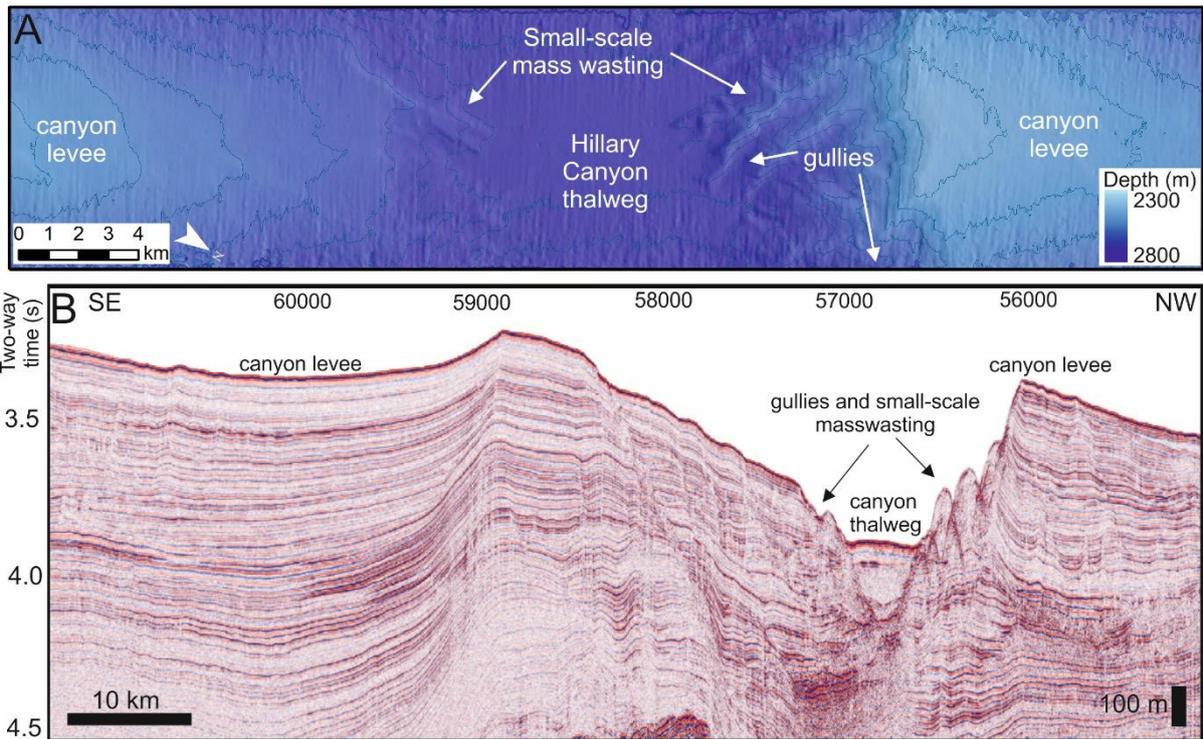
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943 Figure 5



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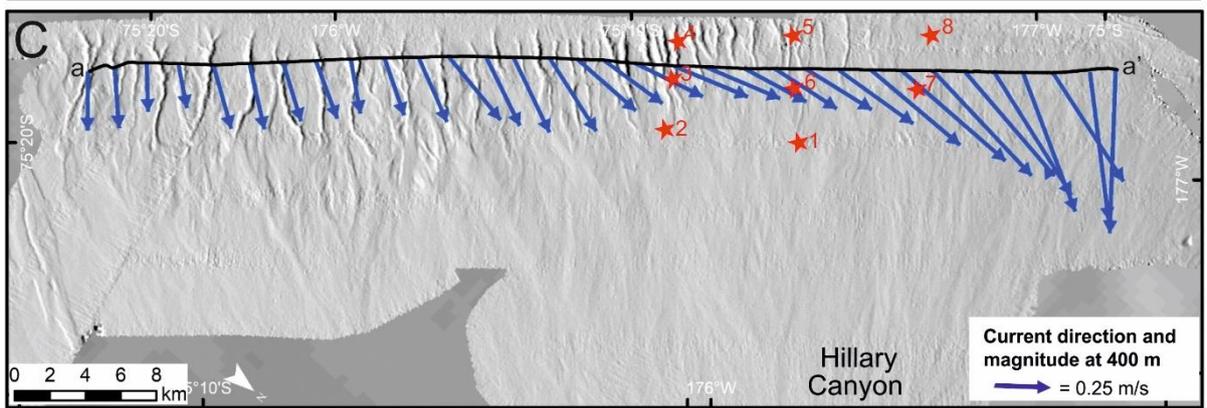
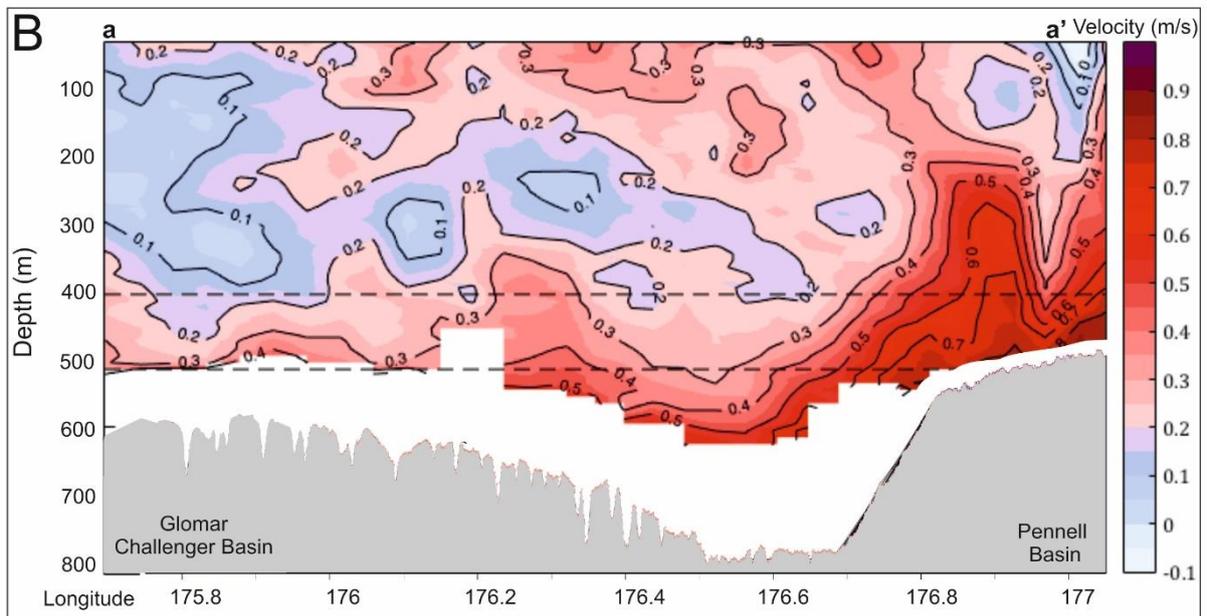
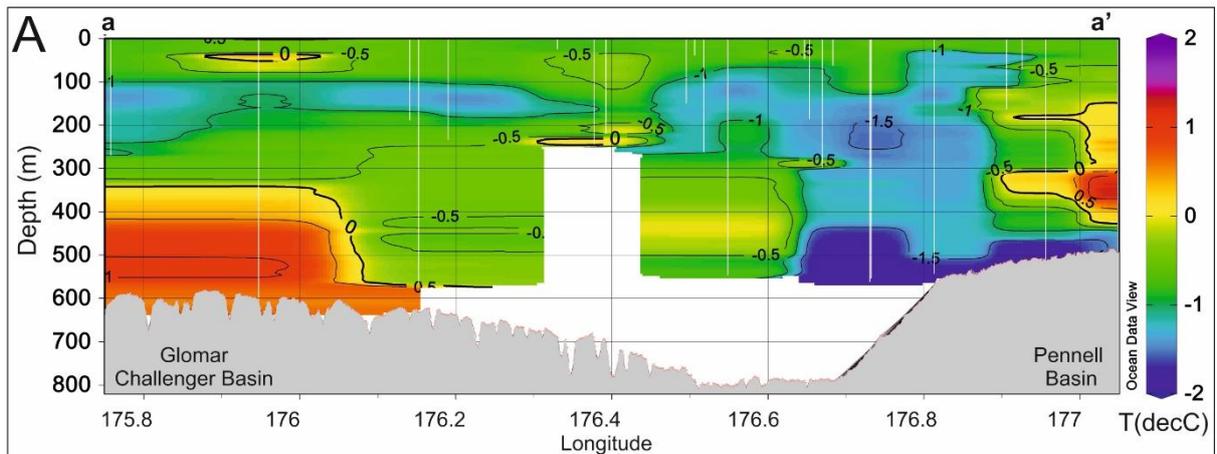
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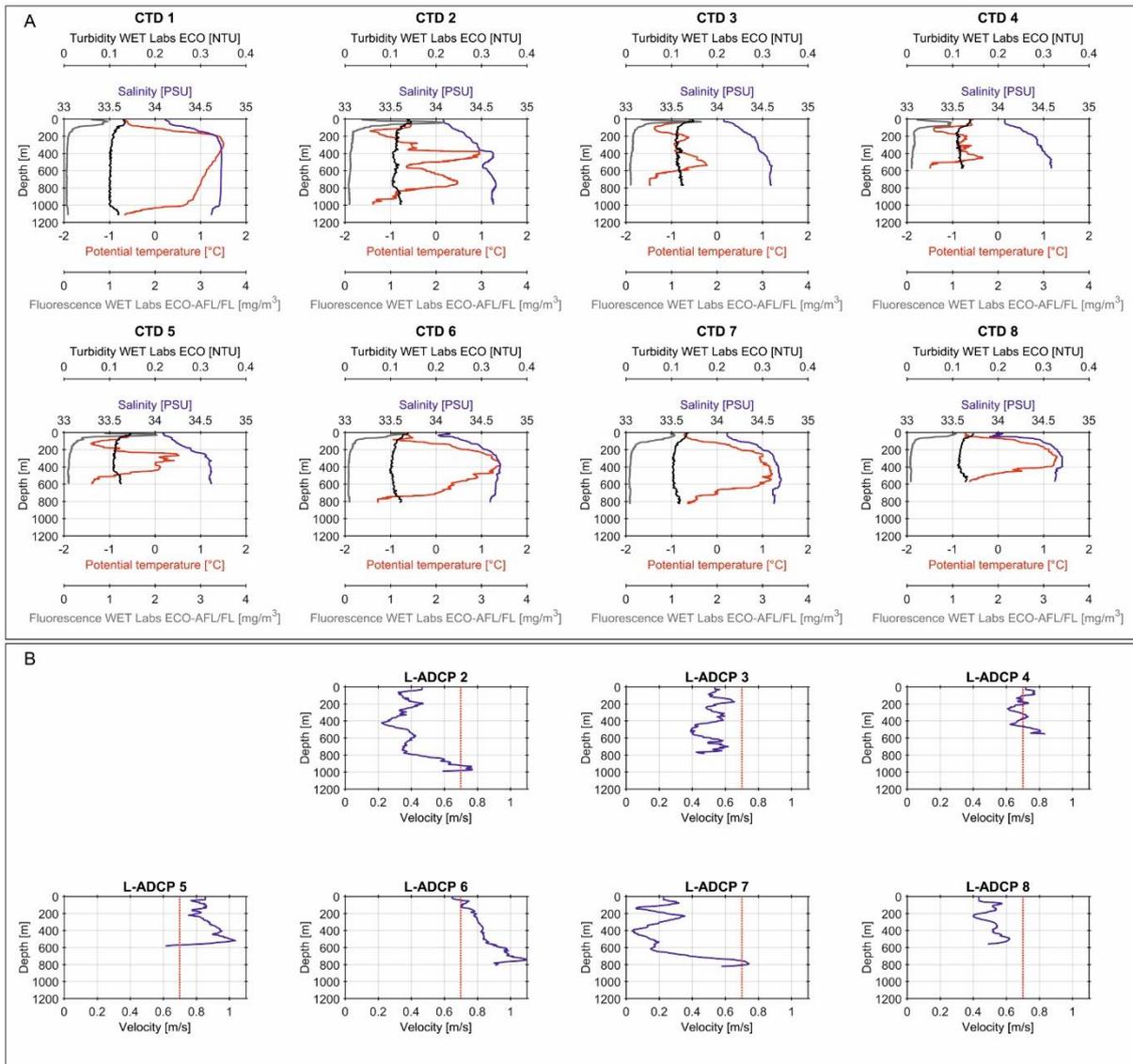


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960 Figure 7



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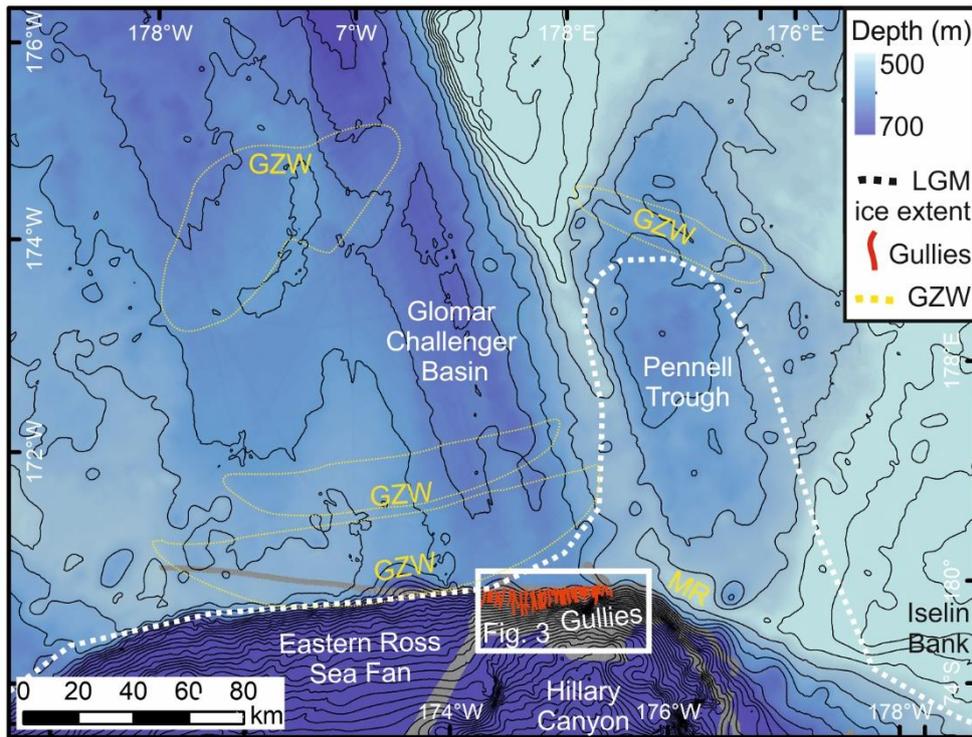
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969 Figure 8.



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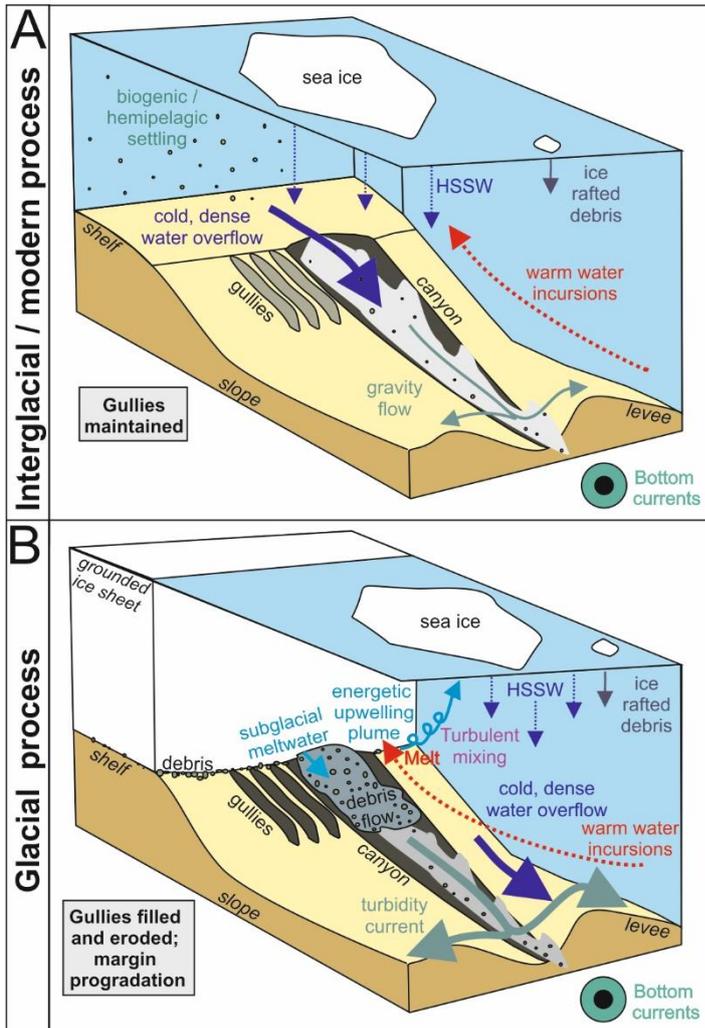
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982 Figure 9



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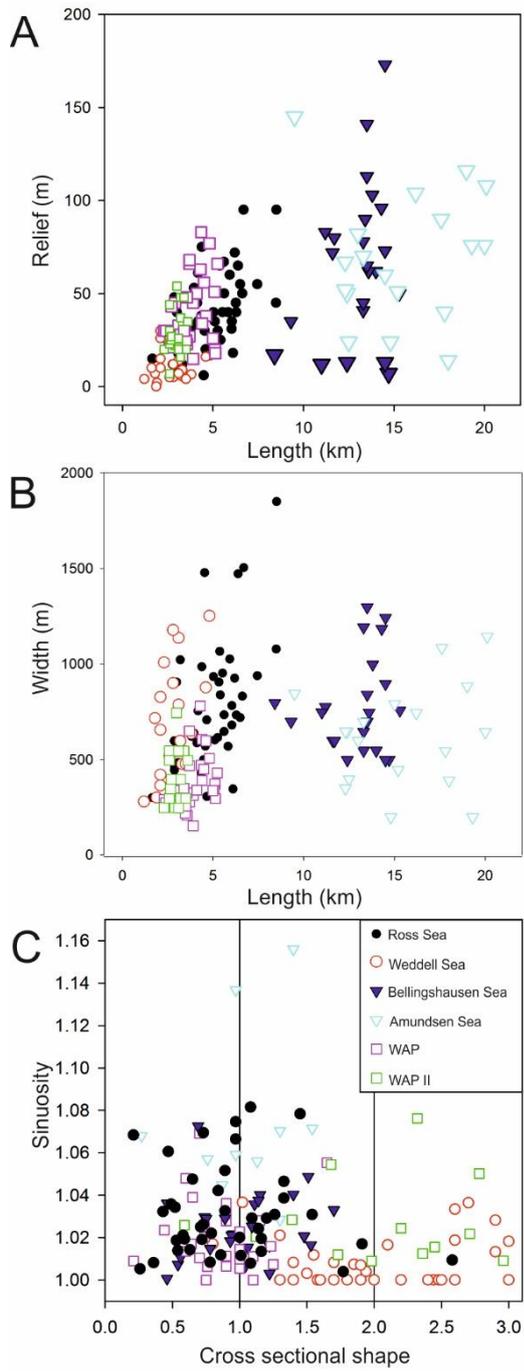
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992 Figure 10



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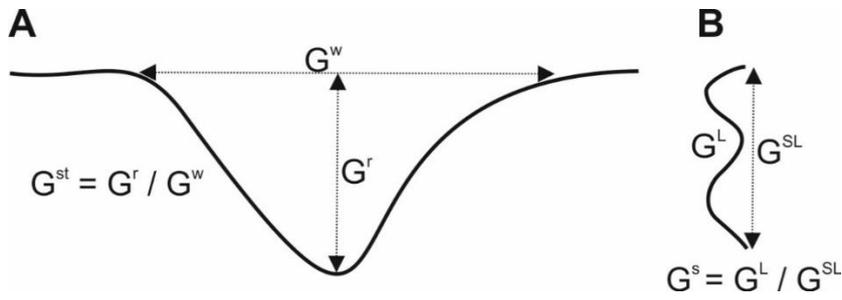
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999 SM. Figure 1



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