

1 Thwaites Glacier thins and retreats fastest where ice-shelf channels 2 intersect its grounding zone

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9 **Abstract.** Antarctic ice shelves buttress the flow of the ice sheet but are vulnerable to increased basal melting from contact
10 with a warming ocean and increased mass loss from calving due to changing flow patterns. Channels and similar features at
11 the bases of ice shelves have been linked to enhanced basal melting and observed to intersect the grounding zone, where the
12 greatest melt rates are often observed. The ice shelf of Thwaites Glacier is especially vulnerable to basal melt and grounding-
13 zone retreat because the glacier has a retrograde bed leading to a deep trough below the grounded ice sheet. We use digital
14 surface models from 2010–2022 to investigate the evolution of its ice-shelf channels, grounding zone position, and the
15 interactions between them. We find that the highest sustained rates of grounding-zone retreat (up to 0.7 km yr⁻¹) are associated
16 with high basal melt rates (up to ~250 m yr⁻¹) and are found where ice-shelf channels intersect the grounding zone, especially
17 atop steep local retrograde slopes where subglacial channel discharge is expected. We find no areas with sustained grounding
18 zone advance, although some secular retreat was distal from ice-shelf channels. Pinpointing other locations with similar risk
19 factors could focus assessments of vulnerability to grounding zone retreat.

20 **1 Introduction**

21 Thwaites Glacier in the Amundsen Sea region of West Antarctica has the potential to contribute up to 65 cm of sea-level rise
22 (Rignot et al., 2019; Morlighem et al., 2020) and has experienced recent speed-up, ice shelf break-up, thinning, and grounding-
23 line retreat (e.g. dos Santos et al., 2021). Thwaites Glacier lies atop an inland-sloping (retrograde) bed leading to a deep trough
24 reaching 1.5 km below sea level (Morlighem et al., 2020). The glacier terminates in two distinct ice shelves, the Thwaites
25 Eastern Ice Shelf (TEIS) and the Thwaites Western Ice Tongue (TWIT), collectively referred to as the Thwaites Glacier Ice
26 Shelf (TGIS). Several studies have suggested that the Thwaites Glacier grounding zone (the region where the ice transitions
27 from grounded to freely floating), may already be undergoing rapid, unstable retreat (Joughin et al., 2014; Goldberg et al.,
28 2015; Rignot et al., 2014; Yu et al., 2018), in a process known as “marine ice sheet instability”, or MISI (Gudmundsson et al.,

29 2012; Schoof, 2007; Weertman, 1974). Much of the basal melting and grounding zone retreat is attributed to contact with
30 warm ocean water (Bevan et al., 2021; Schmidt et al., 2023; Holland et al., 2023) and loss of basal traction inland of the
31 grounding zone (Joughin et al., 2024). Recent evidence suggests that tidal flexure allows seawater to intrude into the embayed
32 grounding zone in the main trunk of the TWIT (located within Box C in Fig. 1), which may accelerate melting below
33 intermittently grounded ice (Rignot et al., 2024). The TWIT has weakened rapidly over the past several decades (e.g. Miles et
34 al., 2020), and the TEIS is expected to weaken significantly in the coming decades (e.g. Wild et al., 2022).

35 Several studies have shown that high rates of ice-shelf thinning and basal melting are often associated with ice-shelf
36 channels, which are curvilinear incisions at the ice-shelf base believed to be maintained by buoyant meltwater plumes entrained
37 within them along-flow (Alley et al., 2016; Drews, 2015; Chartrand and Howat, 2020; Gourmelen et al., 2017; Wearing et al.,
38 2021). Others have shown that ice-shelf channels are also commonly associated with ice-shelf weakening through fracturing
39 as a result of thinning (Vaughan et al., 2012; Dow et al., 2018; Alley et al., 2019). Ice-shelf channels initiate at the grounding
40 zone and extend seaward. They often represent advected extensions of inverted troughs initiated by subglacial channelization
41 beneath the grounded ice or incised by undulations in the bed (e.g. Le Brocq et al., 2013; Alley et al., 2016; Drews et al., 2017).
42 Where subglacial channelization is present, the input of fresh subglacial meltwater may contribute to the growth of a buoyant
43 meltwater plume that can entrain warm ocean water as it travels along the ice-shelf channel (Jenkins, 2011). However, it
44 remains difficult to attribute the formation mechanism to any given channel, particularly if its surface expression does not
45 intersect the grounding zone (e.g. Alley et al., 2016; Chartrand & Howat 2020). The TGIS has at least four previously mapped
46 ice-shelf channels (Alley et al., 2016) and the grounded ice is underlain by at least two persistent subglacial channels
47 intersecting the grounding zone (Hager et al., 2022). Recently, subglacial channels and aligned ice-shelf channels on Greenland
48 ice tongues have been linked to thinning and retreat of the grounding line (Ciraci et al., 2023; Narkevic et al., 2023). It is
49 unknown, however, the extent to which ice-shelf channels on the TGIS and/or subglacial channels within grounded ice may
50 have contributed to the thinning and retreat at Thwaites Glacier.

51 In the absence of a method for directly measuring ice thickness from space, observations of ice-shelf channels and
52 channel-like features (i.e. ice-shelf incisions oriented predominantly along-flow without clear evidence of subglacial initiation
53 or entrained meltwater flow) and their relationship to variations in grounding zone position and other ice-shelf structures are
54 only available from high-resolution (<100 m) measurements of surface height. While relatively frequent and accurate,
55 observations from spaceborne altimetry, such as ICESat and ICESat-2, are limited to ground tracks. Recently, high-resolution
56 digital surface models (DSMs) produced from stereoscopic satellite imagery, combined with altimetry, have been used to map
57 changes in ice-shelf channels and other ice-shelf structures (Chartrand and Howat, 2020; Shean et al., 2019; Zinck et al., 2023).
58 Using the extensive collection of repeat DSMs provided by the Reference Elevation Model of Antarctica (REMA) project
59 (Howat et al., 2019), we map the positions of surface depressions overlying ice-shelf channels on the TGIS and subglacial
60 channels within grounded ice as well as the landward extent of the transition to flotation as a proxy for the grounding line,
61 termed the hydrostatic boundary (HB). We also construct the most comprehensive maps to date of time-evolving surface
62 height, thickness, and basal mass change for the Thwaites Glacier and TGIS. Here we examine the transient locations of the

63 ice-shelf channels relative to those of the HB, as well as to variations in basal melt and flow speed, to assess the potential
64 relationship between channels and grounding-zone retreat.

65 **2 Datasets**

66 We use several geophysical datasets for this study as described throughout this section.

67 **2.1 REMA Digital Surface Model (DSM) strips and annual mosaics**

68 Reference Elevation Model of Antarctica (REMA) DSMs for the TGIS are obtained through stereophotogrammetry applied to
69 pairs of commercial submeter-resolution, panchromatic satellite images acquired by the MAXAR constellation, including
70 Worldview-1, -2 and -3, Quickbird-1 and -2, and GeoEye satellites (Howat et al., 2019). Elevations from REMA DSMs are
71 relative to the WGS84 ellipsoid and are distributed both as individual “strips” representing WGS84 ellipsoid elevations, created
72 from a single pair of images along their overlapping swath, and as seamless, continuous mosaics made from those strips. Both
73 product types are distributed at 2 m resolution, with downsampled versions available. We use the REMA 200 m mosaic for
74 the Thwaites region as a base map in several figures.

75 We utilise the DSM strips at 10 m resolution to map the HB (Section 3.1) and ice-shelf channels (Section 3.2). After
76 removing strips with insufficient coverage or low internal quality, as indicated in the metadata by a root-mean-squared error
77 value greater than 1 m, there are 191 strips for the TGIS.

78 We combine REMA strips into full-coverage (depending on strip availability) annual mosaics at 50 m resolution to
79 compute rates of change across the TGIS (Section 3.3). As strips are more readily available for summer months, REMA strips
80 from November to March are mosaicked to form an austral annual mosaic, for which we assign a nominal date of January 1.
81 For each annual mosaic, each strip is co-registered to every other using the method of Nuth and Kaab (2011). The co-registered
82 strips are then stacked, and the annual mosaic is the median height at each pixel through the stack (Fig. S1).

83 As part of this process, each strip and annual mosaic is registered to correct for single-value elevation biases using
84 ICESat-2, CryoSat-2, or Operation IceBridge (henceforth, IceBridge) altimetry data, depending on availability and/or which
85 dataset is temporally closest to the collection date of the strip. The registration dataset selected is based on a hierarchy; if the
86 first dataset is unavailable, the next dataset will be used, and so on. Priority is given to overlapping ICESat-2 ATL06 Version
87 5 (Smith et al., 2021) or Version 6 (Smith et al., 2023) ground control points (GCPs) that were collected within 10 days of the
88 strip or 100 days of the annual mosaic nominal date, then to GCPs from the IceBridge Airborne Topographic Mapper (ATM)
89 L1B (Studinger, 2013), the IceBridge Land, Vegetation, and Ice Sensor (LVIS) L2 (Blair and Hofton, 2015), or the IceBridge
90 and ICECAP Riegl Laser Altimeter L2 (Blankenship et al., 2012) datasets collected within 10 or 100 days. Strips with no
91 contemporaneous ICESat-2 or IceBridge GCPs were registered to CryoSat-2 SARIn-mode elevations (European Space
92 Agency, 2023) collected within about 365 days of the strip. Unlike ICESat-2 and IceBridge registrations which register the
93 strip to temporally proximal GCPs, the CryoSat-2 registration is determined by a linear fit to elevation differences with respect

94 to time so that the DSM is fit to a temporal model of the elevation data. Strips that do not meet these registration criteria are
95 eliminated, leaving 177 strips over the TGIS (Fig. S2).

96 The strips and annual mosaics are not smoothed, but they are masked to remove artefacts and errors like clouds.
97 Following registration, the absolute residual, and mean and standard deviation thereof, between each DSM and the REMA
98 mosaic is computed. Then, the residual is smoothed by a 50 pixel (500 m for strips or 2500 m for annual mosaics) moving
99 mean, and any DSM pixels that correspond to a smoothed residual that is 2–5 standard deviations greater than the mean residual
100 are masked out; a larger magnitude in the residual mean or standard deviation triggers a smaller standard deviation threshold
101 for that strip. These thresholds were chosen to effectively remove artefacts and errors like clouds, but to maintain differences
102 due to the advection of surface features.

103 Following the procedure in Chartrand and Howat (2020), registered and masked strip and annual mosaic ellipsoid
104 elevations are converted to freeboard heights (h) by referencing to mean sea level using the EIGEN–6C4 geoid model (Förste
105 et al., 2014), correcting for Mean Dynamic Topography using the DTU22 MDT model (Knudsen et al., 2021), accounting for
106 firm density using the firm depth correction from Ligtenberg et al. (2011) provided in BedMachine, and correcting for tidal
107 variations using the CATS2008b inverse tide model (Padman et al., 2018).

108 **2.2 BedMachine Antarctica**

109 Bed heights referenced to the geoid are obtained from BedMachine Antarctica, Version 3 (Morlighem, 2022). The BedMachine
110 bed height and grounded ice thickness are used to estimate subglacial conditions (Section 3.4). This dataset also contains masks
111 for floating ice, grounded ice, ice-free land, and ocean which are used to select strips based on their overlapping area with the
112 relevant masks.

113 **2.3 Ice velocities**

114 Ice surface velocity for the TGIS is obtained from NASA Making Earth System Data Records for Use in Research
115 Environments (MEaSUREs) mosaicked, 450 m posting, InSAR–Based Antarctic Ice Velocity Map, Version 2 (henceforth,
116 “velocity mosaic”; Mouginit et al., 2012; Rignot et al., 2011, 2017), the MEaSUREs 1 km posting, SAR–Based Annual
117 Antarctic Ice Velocity Maps, Version 1 (henceforth, “annual velocity maps”; Mouginit et al., 2017a), and 250 m posting
118 velocity maps for the austral summer quarters (Oct–Dec and Jan–Mar) that we derived using speckle tracking applied to
119 Sentinel 1 A/B images from TGIS and Pine Island Glacier. The 450 m posting MEaSUREs mosaic is used to obtain masks for
120 ice moving $< 20 \text{ m yr}^{-1}$ for registering strips and for filtering the annual and quarterly velocity maps. The MEaSUREs annual
121 velocity maps obtained for 2011–2015 are variable in their spatial coverage and quality, while the quarterly velocity maps
122 obtained for 2016–2023 have more consistent coverage and better quality. To obtain annual velocity maps from 2011–2023
123 with more consistent quality, we initially take different approaches to filling data gaps and reducing noise in each dataset: for
124 the 2011–2015 annual velocity maps, we take the average of each annual map and the velocity mosaic at each pixel; for the

125 2016–2023 quarterly maps, we take the average of each year’s Oct–Dec map and Jan–Mar map at each pixel. Each resulting
126 filled, noise-reduced annual map from the whole study period is then bilinearly interpolated to the same grid as the annual
127 REMA mosaics and further filtered by taking the median velocity and standard deviation at each pixel throughout the time
128 period and masking out pixels in each year where the velocity differs from the median pixel velocity by more than 2.5x the
129 standard deviation for that pixel, then smoothing the velocity for each year with a 600 m moving mean window. The annual
130 velocity maps are used to flow-shift the annual DSM mosaics to obtain Lagrangian rates of change and to investigate changes
131 in velocity where we observe rapid HB retreat.

132 **2.4 Historical grounding lines**

133 The grounding line from the MEaSURES Antarctic Boundaries for the 2007–2009 International Polar Year (IPY) from Satellite
134 Radar, Version 2 dataset (Mouginot et al., 2017b) is used as a reference grounding line from which to measure changes in the
135 grounding zone position and is henceforth termed 07–09 IPY GL or simply IPY GL. This dataset provides a complete and
136 continuous grounding line derived from a variety of satellite platforms. Additional historical grounding lines are obtained for
137 a long-term visual comparison (Fig. 1) from the MEaSURES Antarctic Grounding Line from Differential Satellite Radar
138 Interferometry, Version 2 for 7 February 1992 to 17 December 2014 (Rignot et al., 2016); however, these are not used for
139 analyses.

140 **3 Methods**

141 We use a variety of previously defined and novel techniques to investigate changes on the TGIS, described below.

142 **3.1 Mapping the hydrostatic boundary**

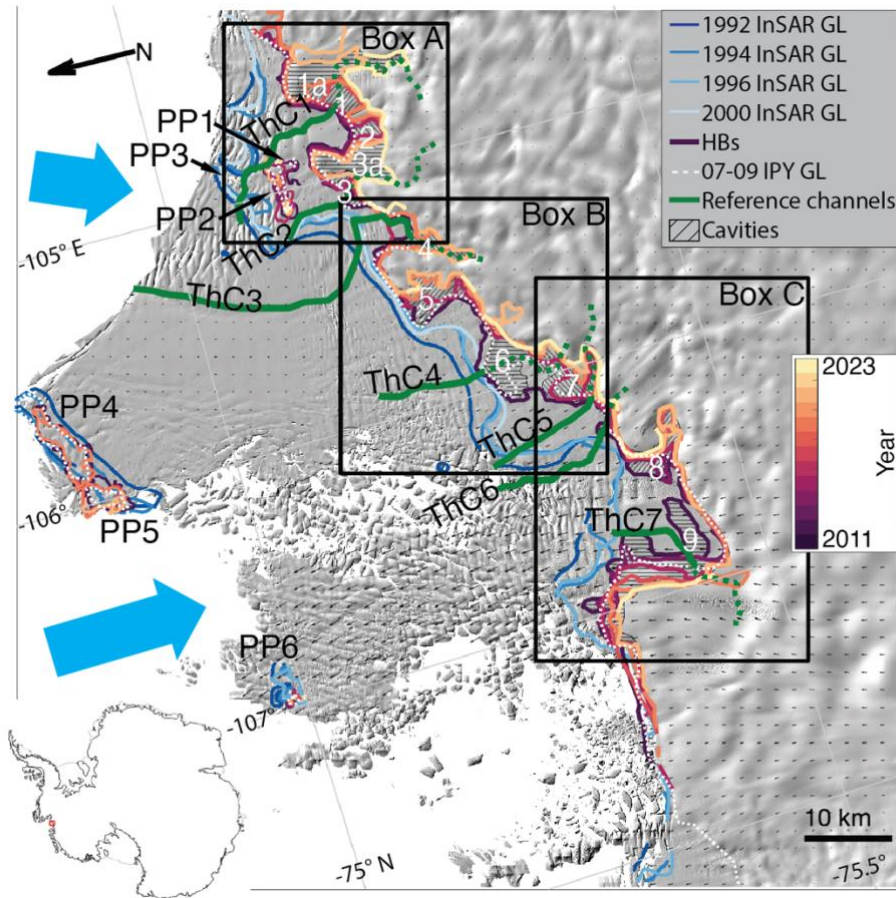
143 The hydrostatic boundary (HB) is defined as the point at which the grounding thickness matches the flotation thickness. The
144 grounding thickness is the distance between the observed ice surface and the bed from BedMachine. Flotation thickness (H_E)
145 is determined from the DSM strip freeboard heights (h) as in Chartrand & Howat (2020, 2023):

$$146 \quad H_E = h \frac{\rho_s}{\rho_s - \rho_i} - H_a \frac{\rho_a - \rho_i}{\rho_i - \rho_s}, \quad (1)$$

147 where ρ_s is seawater density ($1,027 \text{ kg m}^{-3}$), ρ_i is meteoric ice density (918 kg m^{-3}), ρ_a is the firm-air column density (2 kg m^{-3}),
148 and H_a is the thickness of the firm-air column within the freeboard (specifically, the length of the change in firm thickness
149 resulting from compressing the firm column to ice density (Ligtenberg et al., 2011)). The subscript E denotes that H_E is an
150 estimate of ice thickness.

151 We use the 07–09 IPY GL as the basis for where HBs are expected to be mapped. We track HBs at the continental
152 grounding line and six pinning points (PP1–6) delineated in the IPY GL. For each strip, the difference between H_E and the
153 grounding thickness is computed at each pixel. This difference is converted to a contour map, and the coordinates of pixels

154 that lie on the 0 m contours are stored as features representing HBs near the continuous IPY GL and each pinning point. These
 155 features are filtered and simplified for analysis as follows (Fig. S3). For each strip, HB features containing fewer than 25
 156 coordinates are removed, and the remaining HB feature coordinates are smoothed by a 50 point moving mean (a distance of
 157 about 200 m). If the strip has coverage over a given grounding line, a polygon is manually defined to encapsulate the HB
 158 features that most likely represent that grounding line (i.e., the polygon is defined to keep the longest and most continuous HB
 159 features and eliminate small isolated HBs), and points outside of the polygon are eliminated. Then, HB features are combined
 160 by year, and an annual HB with a nominal date of January 1 is manually defined along the most inland features for the
 161 continental GL, and the innermost features for each pinning point, from each year.
 162



163
 164 **Figure 1: REMA 200 m mosaic hillshade for the Thwaites Glacier and Thwaites Glacier ice shelf (TGIS) overlain by historical**
 165 **grounding lines derived from InSAR (curves in shades of blue), the continuous InSAR-derived 07–09 IPY GL (white dashed curve,**
 166 **including pinning points PP1–6), and selected hydrostatic boundaries (HBs) identified in this study (curves in shades of purple,**
 167 **orange, and yellow, with lighter shades representing more recent HBs). The reference channels for the ice-shelf channels and surface**
 168 **depressions identified in this study are shown as solid and dashed green curves, respectively, and labelled ThC1–7. Hatched regions**
 169 **indicate cavities that opened as the HB retreated throughout the study period, labelled 1–9. Large blue arrows indicate the general**

170 location of circumpolar deep water (CDW) influx to the ice-shelf cavity (Dutrieux et al., 2014). The black boxes, Box A–C, indicate
171 the zoomed regions in following figures. Small black arrows represent ice flow.

172

173 3.2 Mapping ice-shelf channels and surface depressions

174 Complementary methods are used to locate persistent curvilinear ice-shelf basal features, including ice-shelf channels, as they
175 are not directly observable using surface elevation alone. Sufficiently large ice-shelf channels (usually > 1 km wide) correspond
176 to surface depressions that are resolvable as stream-like features in the surface topography (Drews, 2015). We do not attempt
177 to verify the presence of entrained meltwater flow in the basal features we identify, and we refer to all consistently mapped
178 basal incisions that are oriented predominantly parallel to flow and associated with surface depressions as ice-shelf channels.
179 Ice-shelf channel locations from Alley et al. (2016) are used to initially query REMA DSM strips.

180 We map surface depressions over both grounded and floating ice by refining the method of using DSM local minima
181 to map surface depressions (Chartrand and Howat, 2020). We compute maps of hypothetical stream channel depth for the ice-
182 shelf surface from strip freeboard heights using the flow accumulation (“flowacc”) function from the MATLAB-based
183 TopoToolbox software (Schwangart & Scherler, 2014). We assume that features with high flow accumulation are surface
184 depressions. We then compare potential depressions with DSM hillshade renderings, eliminating those that align with clear
185 fractures (usually perpendicular to flow), are very short (< ~1 km), or do not intersect the ice shelf.

186 To identify ice-shelf channels on the shelf, we compute flotation thickness from strip freeboard height (Eq. 1) and
187 determine the depth of the hydrostatic ice-shelf draft ($h - H_E$) relative to sea level. We invert the ice-shelf draft by multiplying
188 its depth by -1 , and again compute the hypothetical stream channel depth and flow accumulation across the inverted ice-shelf
189 base, with the locations of stream flow identified as possible ice-shelf channels. As for the surface depressions, we remove
190 spurious features by comparison with DSM hillshade renderings of both the surface and inverted basal topography.

191 Where available, potential ice-shelf channels are verified using IceBridge and pre-IceBridge MCoRDS L2 ice-
192 penetrating radar (IPR) thicknesses (Paden et al., 2011, 2010) (Figs. S4–5). If a basal incision and/or surface depression does
193 not match with a thickness minimum or a previously identified ice-shelf channel (e.g. Alley et al., 2016), we do not rule out
194 that it could be an ice-shelf channel with entrained meltwater flow and look for other evidence of its formation.

195 3.3 Estimating rates of change

196 Time-evolving rates of change are estimated from the annual DSM mosaics within four epochs: 2011–2015, 2016–2019, 2020–
197 2023, and the entire study period from 2011–2023. Within each epoch, rates of change are calculated from all combinations
198 of annual mosaics such that the relevant quantity derived from the earlier mosaic in each combination is subtracted from the
199 later mosaic and divided by the time elapsed between the mosaics. The Eulerian reference frame (fixed coordinate system,
200 denoted by dQ/dt , where Q is the quantity in question) is used over grounded ice, to prevent slope-induced errors, and the
201 Lagrangian reference frame (coordinate system moves with ice flow, denoted by DQ/Dt) is used over floating ice, where height

202 variability is dominated by horizontal advection. The strip-derived annual HB from each year is used to delineate the extent of
 203 floating and grounded ice for each annual mosaic. For grounded ice, we calculate the Eulerian rate of thickness change (dH/dt),
 204 where grounded ice thickness, H , is simply the DSM-derived surface height minus the BedMachine bed height. For floating
 205 ice, we calculate Lagrangian rates of ice-column surface height change (Dh/Dt), thickness change (DH_E/Dt), where flotation
 206 thickness H_E is derived from annual DSM mosaic freeboard heights using Eq. 1, and basal mass loss or gain (M_b). For
 207 Lagrangian calculations, the mosaics are flow-shifted to a common date using the smoothed annual surface velocity maps
 208 (Section 2.3) following the approach of Shean et al. (2019) and Chartrand and Howat (2020). The mosaics are flow-shifted to
 209 1 January of the earliest full year in each epoch (e.g., 1 January 2011 for the 2011–2015 epoch and the full study period).
 210 Lagrangian ice-column thinning can occur as a result of stretching as the ice accelerates (dynamic thinning) or as a result of
 211 surficial or basal ablation, although these mechanisms cannot be attributed by a calculation of DH_E/Dt , which only reflects
 212 how the surface height changes as the column advects due to the hydrostatic assumption.

213 The Lagrangian basal mass change rate M_b (m ice equivalent yr^{-1} , negative values imply basal melt), for floating ice
 214 is determined from mass conservation as:

$$215 \quad M_b = \frac{DH_E}{Dt} + H_E(\nabla \cdot \mathbf{u}) - M_s \quad (2),$$

216 where M_s is the surface accumulation rate (m yr^{-1} , positive for mass gain), and $\nabla \cdot \mathbf{u}$ is the divergence in the column-average
 217 horizontal velocity of the ice \mathbf{u} (m yr^{-1}). As in Shean et al. (2019), the velocity divergence is computed at each time step prior
 218 to flow-shifting the DSM, so the M_b estimate accounts for the flow history of each pixel. The rate of surface accumulation for
 219 Antarctica is obtained from the Regional Atmospheric Climate Model (RACMO) 3p2 (van Wessem et al., 2018) which
 220 provides estimates of M_s for 1979–2016 on a 27 km grid. We bilinearly interpolate the per pixel mean M_s from 2011–2016 to
 221 the mosaic grid coordinates and convert to ice-equivalent mass change rates. The maps of rates of change are smoothed by a
 222 500 m moving mean and extreme values resulting from remaining artefacts from clouds or poorly co-registered strips in the
 223 annual mosaics are filtered out.

224 **3.4 Estimating subglacial conditions**

225 To assess potential spatial relationships between the locations where subglacial hydrologic pathways reach the grounding zone
 226 and align with ice-shelf channels, we derive a map of the subglacial hydraulic potential (Φ) based on observations (Fig. S6) as:

$$227 \quad \Phi = \rho_w g z + \rho_i g H \quad (3),$$

228 where g is the acceleration due to gravity and z and H are equal to the BedMachine bed height and grounded ice thickness,
 229 respectively. Assuming that water is present everywhere at the bed, we again use the TopoToolbox FLOWobj function to
 230 compute the direction that water would flow along the gradient of the hydraulic potential, and the flow accumulation
 231 (“flowacc”) function to find the cumulative number of pixels that contribute to flow in each downstream pixel, and convert
 232 this to the cumulative drainage area, or basal watershed area, for each pixel along the hydraulic potential gradient. While not
 233 intended as an actual estimate of subglacial discharge, this quantity provides a relative metric of where water is likely to be

234 routed in the subglacial system. We compare spatial patterns in inferred subglacial drainage at the grounding zone with the
235 occurrence of mapped ice-shelf channels.

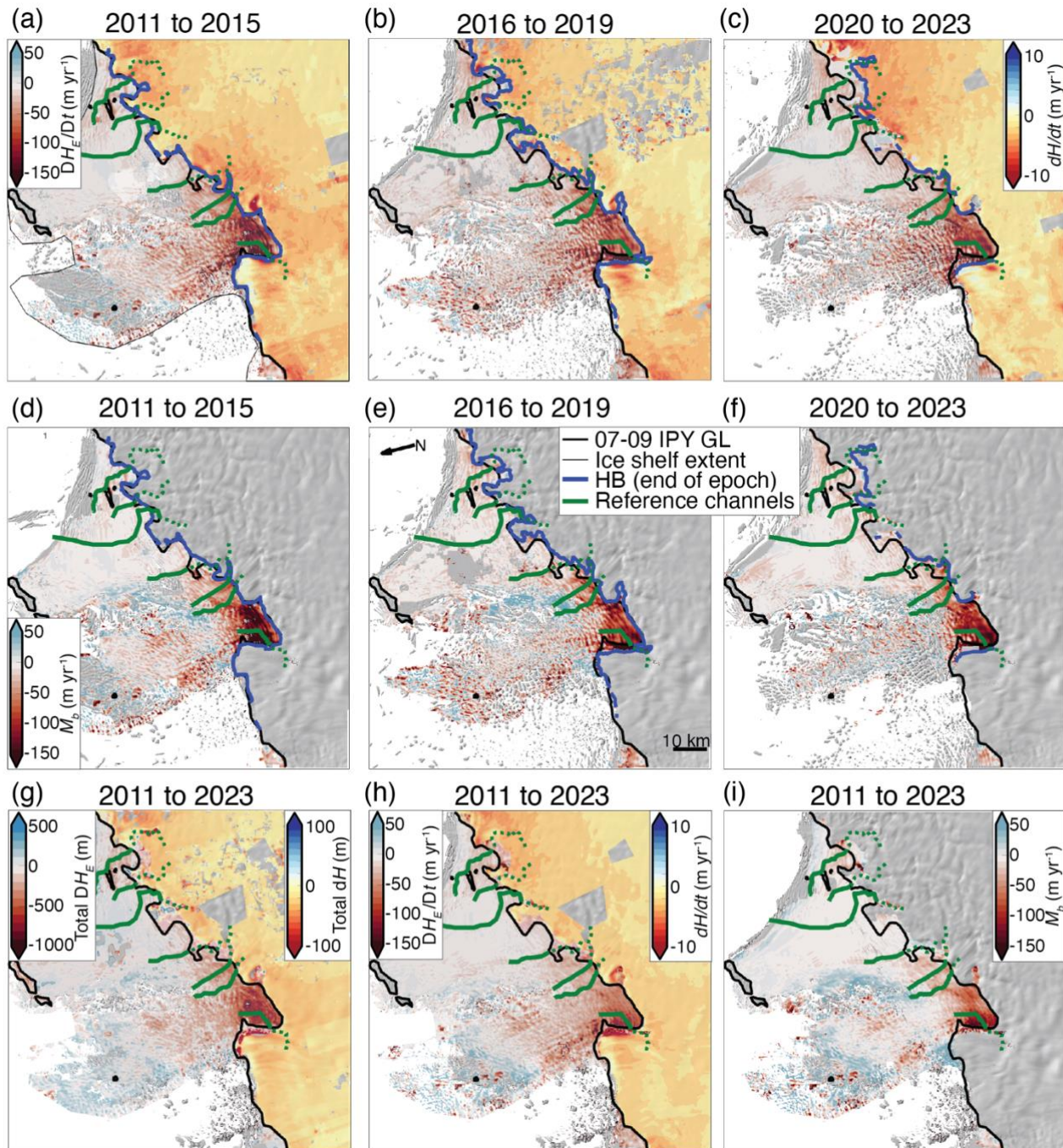
236 **3.5 Uncertainty and sources of error**

237 Estimates of rates of changes in surface height, thickness, and basal mass are subject to uncertainties in the remotely-sensed
238 measurements, model outputs, and assumptions from which they are derived. Our methods follow those of Chartrand and
239 Howat (2020), which showed that uncertainties in DH_E/Dt and M_b range from $\sim 8\text{--}22\text{ m yr}^{-1}$; this is similar to the variability in
240 our estimates (Table 1). We note that M_s is derived from a temporal average of RACMO model output from only part of our
241 study period, which may omit the impact of anomalous precipitation events on our estimates of M_b . However, as we are
242 interested in the spatial variability of grounding zone change over several years, we do not expect the omission of short-term,
243 regional events to significantly impact our results as they will be partially captured in DSM surface heights.

244 Mapping of ice-shelf channel surface depressions and hydrostatic boundaries is subject to uncertainties arising from
245 the hydrostatic assumption, errors in manual delineation, and errors in the BedMachine bed height (Fig. S8). In particular, the
246 hydrostatic assumption may not be valid for portions of the TGIS (Chartrand & Howat, 2023), and an ice shelf's deviation
247 from hydrostatic balance may vary through time in the vicinity of ice-shelf channels (Chartrand & Howat, 2020; Stubblefield
248 et al., 2023). However, hydrostatic imbalance and temporal variations therein are estimated to be a fraction of ice thickness
249 (e.g. Chartrand & Howat, 2023; Stubblefield et al., 2023), and comparable to the BedMachine error in the vicinity and inland
250 of the IPY GL (Fig. S8; Morlighem et al., 2022). As we are interested in relative HB position through time, which occurs over
251 distances longer than ice thickness, rather than absolute HB position, we do not expect hydrostatic imbalance to impact our
252 interpretation of relative HB position inland of the IPY GL. However, the errors in BedMachine bed height increase rapidly to
253 $\sim 400\text{ m}$ between 2–5 km downstream of the IPY GL, so we are not as confident in absolute or relative HB position when it is
254 mapped downstream of the IPY GL. Furthermore, it should be noted that the IPY GL is derived from interferometric SAR
255 imagery and represents the inland limit of ice flexure, whereas the HBs represent the inland limit of hydrostatic balance, which
256 may differ from the limit of flexure by several km (e.g. Fricker et al., 2009). The IPY GL is therefore used only as a reference
257 from which to measure HB change, although it is fortuitous that the IPY GL (delineated from data primarily collected in 2007–
258 2009) represents the grounding line position near the beginning of DSM availability. As described in Section 3.1, manual input
259 is used to delineate the annual HB for each annual epoch, from which relative position is derived; the raw HB features are not
260 subject to manual error. Following masking to remove small, isolated HB features from consideration for the annual HB, points
261 along the most inland HB features from each year are manually selected so that the annual HB consistently represents the
262 grounding line in its most retreated position; we estimate that this manual delineation introduces independent errors $< 200\text{ m}$,
263 or 20 DSM strip pixels, in the position of any point along the annual HB.

264 Manual input for the ice-shelf channel surface depression positions is used to filter out spurious depressions in the
265 TopoToolbox output features, such as those that align with flow-perpendicular crevasses, and to define the reference channel

266 positions based on where a channel was consistently mapped by TopoToolbox in many DSMs. Thus, the reference channel
267 positions represent a manually-defined “average” of each ice-shelf channel’s position through time and may not reflect its
268 position at any given time.



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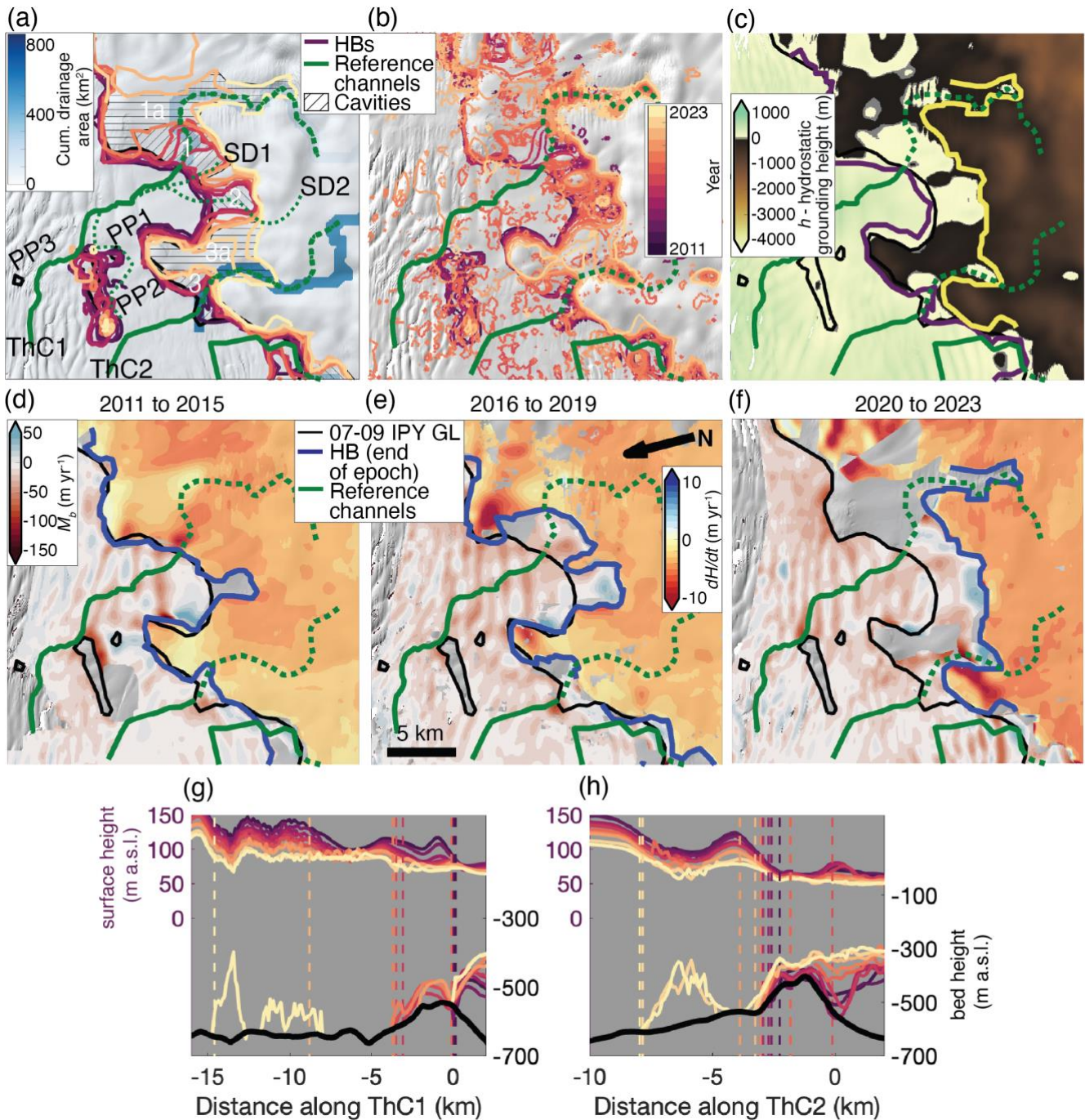
271 **Figure 2: (a–c) Median Lagrangian rate of thickness change (DH_E/Dt) on floating ice (dark red to blue colour scale) and Eulerian**
 272 **rate of thickness change (dH/dt) on grounded ice (dark orange to blue colour scale) for each 4–5 year epoch. (d–f) Median Lagrangian**
 273 **basal mass change rate (M_b , negative for basal melt) for each 4–5 year epoch. All maps (a–f) overlie the most recent annual mosaic**
 274 **hillshade from each epoch. (g) total Lagrangian thickness change (DH_E) on floating ice and total Eulerian thickness change (dH) on**

275 grounded ice for the entire study period. (h) Median DH_E/Dt and dH/dt for the entire study period. (i) Median M_b for the entire study
276 period. Maps for the entire study period (g–i) overlie the 200 m REMA mosaic. All maps show the IPY GL as a black curve and the
277 reference channels as green curves, and (a–f) show the most recent HB in each epoch as a blue curve.

278

279 **4 Results**

280 The TGIS has REMA coverage from November 2010 to December 2022, enabling investigation of time-evolving HB and ice-
281 shelf channel positions as well as ice-column thinning and basal melt rates over the entire ice shelf at unprecedented spatial
282 resolution (Section 4.1). The HB retreated or was stagnant everywhere in the study area (Fig. 1) and ice-column thinning, basal
283 melting and grounded ice thinning dominated rates of change throughout the study period (Figs. 2, S7). We identify nine
284 regions of significant HB retreat and growth of basal cavities, labelled Cavities 1–9 (Fig. 1), discussed in more detail below
285 (Section 4.2). Seven ice-shelf channels that originate near the grounding zone are consistently identified throughout the study
286 period, labelled ThC1–7 (Fig. 1). Six of the channel locations align with inferred subglacial drainage routes (Figs. 3–5a), as
287 discussed in more detail below (Section 4.2).



289

290 **Figure 3:** (a-f) Zoom on Box A from Fig. 1, showing reference channels ThC1-2 (thick green solid/dashed curves) and the 07-09
 291 IPY GL (black curve). (a) Cumulative subglacial drainage area (blue colour scale, explained in Section 3.4) with the smoothed annual
 292 HBs (purple-orange-yellow curves, with less recent, darker features plotted below more recent features as lighter colours) and

293 Cavities 1–3 (hatched regions). Prominent surface depressions that are possibly connected to ThC1 are highlighted by thin dotted
294 green curves. (b) Unfiltered HB features for each year (also on the purple-orange-yellow colour scale used for smoothed annual
295 HBs). Panels (a) and (b) use the REMA v4 200 m mosaic hillshade as the base map. (c) Difference between the 2019 surface height
296 and the hydrostatic grounding height (which is the flotation thickness plus the BedMachine v3 bed height) overlain by the smoothed
297 annual HBs from 2011 (purple curve) and 2023 (yellow curve). (d–f) M_b on the TGIS (dark red to blue colour scale) and dH/dt on
298 grounded ice (dark orange to blue colour scale) for each 4–5 year epoch overlain on the most recent annual REMA mosaic hillshade
299 from each epoch. The most recent HB in each epoch is also plotted (blue curve). Annual surface height (left axis) and ice base (right
300 axis) and BedMachine bed height (right axis, black curve) interpolated to reference channels (g) ThC1 and (h) ThC2. Vertical dashed
301 lines mark the most landward intersection of each year’s HB with the extended channel. Distances are defined from each channel’s
302 intersection with the IPY GL, with positive distance indicating advance and negative distance indicating retreat.

303

304 4.1 Time-evolving rates of basal mass change and ice thickness

305 Figures 2–5 indicate that rates of basal mass change are predominantly negative, indicating basal melting, but are spatially and
306 temporally variable throughout the observation period, as are rates of floating ice-column and grounded ice thickness and
307 surface-height change (Table 1). For ice shelf thickness changes in the Lagrangian frame (DH_E/Dt), thinning refers to change
308 in the same column of ice as it advects with flow, rather than thinning at a fixed coordinate (Eulerian Frame), which we refer
309 to on grounded ice. In general, ice-column thinning and basal melting accelerated from 2011–2015 to 2016–2019 and
310 decelerated slightly in 2020–2023. The banded patterns of positive and negative values visible on the TEIS and TWIT in maps
311 of DH_E/Dt and M_b may be due to changes in ice velocity not accounted for in flow-shifting using annual surface-velocity maps,
312 or due to hydrostatic compensation around growing basal crevasses (e.g. Vaughan et al., 2012).

313 Overall, the TEIS experienced less basal melting than the TWIT. Nevertheless, there was some apparent mass gain
314 in areas of TWIT, particularly in the downstream portion of the TWIT and the shear zone between the TEIS and TWIT (Figs.
315 2, 4–5). We expect that the apparent positive M_b in the downstream portion of the TWIT may be an artefact of hydrostatic
316 disequilibrium due to transient grounding, as evidenced by the presence of isolated HB features in that region (Figs. 5b, S4b).
317 The fastest rates of ice-column thinning and basal melting on the TEIS consistently occur at the eastern ends of pinning points
318 PP2 and PP4, with PP2 located near a zone of rapid HB retreat (Section 4.5) and the opening of Cavity 3 (Fig. 2, Sections 4.2.1
319 and 4.2.4, respectively).

320 The main trunk of the TWIT experienced the most intense basal melting at rates reaching 250 m yr^{-1} in places near
321 the grounding zone throughout the study period (Fig. 5). A closer look within Box B (Fig. 4d–f) shows that consistently high
322 basal melt rates also occurred near the grounding zone in the vicinity of ThC5 and Cavity 7. There is also a flow-parallel band
323 of accelerating ice-column thinning and basal melting along ThC6 near a zone of modest HB retreat along the most pronounced
324 inferred subglacial drainage route (Figs. 2, 5d–f, Section 4.2.3).

325

326 Table 1. Rates of change in each annual mosaic epoch. All values are in units m yr^{-1} (ice equivalent).

	Rate of surface height change		Rate of thickness change				Rate of basal mass change	
	Floating Dh/Dt		Grounded dH/dt		Floating DH_E/Dt		Floating M_b	
	Mean $\pm \sigma$	median	Mean $\pm \sigma$	median	Mean $\pm \sigma$	median	Mean $\pm \sigma$	median
Overall	-0.7 ± 4.9	-0.9	-2.2 ± 0.7	-2.1	-14.2 ± 27.7	-8.6	-6.2 ± 27.9	-3.1
2011–2015	-1.0 ± 6.3	-0.8	-2.8 ± 1.6	-2.6	-21.4 ± 36.9	-11.0	-14.5 ± 37.4	-6.2
2016–2019	-3.0 ± 5.0	-2.0	-2.4 ± 2.2	-2.3	-27.3 ± 35.6	-17.8	-19.0 ± 37.6	-11.1
2020–2023	-2.7 ± 5.1	-1.8	-2.0 ± 2.0	-1.9	-26.2 ± 35.3	-16.2	-16.8 ± 34.8	-9.2

327

328 4.2 Ice-shelf channel–HB interactions

329 As mentioned above, the HB retreated or stagnated relative to its early positions everywhere by 2023, including on pinning
 330 points, with significant variability in the rates of retreat, including some small and temporary areas of advance. In Cavities 6–
 331 9, early HBs appear seaward of the 07–09 IPY GL, likely due to the differences in mapping method, but by 2023 the HB had
 332 also retreated or stagnated relative to the IPY GL everywhere. While few consistent spatiotemporal patterns in HB retreat
 333 emerged, there were no areas of sustained HB advance.

334 We identify several persistent basal incisions and surface depressions along seven ice-shelf channels. We reduce the
 335 impact of noise in individual DSM strips on mapped basal incisions and surface depressions (Fig. S4) by defining fixed
 336 reference locations of seven ice-shelf channels, called ThC1–7, to be where the greatest overlap of these features occurs. The
 337 reference channels are extended above the grounding zone along surface depressions (where present) on grounded ice, enabling
 338 us to measure changes in surface height (Figs. 2–5g–i), thickness (Fig. 6b), and velocity (Figs. 6 and 7) along the reference
 339 channels, as well as changes in their intersections with the HB from each year relative to their intersections with the 07–09
 340 IPY GL (Figs. 3–6), referred to as the ThCX–IPY GL intersection. IceBridge MCoRDS IPR ice-thickness profiles were queried
 341 to verify the presence of the seven ice-shelf channels identified, although data quality along some profiles is poor (Figs. S4–
 342 5). There are also several persistent surface depressions that extend inland of the IPY GL but do not appear aligned with
 343 mapped ice-shelf channels, labelled SD1–7 (Figs. 3a, 5a).

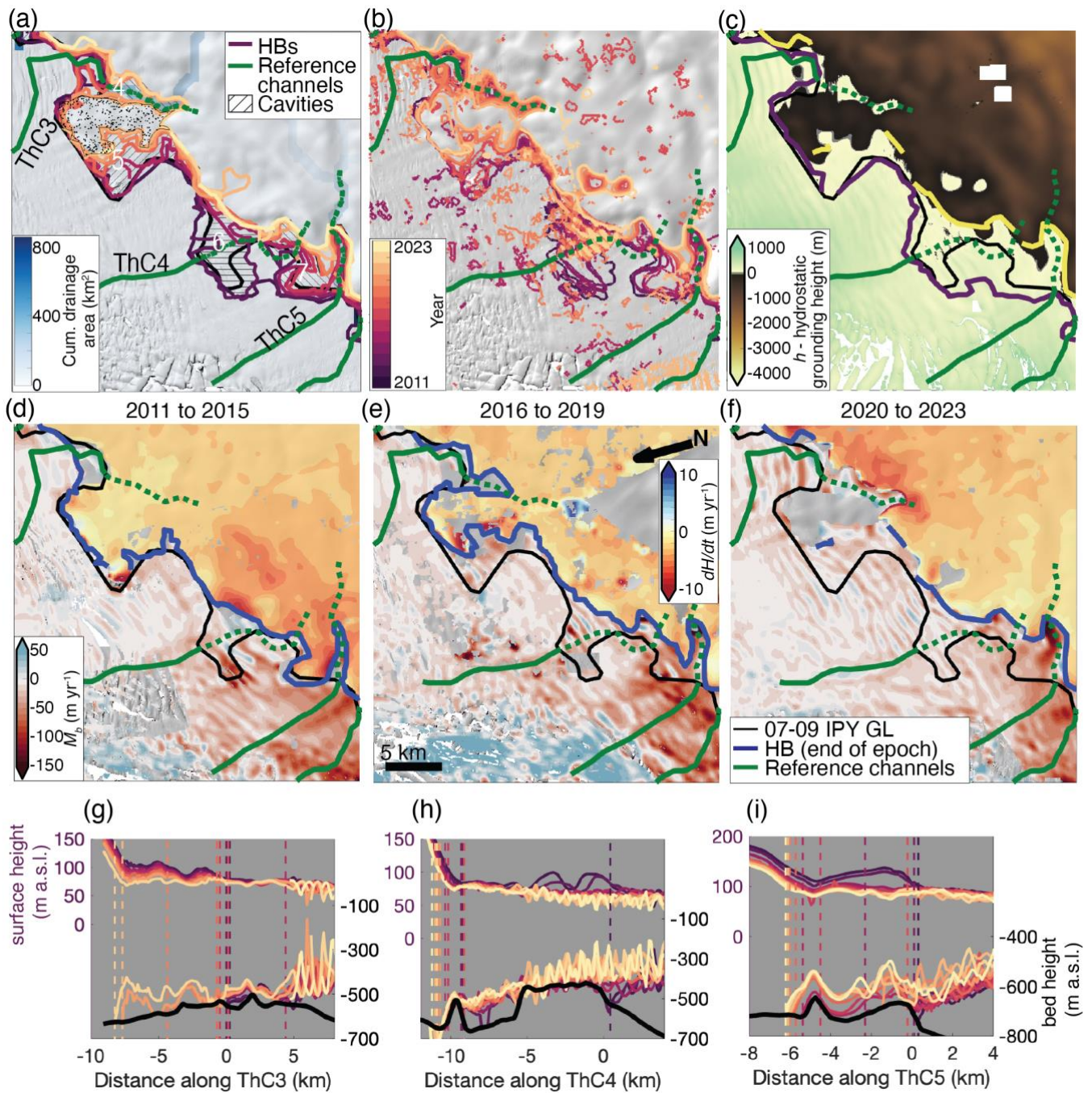
344 A variety of behaviours and interactions between the ice-shelf channels and the grounding zone are observed. We
345 distil these into three major types:

- 346 1. Narrow-cavity HB retreat along a narrow band parallel to an ice-shelf channel (ThC2, ThC3, ThC5; Section 4.2.1)
- 347 2. Wide-cavity retreat along an ice-shelf channel (ThC1, ThC4; Section 4.2.2)
- 348 3. Little to no HB retreat at the inland ends of ice-shelf channels (ThC6, ThC7; Section 4.2.3).

349 Figure 6a shows a time series of the change in position of the HB's intersection along each reference channel and illustrates
350 that Type 1 retreat tended to involve more steady retreat and Type 2 retreat tended to involve cycles of rapid retreat followed
351 by stabilisation. It is important to note that we do not consider these types to be mutually exclusive, as more than one type of
352 HB retreat is observed along several ice-shelf channels throughout the study period.

353 Several TGIS-wide similarities among ice-shelf channels are observed. The grounded ice within ~5 km of each
354 channel's inland end had background thinning rates between -1 to -4 m yr⁻¹ (Figs. 2a–c, 3–5d–f). All channels except for
355 ThC4 originate near areas of high cumulative subglacial drainage area at the grounding zone (Figs. 3–5a). Retreat of the HB
356 exceeding 1 km occurred along all reference channels except for ThC6 and ice-column thinning and basal melting occurred
357 near all channel intersections with the grounding zone within at least one multi-year epoch (Figs. 2–6). Notably, the mean
358 velocity along each HB is slightly slower at the ThCX–HB intersections than in non-channelized portions of the grounding
359 zone (Fig. 6c). However, no strong relationships emerge between changes in velocity and HB retreat rates along all channels;
360 notable correlations between changes in velocity and changes in HB position along individual channels are described in ensuing
361 sections (Fig. 7).

362



363

364 **Figure 4:** (a-f) Zoom on Box B from Fig. 1, showing reference channels ThC3-5 (thick green solid/dashed curves) and the 07-09 IPY
 365 GL (black curve). (a) Cumulative subglacial drainage area (blue colour scale) with the smoothed annual HBs (purple-orange-yellow
 366 curves, with less recent, darker features plotted below more recent features as lighter colours) and Cavities 4-7 (hatched regions).
 367 (b) Unfiltered HB features for each year (also on the purple-orange-yellow colour scale used for smoothed annual HBs). Panels (a)
 368 and (b) overlie the REMA 200 m mosaic hillshade. (c) Difference between the 2019 surface height and the hydrostatic grounding

369 height overlain by the smoothed annual HBs from 2011 (purple curve) and 2023 (yellow curve). (d–f) M_b on the TGIS (dark red to
370 blue colour scale) and dH/dt on grounded ice (dark orange to blue colour scale) for each 4–5 year epoch overlain on the most recent
371 annual REMA mosaic hillshade from each epoch. The most recent HB in each epoch is also plotted (blue curve). Annual surface
372 height (left axis) and ice base (right axis) and BedMachine bed height (right axis, black curve) interpolated to reference channels (g)
373 ThC3, (h) ThC4, and (i) ThC5. Vertical dashed lines mark the most landward intersection of each year’s HB with the extended
374 channel. Distances are defined from each channel’s intersection with the IPY GL, with positive distance indicating advance and
375 negative distance indicating retreat.

376

377 4.2.1 Type 1: Sustained retreat along narrow cavities

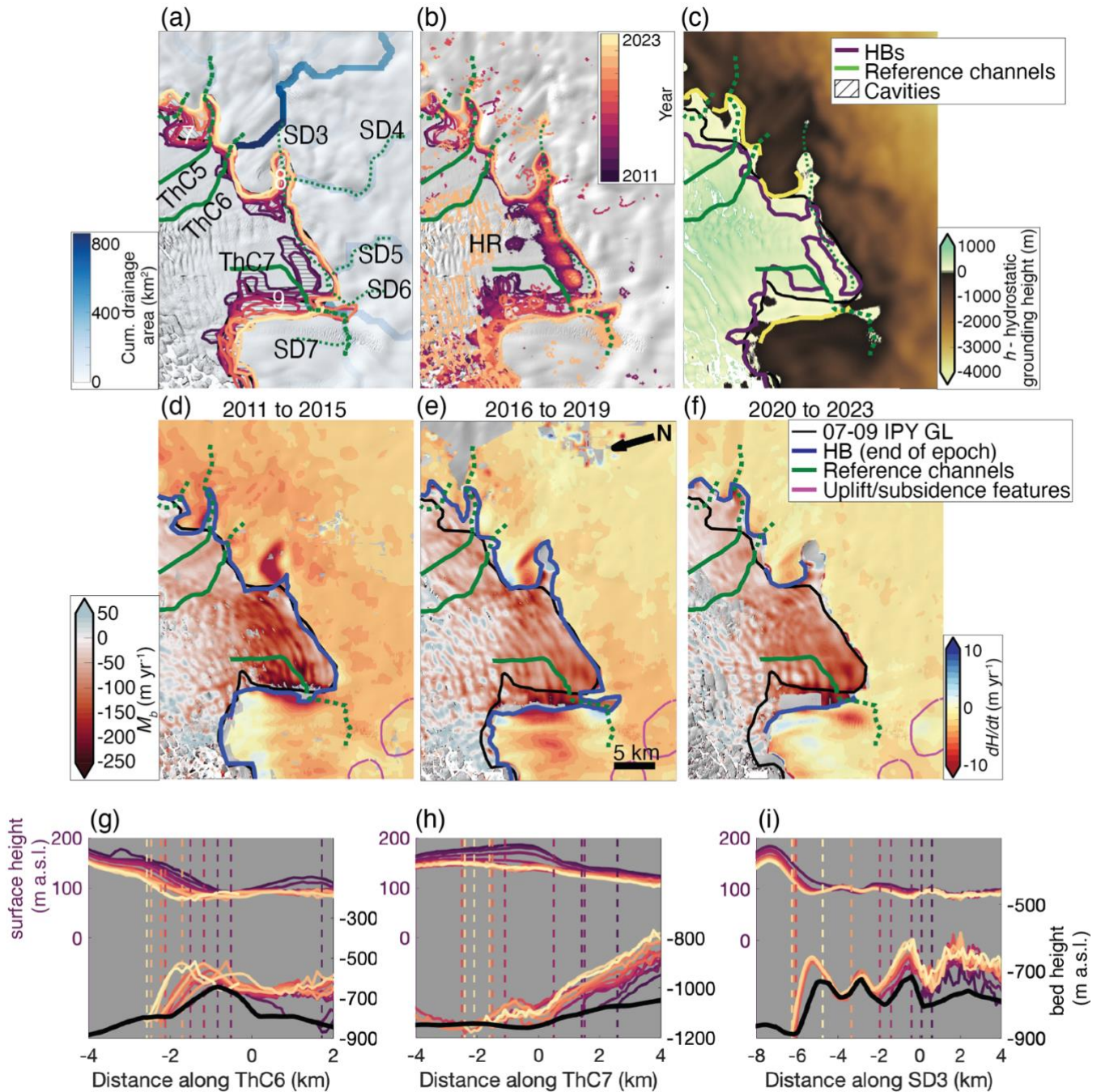
378 Channels ThC2, ThC3, and ThC5, as well as ThC1 in more recent years, were associated with steady, sustained HB retreat
379 along narrow cavities. These features are each directly aligned with an inferred subglacial drainage pathway (Figs. 3–5a) along
380 which HB retreat occurred, such that the cavities may strike oblique to the flow direction, but parallel to the surface depressions.
381 Furthermore, the profiles for these reference channels show large undulations in bed and surface height within 5–10 km of
382 their intersections with the IPY GL (Figs. 3–4g–i) which make it difficult to interpret DH_E/Dt and M_b in these cavities because
383 the flow-shifting does not fully account for height changes due to horizontal advection, leading to alternating bands of apparent
384 ice-column thinning/basal melting and thickening/basal mass gain near the grounding zone (Figs. 3–4d–f).

385 The HB intersection with ThC2 predominantly exhibited Type 1 retreat as the HB retreated in a narrow band along
386 the sinuous ThC2 and its inferred underlying subglacial drainage route throughout the study period (Fig. 3). Cavity 3 widened
387 suddenly following a few years of retreat in a narrow band, temporarily exhibiting Type 2 characteristics, before growing
388 further inland in a narrow band. Between 2021–2023, continued HB retreat opened all of Cavity 3a so that it merged with
389 Cavity 2 (Fig. 3a), coinciding with a 20% increase in surface velocity along ThC2 after relatively steady speeds in earlier years
390 (Fig. 7). Figures 5c and 5h show that Cavity 3a overlies a bedrock ridge, and that the inland end of ThC2 overlies a deepening
391 trough. Rapid ice-column thinning and basal melt occurred near the intersection between ThC2 and the grounding zone
392 throughout the study period (Figs 2, 3d–f).

393 The HB position change along ThC3 exhibits the clearest example of Type 1 retreat (Figs. 1, 4). Figures 4a–b and 4g
394 show that the HB remained relatively stationary relative to ThC3 on a small bedrock ridge through 2018, before retreating
395 rapidly along ThC3 down a retrograde bed slope at a rate of 3.5 km yr⁻¹ between 2018–2020, opening the narrow Cavity 4
396 (Fig. 3a). HB retreat slowed in subsequent years despite the retrograde bed slope continuing inland. Despite the banded pattern
397 in the M_b maps in this region, it appears that basal melt rates generally intensified throughout the study period (Figs. 4d–f). As
398 observed along ThC2, surface velocity along ThC3 accelerated between 2020–2023, although retreat had slowed by this time
399 (Fig. 7).

400 Figure 4 shows that the HB retreated south-eastward along ThC5 at an average rate of ~0.7 km yr⁻¹ between 2013–
401 2018, forming the finger-like southern portion of Cavity 7. During this time, ice-column thinning and basal melt rates reached
402 100 m yr⁻¹ and 60 m yr⁻¹, respectively (Figs. 4d–e), along ThC5 as the HB breached successive bedrock ridges before
403 stagnating on a relatively flat bed about 5–7 km inland of the ThC5–IPY GL intersection (Fig. 4i). Although the HB didn’t

404 retreat much further along ThC5, Cavity 7 widened eastward and merged with Cavity 6. During this time, basal melt rates
405 within Cavity 7 intensified, exceeding 100 m yr^{-1} along ThC5 (Fig. 4f).



407

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410

411

Figure 5: (a-f) Zoom on Box C from Fig. 1, showing reference channels ThC6-7 (thick green solid/dashed curves) and the 07-09 IPY GL (black curve). (a) Cumulative subglacial drainage area (blue colour scale) with the smoothed annual HBs (purple-orange-yellow curves, with less recent, darker features plotted below more recent features as lighter colours) and Cavities 7-9 (hatched regions). Prominent surface depressions SD3-7, some of which are possibly connected to ThC7, are highlighted by thin dotted green

412 curves. (b) Unfiltered HB features for each year (also on the purple-orange-yellow colour scale used for smoothed annual HBs), with
413 an HB in the TWIT labelled “HR” for the “Holland Rumble” that was mapped by Holland et al., (2023). Panels (a) and (b) overlie
414 the REMA 200 m mosaic hillshade. (c) Difference between the 2019 surface height and the hydrostatic grounding height overlain by
415 the smoothed annual HBs from 2011 (purple curve) and 2023 (yellow curve). (d–f) M_b on the TGIS (dark red to blue colour scale)
416 and dH/dt on grounded ice (dark orange to blue colour scale) for each 4–5 year epoch overlain on the most recent annual REMA
417 mosaic hillshade from each epoch. The most recent HB in each epoch is also plotted (blue curve) and uplift/subsidence features
418 identified by Rignot et al. (2024) (magenta curves). Annual surface height (left axis) and ice base (right axis) and BedMachine bed
419 height (right axis, black curve) interpolated to reference channels (g) ThC6, (h) ThC6, and (i) SD3. Vertical dashed lines mark the
420 most landward intersection of each year’s HB with the extended channel. Distances are defined from each channel’s intersection
421 with the IPY GL, with positive distance indicating advance and negative distance indicating retreat.

422

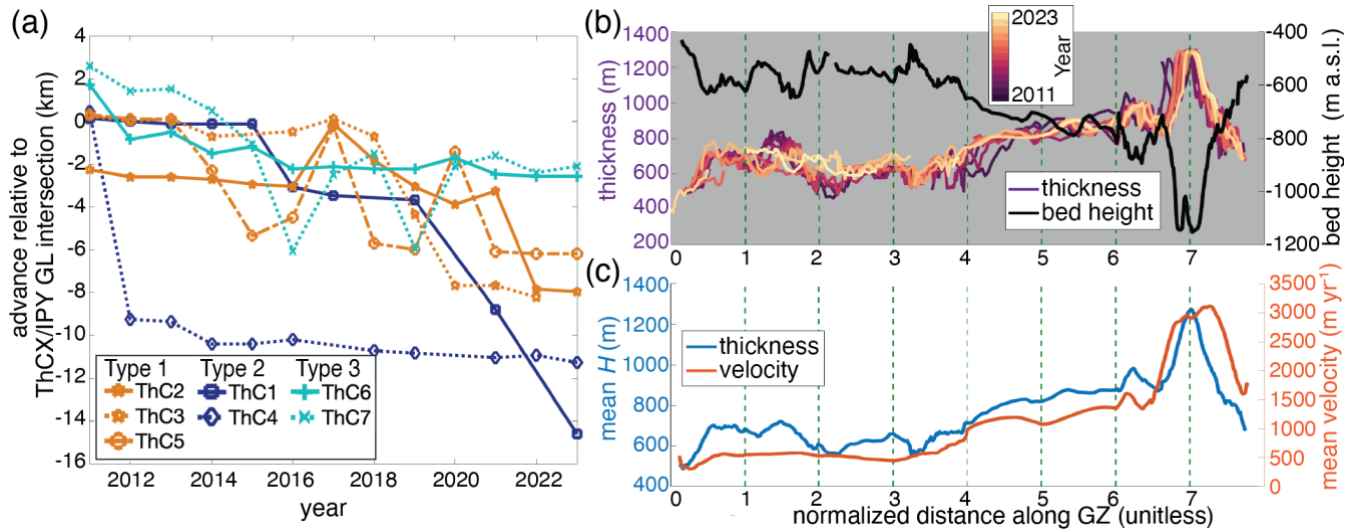
423 4.2.2 Type 2: Wide-cavity retreat

424 Channels ThC1 and ThC4 are associated with sudden, rapid HB retreat off of bedrock highs to form wide Cavities 1a, 2, and
425 6, respectively (Figs. 3–4). Expansion of these relatively large cavities mostly occurred during 2013–2016 as widening across
426 flow.

427 ThC1 and merging surface depressions SD1 and SD2 intersect the IPY GL in a region where the TEIS cavity is deeply
428 embayed. Figure 3a shows that Cavity 1 opened up as the HB retreated suddenly along the ThC1 surface depression and an
429 inferred subglacial channel between 2015 and 2016, and that Cavity 2 opened up as the HB retreated suddenly along SD2
430 (which does not align with an inferred subglacial channel) between 2014 and 2015, forming the lobes that make this region
431 known as the “butterfly” region. At the same time, the velocity along ThC1 accelerated after a period of stability between
432 2011–2015 and continued to accelerate throughout the study period (Fig. 7). As basal melt rates accelerated to near 40 m yr^{-1}
433 by 2016–2019 (Fig. 3e), Cavities 1 and 2 widened but did not extend further inland (Fig. 3a). Cavity 2 merged with Cavity 1
434 between 2021–2022, eliminating the butterfly shape that was prominent in earlier years. After 2019, the HB exhibited Type 1
435 retreat, as the narrower Cavity 1a extended inland along the ThC1 surface depression and underlying inferred subglacial
436 channel at a rate exceeding 2 km yr^{-1} (Figs 3, 6a).

437 ThC4 is situated near the western edge of the TEIS and does not align with an inferred subglacial hydrological route.
438 ThC4 appears to result from the merging of two incisions initiated at two bedrock ridges, the wider of which is located at -5
439 to 0 km along ThC4 in Figs. 4h and 7 (where 0 km is the northernmost ThC4–IPY GL intersection); the narrower ridge is
440 located further south at about -9.2 km in Figs. 4h and 7, near the western end of Cavity 7 (Fig. 4a). Several instances of the
441 sinuous surface depression persist across Cavities 6 and 7 (Fig. S4c), while instances of the basal incision appear to curve
442 around Cavity 6 (Fig. S4a). Figure 4h shows that Cavity 6 opened along ThC4 between 2011–2015 as ice ungrounded from
443 the wide ridge. In subsequent years, Cavity 7 reached its eastern and southern maximum extents and merged with Cavity 6 as
444 the HB retreated off the narrower southern bedrock ridge (Figs. 4a, 4h). Despite basal melt rates consistently exceeding 60 m
445 yr^{-1} near ThC4 (Fig 4d–f, which may be unreliable due to possible intermittent re-grounding, indicated by the isolated HBs in
446 Fig. 4b), the HB did not retreat much further along ThC4, stabilising within a bedrock trough (Fig. 4h). The velocity along
447 ThC4 was highly variable, decelerating by about 5% as Cavity 6 grew, accelerating by about 10% between 2015–2016, slowing

448 again between 2016–2019 as Cavity 7 grew, and speeding up, especially downstream of Cavity 6, as the HB stagnated (Fig.
 449 7).
 450



451
 452 **Figure 6: (a) Time series of HB position along each reference channel, with negative distance indicating retreat, and 0 km marking**
 453 **the ThCX/IPY GL intersection. The orange time series indicate Type 1 retreat, the blue time series indicate Type 2 retreat, and the**
 454 **teal time series indicate Type 3 retreat. (b) Thickness along each annual HB and BedMachine bed height along the IPY GL and (c)**
 455 **mean thickness and velocity along the IPY GL, with unitless distance normalised to the intersection of each year's HB or the IPY**
 456 **GL with each ice-shelf channel ThC1–7.**

457

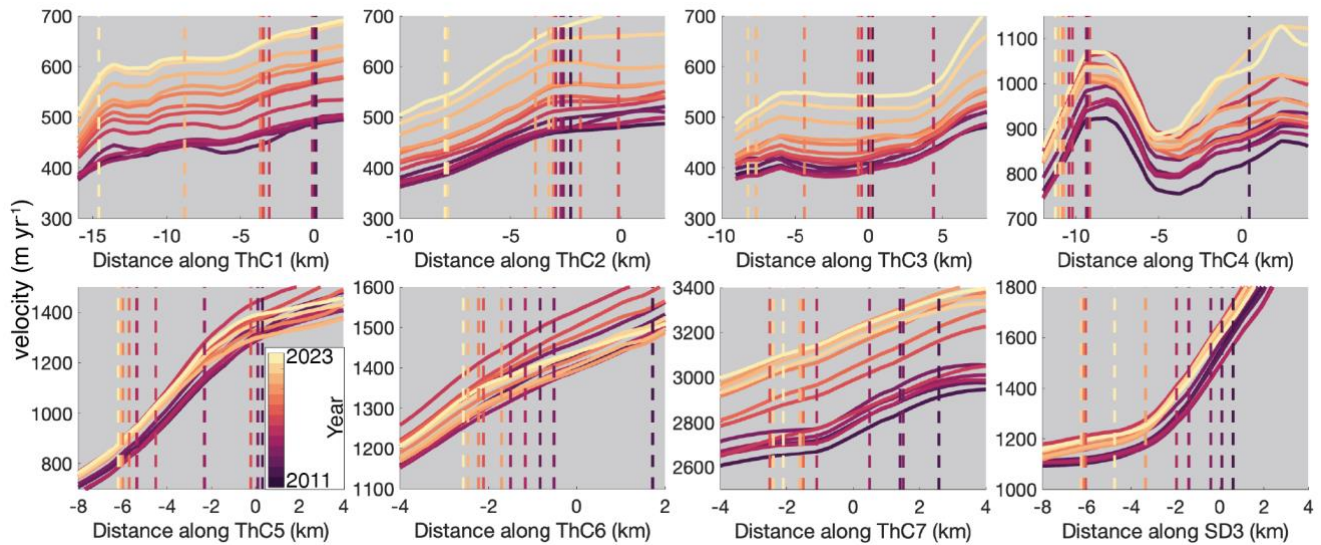
458 4.2.3 Type 3: Little to no HB retreat

459 Ice-shelf channels ThC6 and ThC7 were associated with modest HB retreat that did not fit into Type 1 or Type 2 cavity shapes,
 460 despite their alignment with inferred subglacial drainage routes (Fig. 5).

461 ThC6 is aligned with the strongest inferred subglacial channel just east of the TWIT main trunk, but the surface
 462 depression does not appear to extend inland of the IPY GL (Figs. 4a, S4c). Thus, we manually extended the landward end of
 463 the ThC6 reference channel about 5 km inland of the grounding zone to show retreat past the IPY GL. Throughout the study
 464 period, rapid basal melt and grounded and floating ice thinning occurred near the ThC6–IPY GL intersection (Fig. 5d–g). The
 465 HB retreated at a rate of about 0.3 km yr⁻¹ along ThC6 (Fig. 6a), and the small cavity forms a v-shape along the inferred
 466 subglacial channel (Fig. 5a), potentially indicating that Type 1 retreat will occur in the future. ThC6 also experienced among
 467 the widest fluctuations in velocity, although there was no clear relationship between velocity and HB retreat (Fig. 7).

468 The western flank of the TWIT main trunk grounding zone exhibits complex morphology and changes, but relatively
 469 slow rates of retreat in the vicinity of ThC7 and merging surface depressions SD6–7 (Fig. 5a). At ThC7, the HB retreated ~0.4
 470 km yr⁻¹ as Cavity 9 extended westward between 2013–2023 (Fig. 6a), with basal melt rates consistently exceeding 100 m yr⁻¹

471 ¹ (Fig. 5). Notably, in 2016 and 2019, the HB temporarily retreated along a narrow band along the ThC7 surface depression
 472 (Fig. 5i), within the same timeframe as a 300 m yr⁻¹, or 11%, increase in velocity between 2016–2020 (Fig. 7).
 473



474

475 **Figure 7: MEaSUREs annual velocity (2011–2015) and averaged summer quarterly velocities from InSAR (2016–2023) interpolated**
 476 **to reference channels ThC1–7 and SD3. Vertical dashed lines mark the most landward intersection of each year’s HB with the**
 477 **reference channel. Distances are defined from each channel’s intersection with the IPY GL, with positive distance indicating advance**
 478 **and negative distance indicating retreat.**

479

480 4.2.3 Retreat not associated with ice-shelf channels

481 There are a few regions where HB retreat is observed in the absence of ice-shelf channels and/or inferred subglacial drainage
 482 routes. Between ThC3 and ThC4, there is a region where the HB shifts eastward between 2013 and 2021 at a rate of up to 0.6
 483 km yr⁻¹, opening Cavity 5 (Fig. 4). Notably, the 2022 HB connects with Cavity 4, indicating that a larger cavity may have
 484 opened, but there was insufficient coverage to map the 2023 HB in this region (stippled area in Fig. 4a). In 2022, this region
 485 only has coverage from one or two strips (Fig. S2), resulting in only two mappings of HB features from which the annual HB
 486 was manually delineated. Although some regions are covered by few strips in several years, we are more confident in HB
 487 positions that persist or display a pattern over several years, and one year of data does not provide sufficient evidence to
 488 conclude that this entire region was ungrounded in 2022.

489 The main trunk of the TWIT exhibited complex HB changes seemingly independent of ThC7. Merging surface
 490 depressions SD3 and SD4 are identified inland of the south-eastern corner of the embayed grounding zone in the main trunk
 491 and appear to be associated with HB retreat (Fig. 5a, S4c). SD4 aligns with an inferred subglacial channel, but SD3 does not

492 (Fig. 5a). We also identify a surface depression extending from SD3 and SD4 parallel to the IPY GL at the southern grounding
493 zone of the main trunk, but IPR transects M6 and M7 do not indicate that there are corresponding basal incisions (Fig. S5).
494 The surface depression extending across the main trunk may instead be a dynamical response to the transition of flow off the
495 ridge along the southern grounding zone of the main trunk (Fig. S8a). Figures 5 and 6a show that the HB retreated along SD3
496 at an average rate of $\sim 0.8 \text{ km yr}^{-1}$ between 2011–2018, opening the narrow Cavity 8 along an undulating bedrock topography.
497 The fastest retreat rates ($\sim 2 \text{ km yr}^{-1}$) occurred between 2015–2017, followed by relative stability over a bathymetric low after
498 2018. Ice-column thinning and basal melt rates were consistently high at the downstream end of Cavity 8, exceeding 100 m
499 yr^{-1} along the eastern flank of the main trunk in all three multiyear epochs (Figs. 2, 5d–f).

500 Figure 5b shows that the TWIT main trunk contained many small, isolated HBs throughout most of the study period
501 that may indicate intermittent grounding, although the bed height is unreliable here due to the use of indirect measurements
502 for bed heights in ice-shelf cavities (Fig. S8, Morlighem et al., 2020). The annual HBs from the early years of the study period
503 extended eastward in a narrow band between ThC7 and the IPY GL, narrowing the ice-shelf cavity in the centre of the main
504 trunk to only 2.5 km (Figs. 5a–b). By 2013, Cavity 9 had opened and the HB was approximately at the same location as the
505 IPY GL along the southern grounding zone of the main trunk. Furthermore, the HB at the western flank retreated steadily to
506 the west and up a ridge in the basal topography throughout the study period (Figs. 5a, S8a). The western flank of the main
507 trunk also experienced high rates of ice-column thinning and basal melting throughout the study period as Cavity 9 opened
508 (Figs. 2, 5d–f).

509 **4.3 Pinning points**

510 The HBs at the pinning points exhibited a variety of behaviours. Our HBs maps did not capture PP3, although PP1, PP2, and
511 PP4–6 were mapped. The IPY GL did not contain a pinning point in the main trunk of the TWIT, although Holland et al.
512 (2023) track the evolution of an ice rumple near the centre of the main trunk, which disappeared between 2011–2022. We only
513 map PP1 through 2014, and PP2 grew smaller through 2023 (Fig. 3). Pinning points 1 and 2 experienced relatively high rates
514 of ice-column thinning and basal melting at the eastern extent of the IPY GL of PP2 (Figs. 2, 3d–f). Furthermore, ThC1
515 possibly rerouted as these pinning points shrank; from 2011–2014, the surface and basal manifestations of ThC1 curved toward
516 the west, following the western prong of the “y” shape south of PP1, then straightened toward the eastern prong of the “y”
517 shape between 2015–2022 (Fig S4).

518 Pinning points 4 and 5 were mapped throughout the study period, without much change in HB position, and the
519 surrounding ice shelf experienced ice-column thinning and basal melting at rates similar to the rest of the TEIS (Figs. 1–2).
520 We observe possible north-westward growth of PP4 and PP5 but note that BedMachine is poorly constrained here (Fig. S8).
521 Pinning point 6 was also mapped throughout the study period, although it is largely indistinguishable from other small, noisy
522 HBs that are mapped, but filtered out, on the bedrock high in the TWIT (Figs. 1, S4d).

523 As discussed in Section 4.2.3, the unfiltered HBs in the TWIT (Fig. 5b) indicate the presence of many small pinning
524 points, which appear to shrink or disappear over time as the ice thins. We also map an isolated HB near the ice rumple mapped
525 by Holland et al. (2023; labelled “HR” for “Holland Rimple” in Fig. 5b) which disappears by 2014. However, as noted
526 elsewhere, the bed topography is poorly constrained in this cavity so the locations of HBs and basal incisions inferred using
527 the hydrostatic assumption are uncertain. Notably, the TWIT lost an area of $\sim 1270 \text{ km}^2$ between 2011–2012, retreating from
528 potential pinning points near the front, and continued to lose area throughout the study period (Fig. S1).

529 **5 Discussion**

530 This work reveals high-resolution observations of important processes affecting the shape and structure of the Thwaites Glacier
531 and TGIS. We observe evidence for high rates of grounding zone retreat along ice-shelf channels and inland subglacial
532 channels on the Thwaites Glacier and TGIS using REMA DSMs to map ice-shelf channels, surface depressions, and rates of
533 thickness and basal mass change. We observe three major types of retreat along seven ice-shelf channels and associated surface
534 depressions: narrow-cavity retreat, wide-cavity retreat, and little to no retreat. Regions associated with each type of retreat are
535 often collocated with high rates of ice-shelf basal mass loss and ice-column thinning and down-stream of the grounding zone
536 and grounded ice thinning inland, particularly in the 2011–2015 epoch (Fig. 2).

537 **5.1 Basal melt rates**

538 Aside from some differences in magnitude, the general patterns of persistent HB retreat and rapid ice-column thinning and
539 basal melt along the TGIS grounding zone that we observe are in agreement with other recent observations (e.g. Holland et
540 al., 2023; Milillo et al., 2019; Schmidt et al., 2023; Adusumilli et al., 2020). All confirm that M_b is consistently smaller in
541 magnitude on the TEIS than the TWIT, and that more basal melting occurs near the grounding zone than further seaward.
542 Others have also observed and modelled rapid and potentially unstable retreat of the grounding zone, attributed to enhanced
543 basal melting (Joughin et al., 2014; Rignot et al., 2014; Seroussi et al., 2017; Yu et al., 2018; Milillo et al., 2019; Hoffman et
544 al., 2019). Enhanced basal melt rates are in turn attributed to the intrusion of warm Circumpolar Deep Water (CDW) flowing
545 along bathymetric troughs to the grounding zone (Nakayama et al., 2018; Milillo et al., 2019; Hogan et al., 2020). In the TGIS
546 region, CDW intrusion primarily occurs along two bathymetric troughs (indicated in Fig. 1), allowing it to reach the grounding
547 zone of both the TEIS and the TWIT (Dutrieux et al., 2014; Dotto et al., 2022).

548 The modest basal melt rates that we observe in the vicinity of Cavities 1, 1a, and 2 (Fig. 3d–f) are largely in agreement
549 with those observed by the Icefin submersible in the same region (Schmidt et al., 2023). Holland et al. (2023) show modest
550 apparent basal mass gain along the eastern and southern flanks of the TWIT main trunk in 2011 and basal melt rates reaching
551 250 m yr^{-1} along the southwestern boundary in both 2011 and 2022. High rates of apparent basal mass gain in the TWIT main
552 trunk are also inferred by Milillo et al. (2019). We observe basal melt rates reaching 150 m yr^{-1} throughout the main trunk,
553 especially at the western flank, but we observe no basal mass gain at the southern flank. We suggest that the choice of

554 Lagrangian flow-shifting methods can result in apparent mass gain in the TWIT if the time-evolving flow divergence is not
555 accounted for (Fig. S9). Milillo et al. (2019) posit that, as the ice thins and the grounding line retreats, the bending zone where
556 the ice is deflected below flotation before rebounding also retreats, causing changes at the surface to mask the true magnitude
557 of ice thinning and overestimate M_b . With the caveat that BedMachine is poorly constrained in this ice-shelf cavity (Fig. S8)
558 we also see intermittent re-grounding of ice in the raw HB features (Figs. 5b, S4d), which would further complicate the actual
559 hydrostatic rebound, as well as the hydrostatic assumption and assumptions about ice flow. These factors all reduce confidence
560 in the inferred DH_E/Dt and M_b in the TWIT main trunk. While we do not estimate melt rates below grounded ice, we observe
561 a few regions where isolated HBs inland of the continental HB are aligned or collocated with inferred subglacial channels and
562 regions of grounded ice thinning (e.g. in and around Cavities 1a, 6, 8, and 9); these resemble regions of uplift and subsidence
563 mapped by Rignot et al. (2024) which may indicate enhanced subglacial melting upstream of the grounding zone and are
564 discussed further in Section 5.2.

565 **5.2 Hydrostatic boundaries**

566 In agreement with other studies, we find a mix of stagnation and retreat of the HB along the entire coast of the TGIS. The
567 fastest retreat rates are collocated with retrograde slopes in the bed topography, ice-shelf channels that intersect the IPY GL
568 and/or the positions of inferred subglacial channels, and high basal melt rates. Notably, our M_b estimates and HB retreat for
569 the fast-flowing TWIT main trunk align closely with several other studies; we find HB retreat rates of 0.3–0.6 km yr⁻¹ between
570 2011–2019 and basal melt rates reaching 180 m yr⁻¹ as Cavity 9 opened along the western flank. Milillo et al. (2019) showed
571 that the grounding line along the western flank retreated at a rate of 0.6 km yr⁻¹ to the west between 2011 and 2017. Our results
572 also align with those of Bevan et al. (2021), who documented the opening of an ice-shelf cavity along the retreating western
573 flank between 2014 and 2017, and Rignot et al. (2024), who observe a retreat rate of about 0.5 km yr⁻¹ between 2018–2023 in
574 this region. Milillo et al. (2019) attributed the rapid ungrounding at the point labelled “A” in their figures (which falls within
575 Cavity 9) to its prograde slope, which favours CDW intrusion and efficient cavity opening, consistent with plume theory
576 (Jenkins, 2011).

577 ThC2, ThC3, ThC5, and SD3 are associated with Type 1 HB retreat in narrow bands oblique to the flow direction,
578 but parallel to inferred subglacial channels, lending confidence to our predicted subglacial channel distribution and indicating
579 that subglacial melting is strong (Figs. 3–5). Similar retreat along subglacial channels has been observed on
580 Nioghalvfjærdsfjorden Glacier (N79) Ice Tongue in northeast Greenland (Narkevic et al., 2023) and the Petermann Glacier Ice
581 Tongue in northwest Greenland (Ciraci et al., 2023); in both cases, retreat occurred in narrow bands aligned with the direction
582 of ice flow. Hager et al. (2022) showed that the inclusion of channelized drainage into their model increased effective pressures
583 in non-channelized regions near the grounding line, which may increase basal drag and reduce grounding line retreat and mass
584 loss (Yu et al., 2018) and velocities (Gillet–Chaulet et al., 2016; Joughin et al., 2019). We observe little to no retreat where
585 subglacial channelization is not present, which may be due to high points or prograde slopes in the bed topography but could

586 possibly be due in part to enhanced basal friction in the absence of subglacial water or its concentration within subglacial
587 channels. It is expected that subglacial melt rates are higher where discharge of subglacial meltwater occurs (e.g. Le Brocq et
588 al., 2013; Washam et al., 2019). The basal meltwater volume has been estimated at 3.5 Gt yr^{-1} for the 189,000 km^2 Thwaites
589 Glacier drainage basin, with most of the melt occurring within about 50 km of the grounding zone (Joughin et al., 2009). Our
590 study area extends from ~10–100 km inland of the grounding zone, so ample subglacial water is available, and may discharge
591 in the manner we predict (Figs. 3–5a, S4b), forming a collection of ice-shelf channels when it reaches the ice shelf (Section
592 5.3). While we do not investigate evolution of the subglacial cumulative drainage area over time, we posit that any
593 discrepancies in orientation or position among mapped surface depressions and basal incisions may be due to rerouting of the
594 subglacial drainage system.

595 In contrast with the retreat observed along the continuous continental grounding zone and shrinking or ungrounding
596 of pinning points 1, 2, and 3, PP4 and PP5 exhibit signs of growth throughout the study period, particularly with advance to
597 the northwest of their IPY positions (Figs. 1, S4) Indeed, the bed topographic high on which these pinning points rest extends
598 and grows taller to the northwest (Fig. S8a), and some localised thickening is observed as the TEIS flows onto PP5, although
599 the region is dominated by ice-column thinning and basal melting (Fig. 2). Due to gaps in coverage (Fig. S2), it is difficult to
600 tell whether the ice-shelf area to the north of the pinning points is changing. Wild et al. (2022) demonstrate that although PP5
601 is structurally sounder than PP4, the two used to be connected and their separation and the disconnection of the TEIS and
602 TWIT has altered ice flow. This change, along with the advection of thinner and more damaged ice on the TEIS portends
603 ungrounding from the pinning point within the next decade (Wild et al., 2022).

604 The unfiltered and unsmoothed HBs observed throughout the Thwaites Glacier and TGIS provide insight into
605 potential future behaviour. We observe isolated HBs inland of the continental grounding zone, indicating that the ice surface
606 is below the “hydrostatic grounding height” (ice surface height resulting from adding the flotation thickness H_E for all ice to
607 the bed height, Figs. 3–5b–c) above bedrock lows. The bed heights from BedMachine v3 are relatively reliable inland of the
608 IPY GL, with errors $< 50 \text{ m}$ (Fig. S8b), which is similar to the uncertainty in our calculation of H_E , promoting confidence in
609 the existence of pockets where the surface is below the hydrostatic grounding height inland of the grounding zone. Several of
610 the isolated HBs inland of the grounding zone persist over multiple epochs and align with inferred subglacial drainage
611 pathways. The isolated HBs inland of the IPY GL are necessarily located at bed topographic lows, and likely contribute to the
612 cycle of rapid retreat and temporary stabilisation observed along several reference channels. Indeed, Figs. 3–5b show that the
613 continental HB retreated far enough for several cavities to encompass some of these isolated HBs from earlier epochs. Notably,
614 the grounding zone near ThC6, which aligns with the inferred subglacial channel location with the highest flow accumulation,
615 experienced modest HB retreat but high rates of grounded ice thinning and ice-shelf basal melting and ice-column thinning
616 (Fig. 5), potentially foreshadowing the formation of a Type 1 cavity; however, there are few isolated HBs further inland (Fig.
617 5b–c).

618 We also observe isolated HBs throughout the TWIT, indicating that the ice surface is above the hydrostatic grounding
619 height above bedrock highs (Figs. 3–5b–c). In contrast to our certainty in mapping isolated HBs above the grounding zone,

620 BedMachine errors increase rapidly to 400 m downstream of the IPY GL, where the bed is inferred from gravity inversion, so
621 we are less confident in the existence of additional pinning points within the TGIS, with the exception of the “HR” ice rumple
622 which was independently mapped by Holland et al. (2023) using similar methods (Fig. 5b). We expect that the isolated HBs
623 that persist on high points within ~2 km downstream of the IPY GL throughout several years (where BedMachine errors are
624 around 100 m, Fig. S8b), may have been or currently are pinning points. Likewise, we expect sufficiently high points between
625 the isolated HBs inland of the grounding zone (Figs. 3–5b, S8) to serve as temporary pinning points as new cavities open
626 around them as the continental HB retreats.

627 **5.3 Ice-shelf channels**

628 Based on our inferred subglacial drainage pattern and mapped surface depressions and basal incisions, we suggest that all ice-
629 shelf channels identified in this study except ThC4 are subglacially sourced. Ice-shelf channels have been mapped previously
630 on the TGIS by Alley et al. (2016), and several subglacial channels were also identified by Milillo et al. (2019), many of which
631 align with our DSM-derived channel positions. Comparisons between these observations provide insights into the formation
632 of each ice-shelf channel.

633 There is relatively strong evidence that ThC1 is a subglacially-sourced ice-shelf channel. The downstream end of
634 ThC1 is about 6 km away from an ice-shelf channel identified by Alley et al. (2016) (Fig. S10), and its inland end roughly
635 aligns with where Milillo et al. (2019) document the formation of an approximately 1 km wide subglacial channel near Cavities
636 1, 1a, and 2 (points C and D from Fig. 1 in Milillo et al. (2019)) before the grounding line retreated to its 2017 extent. They
637 observed no change in velocity along the subglacial channel, and thus attribute thinning in this region to ocean-induced basal
638 melting rather than dynamic thinning (Milillo et al., 2019; Millgate et al., 2013). Schmidt et al. (2023) confirmed strong basal
639 melting in this region, with the fastest rates along the steep slopes of terraces at the ice-shelf base, consistent with observations
640 at Pine Island Glacier (Dutrieux et al., 2014). Although Schmidt et al. (2023) did not sample at the location of ThC1, their
641 finding that the greatest basal melting occurs along steep basal slopes in this region provides further evidence that ThC1 is a
642 subglacially-sourced channel whose steep sides promote high basal melt rates and retreat along its trunk.

643 Two channels mapped by Alley et al. (2016) roughly align with the locations of ThC3 and ThC4, and a third runs
644 parallel to the end of ThC7 but begins further downstream (Fig. S10). Alley et al. (2016) considered the ice-shelf channels
645 parallel to ThC4 and ThC7 to be subglacially sourced. Our observations support this claim for ThC7 but suggest that ThC4
646 may be a grounding-zone sourced incision as ice flows over local bedrock topographic highs as described in Sections 4.2.2
647 (Fig. 5), although we cannot confirm whether it entrains buoyant plumes. Furthermore, the retreat along SD3 also appears to
648 be coincident with a subglacial drainage channel modelled and mapped by Hager et al. (2022, Fig. 5), although due to the
649 breakdown of the hydrostatic assumption in the TWIT main trunk, we cannot confirm whether this inferred subglacial channel
650 forms an ice-shelf channel in the ice shelf.

651 One of the channels we observe, ThC7, initiates near where two subglacial drainage channels discharge to the ocean
652 (Rignot et al., 2024). Using differential SAR interferometry, Rignot et al. (2024) observed several circular areas ~4–6 km in
653 diameter with time-varying uplift and subsidence (10–20 cm). These features are located above subglacial topographic
654 depressions that abut km-scale subglacial ridges. The major features are all adjacent to prominent subglacial drainage channels
655 and resemble the isolated HBs we infer inland of the grounding zone in and around Cavities 1a, 6, 8, and 9 (Figs. 3–5b). Rignot
656 et al. (2024) conclude that the filling and draining of the more inland features is driven by fluctuations in the subglacial water
657 flow through the nearby channels. For the large ‘bull’s eye’ feature just ~6 km above the grounding zone (see Fig. 4c in Rignot
658 et al. 2024), however, they speculate that the vertical motion is due to tidally-forced seawater intrusion, which they suggest
659 should cause enhanced subglacial melting. They do not specify the magnitude of this melt other than to say it should be much
660 lower than 20 m yr^{-1} . If this non-steady melting is significantly above the background subglacial melt rate, we would expect
661 to see a signature in the long-term thinning rates. Instead, the 2020–2023 elevation change data show thinning of $1\text{--}2 \text{ m yr}^{-1}$
662 in the area surrounding the feature near ThC7 with minor thickening ($< 0.5 \text{ m yr}^{-1}$) near its centre in 2020–2023, providing
663 little or no indication of enhanced subglacial melt (Figs. 5, S7). We also note that dH/dt derived from annual DSM mosaics
664 does not provide the fine temporal resolution (up to sub-daily) over which uplift/subsidence features were observed in this
665 study. We do not observe increased rates of thinning for most of these closed regions, even when they are near the main HB,
666 suggesting that any enhanced subglacial melting due to incursion of seawater may not persist long enough to significantly
667 impact the signal on multi-annual timescales for most of the glacier. Furthermore, Bradley and Hewitt (2024) show through
668 modelling that Thwaites Glacier is likely not susceptible to runaway melting as a result of seawater intrusion processes. An
669 alternate hypothesis is that all of the circular features are driven by subglacial water flow rather than seawater intrusion. This
670 hypothesis is supported by a strong gradient in the hydraulic potential between the grounding zone and the ‘bull’s eye’ feature,
671 which should drive the water toward – not away from – the ocean (Fig. S6). Seawater intrusion is also problematic because it
672 needs to occur over an area where the predominant flow direction should be seaward to accommodate major subglacial
673 outflows. These features likely fill and drain through exchange of water with the adjacent subglacial channels, similar to how
674 lakes located much farther inland fill and drain below Thwaites Glacier (Smith et al., 2017) and Jutulstraumen Glacier (Neckel
675 et al., 2021). If this is the case, the pressure boundary condition where these channels meet the ocean should be subject to tidal
676 modulation (10 kPa) sufficient to explain the observed ~10–20 cm uplift/subsidence ($1\text{--}2 \text{ kPa}$).

677 **6 Conclusions**

678 This study presents novel, time-evolving rates of ice-shelf thickness and basal mass change and proxies for grounding line and
679 ice-shelf channel position on the TGIS derived from high resolution REMA DSM products, providing further evidence linking
680 high basal melt rates along ice-shelf channels to rapid rates of grounding-zone retreat (e.g. Narkevic et al., 2023; Holland et
681 al., 2023; Ciraci et al., 2023). Hydrostatic boundary retreat rates averaging 0.6 km yr^{-1} and at times $> 3 \text{ km yr}^{-1}$ were observed
682 concurrently with persistent ice-shelf channels and basal melt rates as high as 250 m yr^{-1} . The retreat is not fully attributable

683 to the presence of ice-shelf channels, as several regions where HB retreated along reference channels also had deep retrograde
684 bed slopes and/or were likely to be in contact with warm CDW. This study does not deconvolve all potential causes and effects
685 of HB retreat, such as changes in ice velocity through time (e.g. dos Santos et al., 2021), varying subglacial discharge (e.g.
686 Hager et al., 2022), or changing ocean currents (e.g. Holland et al., 2023; Dotto et al., 2022), but supports the hypothesis that
687 ice-shelf channels, whether initiated at the grounding zone or subglacially, are associated with more rapid grounding zone
688 retreat than non-channelized areas. Our observations are consistent with other work that suggests buoyant meltwater plumes
689 can entrain CDW to form plumes with strong basal melting capabilities (e.g. Le Brocq et al., 2013).

690 These results also provide additional evidence for the recent opening of new ice-shelf cavities not associated with ice-
691 shelf channels, as observed by Milillo et al. (2019), Bevan et al. (2021), and Schmidt et al. (2023), and point to the potential
692 for continued, unstable retreat of the grounding zone (e.g. Yu et al., 2018; Joughin et al., 2014), particularly along inferred
693 subglacial drainage pathways. As the grounding zone continues to retreat and subglacial pressures change, we suggest that
694 retreat along existing and/or rerouted subglacial channels that intersect the grounding zone will continue to form narrow, Type
695 1 cavities in the future, complicating the task of accurately predicting future grounding zone retreat.

696 Milillo et al. (2019) point out that several of the newly opened ice-shelf cavities have less than 100 m between the
697 ice-shelf base and the sea floor, and to simulate these basal melt and retreat processes would require a significantly finer spatial
698 resolution than is currently available to ocean models. This methodology can be applied to other ice shelves to further
699 investigate the prevalence of HB retreat at channelized and non-channelized grounding zones to further investigate relevant
700 changes in ice-shelf structure, velocity, basal and subglacial melt rates, and subglacial drainage. This study is an important
701 step towards better understanding these complex and critical regions of the Antarctic ice sheet and the relevant temporal and
702 spatial scales over which these processes occur.

703 **Data and Code Availability**

704 REMA v4.1 2 m strips (DOI:10.7910/DVN/X7NDNY) and 200 m mosaics (DOI: 10.7910/DVN/EBW8UC) are available at
705 the Polar Geospatial Center. The following datasets are available at NSIDC DAAC: BedMachine Antarctica V003 bed heights,
706 firn and ice thicknesses, Eigen-6C4 geoid data (DOI: 10.5067/FPSU0V1MWUB6), MEaSURES Antarctic Boundaries for IPY
707 2007–2009 from Satellite Radar V002 (DOI: 10.5067/AXE4121732AD), MEaSURES Antarctic Grounding Line from
708 Differential Satellite Radar Interferometry V002 (DOI: 10.5067/IKBWW4RYHF1Q), MEaSURES InSAR-Based Antarctica
709 Ice Velocity Map V002 (DOI: 10.5067/D7GK8F5J8M8R), MEaSURES Annual Antarctic Ice Velocity Maps V001 (DOI:
710 10.5067/9T4EPQXTJYW9), ATLAS/ICESat-2 L3A Land Ice Height V005 (DOI: 10.5067/ATLAS/ATL06.005) and V006
711 (DOI: 10.5067/ATLAS/ATL06.006), IceBridge MCoRDS L2 Ice Thickness V001 (DOI: 10.5067/GDQ0CUCVTE2Q),
712 IceBridge ATM L1B Elevation and Return Strength V002 (DOI: 10.5067/19SIM5TXKPGT), IceBridge LVIS-GH L2
713 Geolocated Surface Elevation Product V001 (DOI: 10.5067/RELPCEXB0MY3), and IceBridge Riegl Laser Altimeter L2
714 Geolocated Surface Elevation Triplets V001 (DOI: 10.5067/JV9DENETK13E). The DTU22 MDT model is available at

715 [\(2019\)](#). The CATS2008b tide model (DOI: 10.15784/601235) is available at USAP–DC. RACMO 3p2 data are available at
716 https://www.projects.science.uu.nl/iceclimate/publications/data/2018/vwessem2018_tc/RACMO_Yearly/. TopoToolbox
717 v2.3.1 is available on the Mathworks File Exchange.

718 All code, gridded products generated in this study (annual mosaics, annual velocities derived from the Amundsen Sea
719 quarterly velocities, and rates of change), and shapefiles of the reference channels, HBs, and strips with registration information
720 are freely available at DOI: 10.5281/zenodo.13667120.

721 **Author Contributions**

722 AC conceived the ideas and carried out analyses with support from IH. IH created the annual REMA DSM mosaics, IJ provided
723 the quarterly velocity maps for the Amundsen Sea regions, and BS performed the CryoSat–2 registrations for the REMA DSM
724 strips. AC prepared the manuscript with contributions from all authors.

725 **Competing Interests**

726 Some authors are members of the editorial board of The Cryosphere.

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