

Net retreat of Antarctic glacier grounding lines

Hannes Konrad^{1,2}, Andrew Shepherd¹, Lin Gilbert³, Anna E. Hogg¹,

Malcolm McMillan¹, Alan Muir³, Thomas Slater¹

Affiliations

¹Centre for Polar Observation and Modelling, University of Leeds, United Kingdom

²Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany

³Centre for Polar Observation and Modelling, University College London, United Kingdom

Grounding lines are a key indicator of ice-sheet instability, because changes in their position reflect imbalance with the surrounding ocean and impact on the flow of inland ice. Although the grounding lines of several Antarctic glaciers have retreated rapidly due to ocean-driven melting, records are too scarce to assess the scale of the imbalance. Here, we combine satellite altimeter observations of ice-elevation change and measurements of ice geometry to track grounding-line movement around the entire continent, tripling the coverage of previous surveys. Between 2010 and 2016, 22%, 3%, and 10% of surveyed grounding lines in West Antarctica, East Antarctica, and at the Antarctic Peninsula retreated at rates faster than 25 m/yr – the typical pace since the last glacial maximum – and the continent has lost $1463 \text{ km}^2 \pm 791 \text{ km}^2$ of grounded-ice area. Although by far the fastest rates of retreat occurred in the Amundsen Sea Sector, we show that the Pine Island Glacier grounding line has stabilized - likely as a

23 **consequence of abated ocean forcing. On average, Antarctica’s fast-flowing ice streams**
24 **retreat by 110 meters per meter of ice thinning.**

25

26 Grounding lines are the junction where marine-based ice sheets become sufficiently buoyant
27 to detach from the sea floor and float^{1,2}. Knowledge of this position is critical for quantifying
28 ice discharge³, as a boundary condition for numerical models of ice flow⁴, and as an indicator
29 of the ice sheet’s state during periods of advance or retreat^{1,5}. In Antarctica, grounding lines
30 are of particular interest because ice-shelf thinning and collapse have driven grounding-line
31 retreat and glacial imbalance around the continent⁶⁻⁸. Although Antarctic grounding lines
32 have retreat since the Last Glacial Maximum⁹, the pace of retreat at several Antarctic ice
33 streams has been much higher during the satellite era¹⁰⁻¹⁵ and numerical simulations have
34 indicated that this rapid retreat may be followed by centennial-scale collapse of the inland
35 catchment areas.^{16,17}

36 Tracking the position of ice-sheet grounding lines using satellite observations has relied on
37 three general approaches; identifying mismatch between surface elevation and freeboard
38 determined through buoyancy calculations¹⁸, breaks in the surface slope associated with the
39 transition from grounded to floating ice¹⁹, and the contrast between vertical motion of floating
40 and grounded ice due to ocean tides²⁰. The latter method is by far the most accurate, because
41 it relies on mapping the hinge line – the limit of tidal flexure at the ice surface which is more
42 readily detectable than the grounding line itself²¹. Although grounding-line migration is
43 usually quantified by repeating the above techniques over time¹², the necessary satellite
44 observations have been infrequently acquired, and so estimates exist at only a handful of
45 locations¹²⁻¹⁴. Here, we extend a method^{11,13,22} for detecting grounding line motion from
46 satellite measurements of surface-elevation change and airborne surveys of the ice-sheet

47 geometry to produce the first continental-scale assessment of Antarctic grounding-line
48 migration.

49

50 Changes in the mass of the firn and ice column around the grounding line cause horizontal
51 migration of the grounding line as the area, in which the ice is buoyant, grows or shrinks. We
52 convert surface-elevation rates, $\frac{\partial S}{\partial t}$, obtained from CryoSat-2 observations²³ into rates of
53 grounding-line migration, v_{GL} , at known grounding-line locations (see *Methods* for a detailed
54 derivation):

$$55 \quad v_{GL} = [-(\alpha + (\rho_o/\rho_i - 1)\beta)^{-1}] \frac{\rho_m}{\rho_i} \frac{\partial S}{\partial t}.$$

56 The term in square brackets, which we will refer to as propensity for retreat, takes slopes of
57 the surface (α – also from CryoSat-2) and the bedrock²⁴ topographies (β) in the direction of
58 grounding-line migration as well as the contrast between ocean (ρ_o) and ice densities (ρ_i) into
59 account. The material density ρ_m allows thickness changes to occur at densities between snow
60 and ice. The direction of grounding-line migration is empirically defined based on grounding-
61 line perpendiculars, flow directions, and bedrock inclinations. We restrict our solution to
62 sections where the grounding-line location has been derived from satellite Interferometric
63 Synthetic Aperture Radar (InSAR)²⁵ and where the propensity for retreat is not excessively
64 high (Figure 1). To estimate the overall error, we consider uncertainties in the density
65 assumptions, satellite-derived surface elevation and elevation rate, and bedrock topography.
66 Altogether, we are able to quantify grounding-line migration along 33.4% of Antarctica’s
67 47,000 km grounding line, including 61 glacier basins³ – three times the combined length and
68 four times as many glaciers as mapped in previous surveys.^{11-14,19}

69

70 We estimate that, between 2010 and 2016, 10.7% of the Antarctic grounding line retreated
71 and 1.9% advanced faster than 25 m/yr, the typical rate of ice-stream retreat during the last
72 deglaciation^{26,27}. There were notable regional differences: Whilst the Peninsula matched this
73 overall picture quite well (9.5% retreating, 3.5% advancing), the West Antarctic Ice Sheet saw
74 21.7% retreating (59.4% in the Amundsen Sea Sector) and only 0.7% advancing, whereas the
75 grounding line of the East Antarctic Ice Sheet retreated and advanced in 3.3% and 2.2% of its
76 length, respectively (Table S1). These changes amounted to a net $209 \text{ km}^2 \pm 113 \text{ km}^2$ loss of
77 grounded-ice area per year over the CryoSat-2 period in the surveyed sections, of which the
78 major part took place along the West Antarctic Ice Sheet ($177 \text{ km}^2 \pm 48 \text{ km}^2$).

79 Large-scale patterns of grounding-line retreat (Figure 1) coincided with sectors in which ice
80 streams are known to be thinning (see Figure S1), for example at the Amundsen and
81 Bellingshausen Sea coasts of West Antarctica^{23,28}. This general link is modulated by the local
82 ice-sheet geometry (propensity), which introduces higher spatial variability into the pattern of
83 retreat. Bedrock slopes along swathes of the English Coast and Wilkes Land make these
84 sectors unfavorable for either retreat or advance, despite the relatively large changes in ice-
85 sheet thickness that have occurred. Long sections were in a state of advance in Dronning
86 Maud Land, East Antarctica, where mass gains have been associated with increased snow
87 accumulation^{29,30}.

88 To investigate regional patterns of grounding-line migration, we examine changes within 61
89 drainage basins³ (Figure 1). Highly localized extremes of rapid grounding-line retreat between
90 500 m/yr and 1200 m/yr have occurred at Fleming, Thwaites, Haynes, Pope, Smith, Kohler,
91 and Hull Glaciers, at Ferrigno Ice Stream, and at glaciers feeding the Getz Ice Shelf, all
92 draining into either the Amundsen Sea or the Bellingshausen Sea, where broad sections of the
93 coast have retreated at rates of 300 m/yr and 100 m/yr, respectively. In East Antarctica, Frost
94 and Totten Glaciers have retreated, locally, at rates of up to 200 m/yr, whereas Mercer and

95 Dibble Ice Streams and glaciers of the Budd Coast have advanced, locally, at up to 60 m/yr to
96 230 m/yr.

97 In general, fast grounding-line migration occurs in areas of fast ice flow (Figure S2), and so
98 we also computed average rates of migration in areas where the ice speed exceeds 25 m/yr
99 and 800 m/yr (Figure 1, Table S2). In total, the central portions of ten fast flowing ice streams
100 have retreated at rates of more than 50 m/yr on average (applying either of the two
101 thresholds), including those of the Amundsen Sea, Totten Glacier, and several in the
102 Bellingshausen Sea – areas in which change is known to be driven at least partially by warm
103 ocean water^{31–34}. The fast flowing sections of Lidke Glacier, Berg Ice Stream, and glaciers
104 flowing into Venable and Abbot Ice Shelves have also retreated, albeit at slower average rates
105 of up to 50 m/yr. Elsewhere in East Antarctica, rates of migration are mostly centered around
106 zero, apart from Frost, Denman and Recovery Glaciers which have retreated at average rates
107 between 19 and 45 m/yr, and from Mertz, Budd, and Shirase Glaciers and Slessor Ice Stream,
108 which advanced at average rates between 14 and 48 m/yr.

109

110 Widespread grounding-line retreat has been recorded in the Amundsen Sea Embayment using
111 satellite InSAR^{11,12}, with which we contrast our altimetry-based results. At Thwaites Glacier,
112 the average rate of grounding-line retreat has increased from 340 ± 280 m/yr between 1996
113 and 2011¹² to 420 ± 240 m/yr according to our method (Figure 2A). Retreat at Pine Island
114 Glacier appears to have stagnated at $40 \text{ m/yr} \pm 30 \text{ m/yr}$ during the CryoSat-2 period, after it
115 migrated inland at a rate of around 1 km/yr between 1992 and 2011 as documented by the
116 previous studies¹¹ (Figure 2B). The recent stagnation coincides with a deceleration of thinning
117 from 5 m/yr in ~2009 to less than 1 m/yr across a 20 km section inland of the 2011 grounding
118 line³⁵, which in principle explains the reduced retreat rate. The slowdown in surface lowering
119 could, however, also be due to further ungrounding, and so we first examine this possibility:

120 To maintain contact with the upstream parts of the ~120 km long central trunk, which are in
121 our data thinning at a maximum rate of 2 m/yr (Figure S3), the grounding line would have had
122 to retreat by at least 15 km since 2011 – more than double that of the previous two
123 decades^{11,12}, at a time when thinning has abated across the lower reaches of the glacier. This
124 leads us to conclude that the main trunk's grounding line has stabilized, potentially due to the
125 absence of warm sub-shelf water³⁶ which drove retreat until 2011. This finding is supported
126 by two recent studies^{37,38}, which also report a substantial reduction in the pace of retreat since
127 2011.

128 We also observe a continuation of retreat at other, less frequently sampled ice streams. For
129 example, high local rates of retreat of ~1.2 km/yr in our results on Haynes, Smith and Kohler
130 glaciers are comparable to peak rates of 1.8 to 2.0 km/yr detected by InSAR between 1992
131 and 2014^{12,15}. In the Bellingshausen Sea, slower rates of retreat recorded over the last 40
132 years¹³ are similar to those we have derived; at Ferrigno Ice Stream, rates of retreat remain in
133 the range 50 to 200 m/yr, at Lidke Glacier, Berg Ice Stream, Venable and Abbot Ice Shelves,
134 our rates of retreat are in the range of 10 to 40 m/yr compared to the multi-decadal range of 10
135 to 90 m/yr¹³, and at the Cosgrove Ice Shelf we detect no significant retreat, in agreement with
136 previously observed rates between -40 m/yr and +11 m/yr¹³. In East Antarctica, Totten
137 Glacier is the only location where grounding-line retreat has been documented, and our result
138 of 154 m/yr \pm 24 m/yr retreat in its fast-flowing section is consistent with the maximum rate
139 of 176 m/yr recorded between 1996 and 2013¹⁴.

140

141 We compared rates of ice thickness change and grounding-line migration to assess the degree
142 to which the processes are related (Figure 3). Within ice-stream sections flowing faster than
143 800 m/yr, thickness changes and grounding-line migration were approximately proportional
144 with 110 \pm 6 metres of retreat occurring with each metre of ice thinning. This comparison for

145 the first time informs on the remarkably consistent geometry-driven propensity for retreat at
146 these ice streams, leading to an intimate relation between thinning and retreat, despite very
147 different processes driving them^{30,31,39}. A possible reason for this stable relationship is that the
148 geometry at fast moving ice-sheet margins may be comparable due to the involved processes:
149 The shape of the surface topography is formed by the nonlinear viscous flow of ice and the
150 sliding conditions at the bedrock⁴⁰. In turn, bedrock topography emerges through tectonical
151 evolution and pre-glacial erosion, and is interactively shaped by the ice-dynamical
152 environment via sedimental erosion through overriding and subglacial hydrology⁴¹ and via
153 glacial-isostatic adjustment of the solid Earth to the overlying ice mass⁴². Even though these
154 processes occur on different spatial and temporal scales and depend on many parameters, it
155 appears that the average propensity for retreat can nevertheless be approximated for different
156 geological settings – a convenient proxy relationship that may be used as a benchmark in
157 investigations which cannot rely on detailed glacial geometry or dynamics.

158 We have compiled the first comprehensive record of present-day rates of grounding-line
159 migration around Antarctica, spanning one third of the continent's margin. In the Amundsen
160 and Bellingshausen Sea sectors of West Antarctica and at Totten Glacier, our results
161 complement and extend earlier assessments of grounding-line retreat^{11–14}, and elsewhere we
162 provide the first observations of migration in key sectors, such as in the Getz Ice Shelf and
163 large parts of East Antarctica and the Peninsula (Figure 1). Although most of the grounding
164 line is stable, we estimate that 3.3%, 21.7%, and 9.5% of East Antarctica, West Antarctica,
165 and the Antarctic Peninsula, respectively, are measurably in a state of retreat. By far the
166 largest rates of grounding-line retreat (> 50 m/yr) occur at ice streams flowing into the
167 Amundsen and Bellingshausen Sea which, on average, are retreating at rates of 134 m/yr and
168 57 m/yr, as well as at Totten Glacier, all of which experience glacial change driven by warm
169 ocean water^{32,33,43}, indicating that the ocean as a driver generates fastest retreat today. The

170 alternation between retreat and advance of glaciers in Wilkes Land could be explained by
171 regional drivers of migration being surpassed by local ones in places. There is a robust
172 relationship between changes in ice thickness and grounding-line migration at Antarctic outlet
173 glaciers, indicating that the geometrical propensity for retreat is relatively uniform in areas of
174 fast flow. The extent of our record could be substantially increased with a more detailed map
175 of the grounding-line position, ideally acquired in the same period as satellite altimetry
176 observations. Overall, our method is a novel and potent approach for detecting and monitoring
177 ice-sheet imbalance in Antarctica; it can be used to pinpoint locations which merit more
178 detailed analysis through field campaigns or dedicated InSAR surveys, e.g. where fast
179 migration occurs or a high geometric propensity for retreat prevails.

180

181 **References**

- 182 1. Weertman, J. Stability of the junction of an ice sheet and an ice shelf. *J. Glaciol.* **13**, 3–
183 11 (1974).
- 184 2. Schoof, C. Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *J.*
185 *Geophys. Res.* **112**, F03S28 (2007).
- 186 3. Rignot, E. *et al.* Recent Antarctic ice mass loss from radar interferometry and regional
187 climate modelling. *Nat. Geosci.* **1**, 106–110 (2008).
- 188 4. Docquier, D., Perichon, L. & Pattyn, F. Representing Grounding Line Dynamics in
189 Numerical Ice Sheet Models: Recent Advances and Outlook. *Surv. Geophys.* **32**, 417–
190 435 (2011).
- 191 5. Pollard, D. & DeConto, R. M. Modelling West Antarctic ice sheet growth and collapse
192 through the past five million years. *Nature* **458**, 329–332 (2009).

- 193 6. Scambos, T. A., Bohlander, J. A., Shuman, C. A. & Skvarca, P. Glacier acceleration
194 and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophys.*
195 *Res. Lett.* **31**, L18402 (2004).
- 196 7. Shepherd, A. *et al.* Recent loss of floating ice and the consequent sea level
197 contribution. *Geophys. Res. Lett.* **37**, L13503 (2010).
- 198 8. Rignot, E., Jacobs, S., Mouginot, J. & Scheuchl, B. Ice-Shelf Melting Around
199 Antarctica. *Science* **341**, 266–270 (2013).
- 200 9. Bentley, M. J. *et al.* A community-based geological reconstruction of Antarctic Ice
201 Sheet deglaciation since the Last Glacial Maximum. *Quat. Sci. Rev.* **100**, 1–9 (2014).
- 202 10. Rignot, E. J. Fast Recession of a West Antarctic Glacier. *Science* **281**, 549–551 (1998).
- 203 11. Park, J. W. *et al.* Sustained retreat of the Pine Island Glacier. *Geophys. Res. Lett.* **40**,
204 2137–2142 (2013).
- 205 12. Rignot, E., Mouginot, J., Morlighem, M., Seroussi, H. & Scheuchl, B. Widespread,
206 rapid grounding line retreat of Pine Island, Thwaites, Smith, and Kohler glaciers, West
207 Antarctica, from 1992 to 2011. *Geophys. Res. Lett.* **41**, 3502–3509 (2014).
- 208 13. Christie, F. D. W., Bingham, R. G., Gourmelen, N., Tett, S. F. B. & Muto, A. Four-
209 decade record of pervasive grounding line retreat along the Bellingshausen margin of
210 West Antarctica. *Geophys. Res. Lett.* **43**, 5741–5749 (2016).
- 211 14. Li, X., Rignot, E., Morlighem, M., Mouginot, J. & Scheuchl, B. Grounding line retreat
212 of Totten Glacier, East Antarctica, 1996 to 2013. *Geophys. Res. Lett.* **42**, 8049–8056
213 (2015).
- 214 15. Scheuchl, B., Mouginot, J., Rignot, E., Morlighem, M. & Khazendar, A. Grounding
215 line retreat of Pope, Smith, and Kohler Glaciers, West Antarctica, measured with

- 216 Sentinel-1a radar interferometry data. *Geophys. Res. Lett.* **43**, 8572–8579 (2016).
- 217 16. Joughin, I., Smith, B. E. & Medley, B. Marine Ice Sheet Collapse Potentially Under
218 Way for the Thwaites Glacier Basin, West Antarctica. *Science* **344**, 735–738 (2014).
- 219 17. DeConto, R. M. & Pollard, D. Contribution of Antarctica to past and future sea-level
220 rise. *Nature* **531**, 591–597 (2016).
- 221 18. Fricker, H. A. *et al.* Redefinition of the Amery Ice Shelf, East Antarctica, grounding
222 zone. *J. Geophys. Res. Solid Earth* **107**, ECV 1-1-ECV 1-9 (2002).
- 223 19. Horgan, H. J. & Anandakrishnan, S. Static grounding lines and dynamic ice streams:
224 Evidence from the Siple Coast, West Antarctica. *Geophys. Res. Lett.* **33**, 1–6 (2006).
- 225 20. Goldstein, R. M., Engelhardt, H., Kamb, B. & Frolich, R. M. Satellite Radar
226 Interferometry for Monitoring Ice Sheet Motion: Application to an Antarctic Ice
227 Stream. *Science* **262**, 1525–1530 (1993).
- 228 21. Gray, L. *et al.* RADARSAT interferometry for Antarctic grounding-zone mapping.
229 *Ann. Glaciol.* **34**, 269–276 (2002).
- 230 22. Shepherd, A., Wingham, D. J. & Mansley, J. A. D. Inland thinning of the Amundsen
231 Sea sector, West Antarctica. *Geophys. Res. Lett.* **29**, 2–4 (2002).
- 232 23. McMillan, M. *et al.* Increased ice losses from Antarctica detected by CryoSat-2.
233 *Geophys. Res. Lett.* **41**, 3899–3905 (2014).
- 234 24. Fretwell, P. *et al.* Bedmap2: improved ice bed, surface and thickness datasets for
235 Antarctica. *Cryosph.* **7**, 375–393 (2013).
- 236 25. Rignot, E., Mouginot, J. & Scheuchl, B. Antarctic grounding line mapping from
237 differential satellite radar interferometry. *Geophys. Res. Lett.* **38**, L10504 (2011).
- 238 26. Smith, J. A. *et al.* New constraints on the timing of West Antarctic Ice Sheet retreat in

- 239 the eastern Amundsen Sea since the Last Glacial Maximum. *Glob. Planet. Change* **122**,
240 224–237 (2014).
- 241 27. Pollard, D., Chang, W., Haran, M., Applegate, P. & DeConto, R. Large ensemble
242 modeling of the last deglacial retreat of the West Antarctic Ice Sheet: comparison of
243 simple and advanced statistical techniques. *Geosci. Model Dev.* **9**, 1697–1723 (2016).
- 244 28. Pritchard, H. D., Arthern, R. J., Vaughan, D. G. & Edwards, L. A. Extensive dynamic
245 thinning on the margins of the Greenland and Antarctic ice sheets. *Nature* **461**, 971–
246 975 (2009).
- 247 29. Lenaerts, J. T. M. *et al.* Recent snowfall anomalies in Dronning Maud Land, East
248 Antarctica, in a historical and future climate perspective. *Geophys. Res. Lett.* **40**, 2684–
249 2688 (2013).
- 250 30. Boening, C., Lebsock, M., Landerer, F. & Stephens, G. Snowfall-driven mass change
251 on the East Antarctic ice sheet. *Geophys. Res. Lett.* **39**, L21501 (2012).
- 252 31. Jenkins A. *et al.* Observations beneath Pine Island Glacier in West Antarctica and
253 implications for its retreat. *Nat. Geosci.* **3**, 468–472 (2010).
- 254 32. Rintoul, S. R. *et al.* Ocean heat drives rapid basal melt of the Totten Ice Shelf. *Sci. Adv.*
255 **2**, e1601610 (2016).
- 256 33. Holland, P. R., Jenkins, A. & Holland, D. M. Ice and ocean processes in the
257 Bellingshausen Sea, Antarctica. *J. Geophys. Res. Ocean.* **115**, C05020 (2010).
- 258 34. Hogg, A. E. *et al.* Increased ice flow in Western Palmer Land linked to ocean melting.
259 *Geophys. Res. Lett.* **44**, 4159–4167 (2017).
- 260 35. Konrad, H. *et al.* Uneven onset and pace of ice-dynamical imbalance in the Amundsen
261 Sea Embayment, West Antarctica. *Geophys. Res. Lett.* **44**, 910–918 (2017).

- 262 36. Dutrieux, P. *et al.* Strong Sensitivity of Pine Island Ice-Shelf Melting to Climatic
263 Variability. *Science* **343**, 174–178 (2014).
- 264 37. Joughin, I., Shean, D. E., Smith, B. E. & Dutrieux, P. Grounding line variability and
265 subglacial lake drainage on Pine Island Glacier, Antarctica. *Geophys. Res. Lett.* **43**,
266 9093–9102 (2016).
- 267 38. Milillo, P. *et al.* On the Short-term Grounding Zone Dynamics of Pine Island Glacier,
268 West Antarctica, Observed With COSMO-SkyMed Interferometric Data. *Geophys.*
269 *Res. Lett.* **44**, 436-444 (2017).
- 270 39. Joughin, I. *et al.* Continued deceleration of Whillans ice stream West Antarctica.
271 *Geophys. Res. Lett.* **32**, L22501 (2005).
- 272 40. Robin, G. de Q. Surface Topography of Ice Sheets. *Nature* **215**, 1029–1032 (1967).
- 273 41. Smith, A. M. *et al.* Rapid erosion, drumlin formation, and changing hydrology beneath
274 an Antarctic ice stream. *Geology* **35**, 127–130 (2007).
- 275 42. van den Berg, J., van de Wal, R. S. W., Milne, G. A. & Oerlemans, J. Effect of isostasy
276 on dynamical ice sheet modeling: A case study for Eurasia. *J. Geophys. Res.* **113**,
277 B05412 (2008).
- 278 43. Jacobs, S. S., Jenkins, A., Giulivi, C. F. & Dutrieux, P. Stronger ocean circulation and
279 increased melting under Pine Island Glacier ice shelf. *Nat. Geosci.* **4**, 519–523 (2011).
- 280 44. Wessel, P. & Smith, W. H. F. Free software helps map and display data. *Eos Trans.*
281 *AGU* **72**, 441 (1991).
- 282 45. Mouginot, J., Scheuchl, B. & Rignot, E. Mapping of Ice Motion in Antarctica Using
283 Synthetic-Aperture Radar Data. *Remote Sens.* **4**, 2753–2767 (2012).

284

285 **Acknowledgements**

286 We acknowledge the European Space Agency (ESA) for the provision of CryoSat-2 data and
287 ESA's Antarctic_Ice Sheet_cci as well as the UK Natural Environment Research Council's
288 (NERC) Centre for Polar Observation and Modelling (CPOM) for processing of these data.
289 HK. was funded through the NERC's iSTAR Programme and NERC Grant Number
290 NE/J005681/1. AEH was supported by an independent research fellowship (no.
291 4000112797/15/I-SBo) jointly funded by ESA, the University of Leeds, and the British
292 Antarctic Survey. The figures were produced using the Generic Mapping Tool⁴⁴.

293

294 **Author contributions**

295 HK, AS, AEH, and MM designed the study. LG and AM processed CryoSat-2 data. HK, AS,
296 AEH, MM, and TS analyzed the results. HK and AS wrote the manuscript. All authors
297 contributed to revising the manuscript.

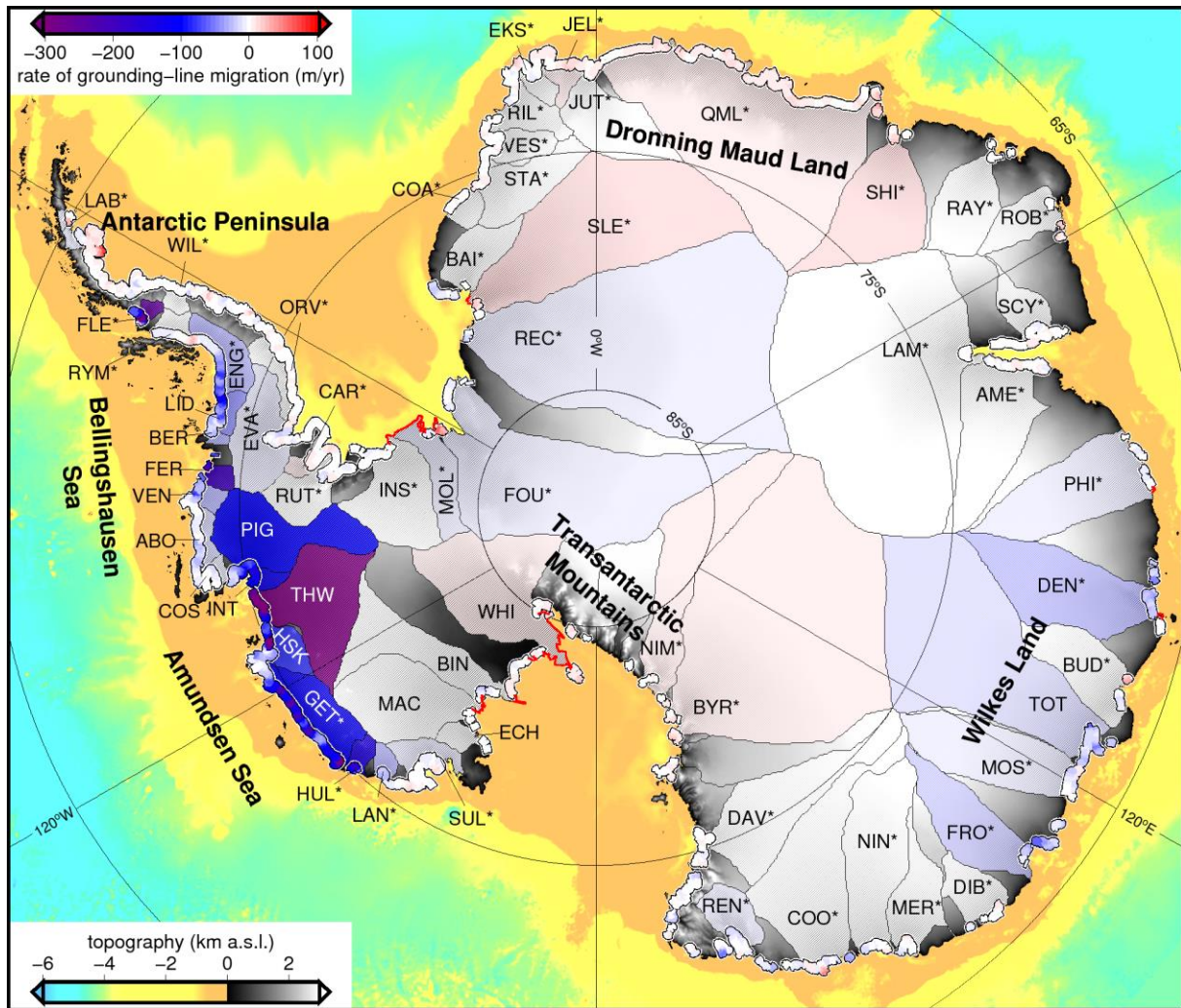
298

299 **Corresponding author**

300 Hannes Konrad, hannes.konrad@awi.de

301

302

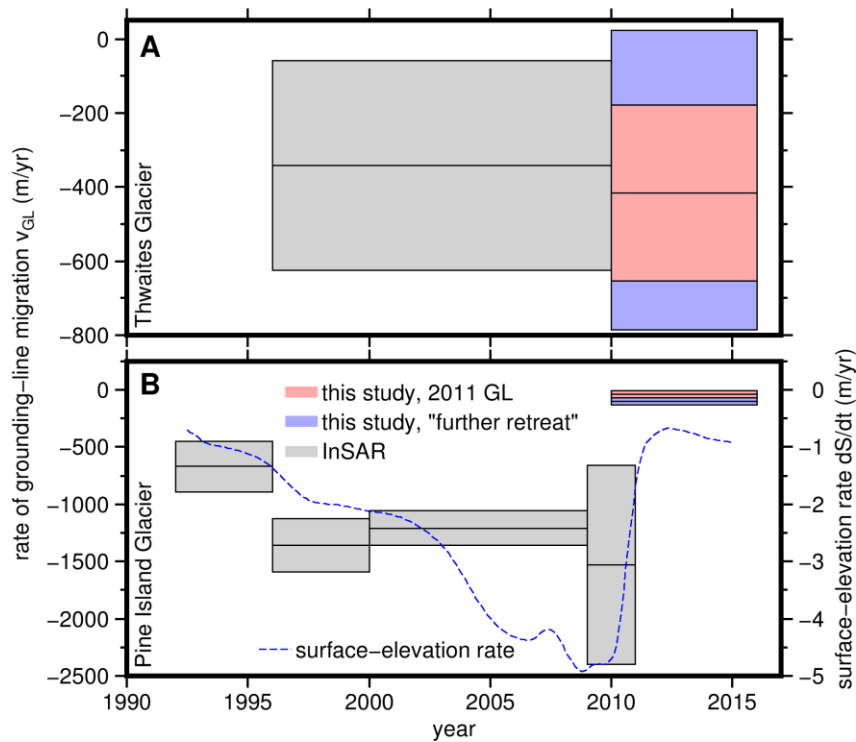


303

304 **Caption Figure 1**

305 **Rates of grounding-line migration between 2010 and 2016 along the Antarctic**
 306 **grounding line⁸ derived from CryoSat-2 and bedrock topography²⁴ observations. Red**
 307 **lines indicate long (>30 km) sections of high propensity for retreat (>500), which we**
 308 **excluded from our analysis. Color-coded basins³ illustrate rates averaged in the areas flowing**
 309 **faster than 25 m/yr (see also Table S2); basins for which we provide the first estimate of**
 310 **grounding-line retreat during the satellite era are marked with an asterisk. Background colors**
 311 **indicate the bathymetry in the ice shelves and ocean and the ice sheet's surface elevation²⁴.**

312



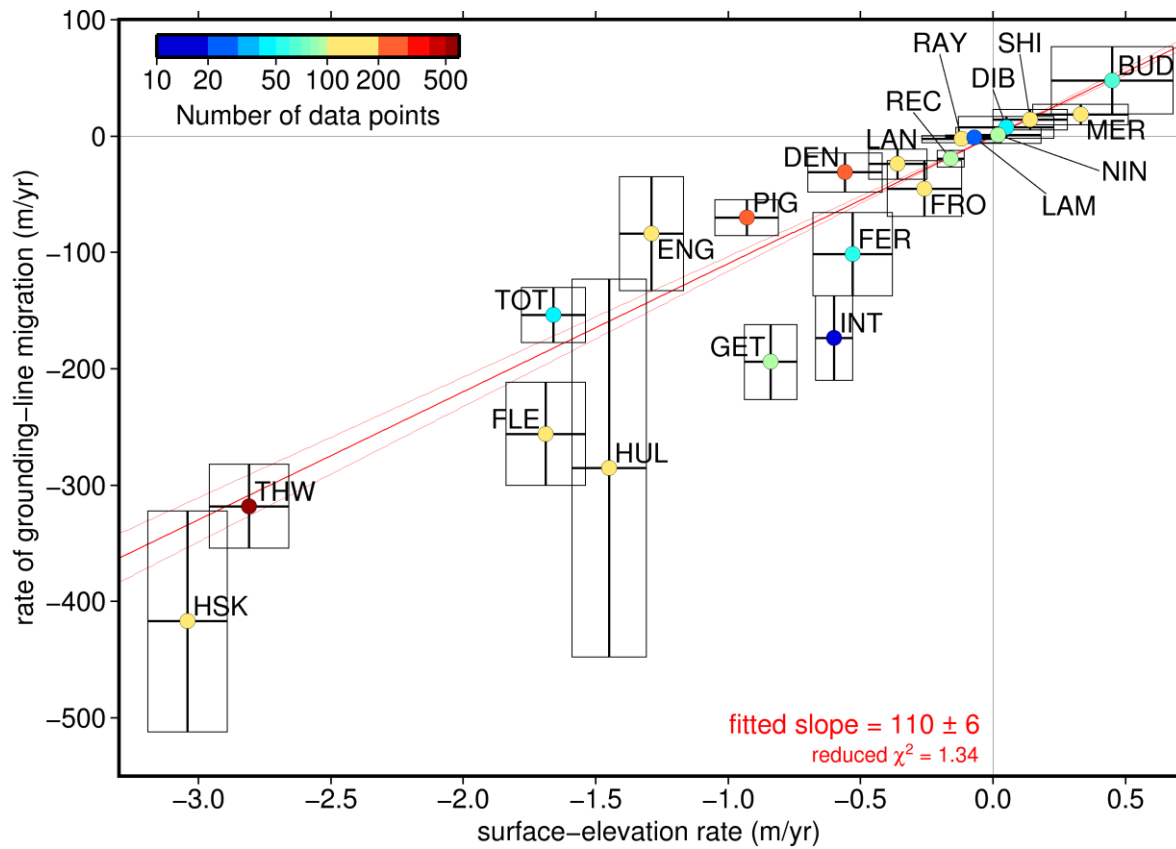
313

314 **Caption Figure 2**

315 **Continuation of rates of grounding-line migration in the Amundsen Sea^{11,12} by our**
 316 **approach.** Rates are averaged across a 43 km wide along-flow swath defined using velocity
 317 observations⁴⁵ at Thwaites Glacier (A) and across the central 10 km of Pine Island Glacier
 318 (B), evaluated along the 2011 grounding line and in a ‘further retreat’ scenario designed to
 319 investigate the impact of strongly dislocated grounding lines (see Methods). Uncertainties
 320 comprise the average local uncertainties in the respective sections and the rates’ spatial
 321 variabilities therein. Also shown in B is a long-term time series of surface-elevation rates
 322 upstream of the 2011 grounding line³⁵.

323

324



325

326 **Caption Figure 3**

327 **Rates of grounding-line migration versus rates of surface elevation.** Both are averaged

328 over fast (>800 m/yr) flowing sections of 21 Antarctic drainage basins. Uncertainties

329 comprise the average local uncertainties of the rates in the respective sections and the spatial

330 variabilities in these sections. Also shown is a straight line fitted in a total least squares sense

331 to these data after re-weighting the average uncertainties by the square root of the number of

332 data points (color coded) in each basin.

333

334 **Methods**

335 We base our analysis of grounding-line migration on a hydrostatic consideration: While the
 336 mass (per unit area) of the ice column σ_i is smaller than that of the water column σ_o in the ice
 337 shelves and vice versa for a (marine-based) ice sheet, they are equal directly at the grounding
 338 line:

339
$$[\sigma_i - \sigma_o]_{GL} = 0 . \quad \text{Eq. (1)}$$

340 At the grounding line and in the interior of an ice sheet, the mass of the ice column can be
 341 expressed in terms of the average density of snow, firn, and ice in the column, ρ_i , the ice
 342 sheet's surface elevation, S , and the bedrock topography, B :

343
$$\sigma_i = \rho_i(S - B) . \quad \text{Eq. (2)}$$

344 Likewise, σ_o can be similarly expressed where the bedrock topography lies below sea level,
 345 substituting S by sea-level height E :

346
$$\sigma_o = \rho_o(E - B) . \quad \text{Eq. (3)}$$

347 Here, ρ_o is the vertically averaged density in the ocean water column. As S and B are usually
 348 referenced to E , we assume $E \equiv 0$.

349 The total temporal derivative of the left-hand side in Eq. (1) containing both partial
 350 derivatives and advective contributions related to grounding-line motion by horizontal
 351 velocity \vec{v}_{GL} must also equal zero.

352
$$\left[\frac{\partial \sigma_i}{\partial t} + \vec{v}_{GL} \cdot \vec{\nabla} \sigma_i - \frac{\partial \sigma_o}{\partial t} - \vec{v}_{GL} \cdot \vec{\nabla} \sigma_o \right]_{GL} = 0 . \quad \text{Eq. (4)}$$

353 Eq. (2) and (3) allow us to replace σ_o and σ_i by the densities and geometric quantities.

354 However, the term $\frac{\partial \sigma_i}{\partial t}$ needs special consideration: If neglecting basal melt, ice thickness

355 changes are consequences of surface processes (surface mass balance, i.e. interaction with the

356 atmosphere; contribution \dot{h}_{surf}), of firn compaction (contribution \dot{h}_{fc}), and of ice dynamics
 357 (contribution \dot{h}_{ice})⁴⁶:

$$358 \quad \frac{\partial(S-B)}{\partial t} = \dot{h}_{\text{surf}} + \dot{h}_{\text{fc}} + \dot{h}_{\text{ice}} . \quad \text{Eq. (5)}$$

359 These contribute differently to the mass σ_i : Ice-dynamical thinning or thickening \dot{h}_{ice} would
 360 change the mass at ρ_i , snow fall variations \dot{h}_{surf} would change it at lower densities, firn
 361 compaction does not affect the mass at all.⁴⁷ We thus introduce the *ad hoc* ‘material density’,
 362 ρ_m to represent which of the above processes is dominant (e.g. McMillan et al.²³):

$$363 \quad \frac{\partial \sigma_i}{\partial t} = \rho_m \frac{\partial(S-B)}{\partial t} . \quad \text{Eq. (6)}$$

364 The introduction of this material density allows us to rearrange Eq. (4):

$$365 \quad \left[f_1 \frac{\partial S}{\partial t} + \vec{v}_{\text{GL}} \cdot \vec{\nabla}(S + f_2 B) \right]_{\text{GL}} = 0 . \quad \text{Eq. (7)}$$

366 Here, it is $f_1 = \rho_m/\rho_i$ and $f_2 = \rho_o/\rho_i - 1 \approx 0.15$. We have neglected any contributions from
 367 bedrock motion (including sea-level rise or similar), as $\frac{\partial B}{\partial t}$ is assumed to contribute ~ 1 cm/yr
 368 at maximum only, if we assume it to be governed by viscoelastic bedrock motion due to paleo
 369 ice-mass changes⁴⁸.

370 Assuming that we know the direction of grounding-line motion \vec{n} (i.e. $\vec{v}_{\text{GL}} = v_{\text{GL}} \vec{n}$), we
 371 define the propensity for retreat as $P = -[\vec{n} \cdot \vec{\nabla}(S + f_2 B)]^{-1}$ (high absolute values indicate a
 372 geometry that favors migration; low values occur in stable areas) and solve for the magnitude
 373 of the rates of grounding-line migration:

$$374 \quad v_{\text{GL}} = \left[f_1 P \frac{\partial S}{\partial t} \right]_{\text{GL}} = - \left[\frac{\rho_m}{\rho_i} \left(\vec{n} \cdot \vec{\nabla} \left(S + \left[\frac{\rho_o}{\rho_i} - 1 \right] B \right) \right)^{-1} \frac{\partial S}{\partial t} \right]_{\text{GL}} . \quad \text{Eq. (8)}$$

375 We obtain S and $\frac{\partial S}{\partial t}$ from CryoSat-2²³ and B from Bedmap2²⁴ and make reasonable
376 assumptions for densities. The two critical points in our approach are the direction of motion
377 \vec{n} and where to evaluate the respective fields. The uncertainty in \vec{v}_{GL} is calculated from the
378 individual uncertainties associated with the altimetry measurements²³, Bedmap2 bedrock
379 topography²⁴, and density assumptions and ranges from 4 cm/yr to 2.5 km/yr, with 90% of all
380 data points between 40 cm/yr and 23 m/yr. Surface-elevation rates and the propensity for
381 retreat along the grounding line are shown in Figure S1.

382

383 *Surface elevation and surface-elevation rates from CryoSat-2*

384 Surface-elevation measurements by CryoSat-2 in SARIn mode between 2010 and 2016 were
385 binned into 5 km x 5 km grid cells. A function quadratic in the component-wise differences x
386 and y to the cell's centre in polar stereographic coordinates and linear in time t since the
387 centre of the time interval for which observations are available is fitted to the data in each cell
388 in a least-squares sense^{23,35},

$$389 \quad f(x, y, t) = a_0 + a_1x + a_2y + a_3xy + a_4x^2 + a_5y^2 + a_6(h) + bt. \quad \text{Eq. (9)}$$

390 The offset between ascending and descending track (heading h) was corrected for by fitting
391 respective parameters $a_6(h)$.⁴⁹ The parameter $a_0 \equiv S$ represents the grid cell's mean surface
392 elevation as used in P in Eq. (8), and the parameter $b \equiv \frac{\partial S}{\partial t}$ is the respective change rate. The
393 uncertainty of the retrieved surface elevation is given by the root mean square of the
394 differences between residues in each grid cell. The uncertainty of the change rates is given
395 through the one-sigma confidence interval of the respective fit parameter. The data were
396 smoothed by a median filter with a 3 cell-wide window (separately for floating and grounded
397 ice in the case of the rates) before they are interpolated (bilinearly for surface elevation and
398 nearest neighbor for surface-elevation rates) onto the 1 km x 1 km Bedmap2 grid (see below).

399 After regridding, the surface gradient was determined, which was again smoothed by a
400 median filter with a 5 cell-wide window. This smoothing as well as that of the bedrock
401 gradients (see below) approximately accounted for the grounding line experiencing different
402 rates solely through the presence of a different geometry as it moved.

403

404 *Bedrock topography from Bedmap2*

405 The bedrock topography B in P in Eq. (8) and its uncertainty between 66 m and 1008 m (not
406 necessarily peaking at the grounding line) were taken from the Bedmap2 data set²⁴ available
407 on a 1 km x 1 km grid. Gradients of bedrock topography were computed and the respective
408 components were smoothed by applying a 5 cell-wide median filter.

409

410 *Density assumptions*

411 The ocean water density was assumed to be $\rho_o = 1027 \frac{\text{kg}}{\text{m}^3} \pm 5 \frac{\text{kg}}{\text{m}^3}$.⁵⁰ Mean ice density in the
412 column was considered to be $\rho_i = 887 \frac{\text{kg}}{\text{m}^3} \pm 23 \frac{\text{kg}}{\text{m}^3}$, allowing for only ice being present or an
413 approximately 100 m thick firn layer on top of the ice at an average thickness of 1000 m at the
414 two extremes of this choice. There are no Antarctic-wide observations available that show
415 how much of a thickness change in a certain area is due to ice-dynamical imbalance or due to
416 (interannual, decadal, or centennial) trends in snow fall; in the absence of such information,
417 we opted for an empirical scheme to define the material density and mostly utilized the
418 surface-elevation rates as a guidance: An absolute rate below 0.3 m/yr was defined to stem
419 from snow fall anomalies only, $\rho_m = 400 \frac{\text{kg}}{\text{m}^3} \pm 50 \frac{\text{kg}}{\text{m}^3}$. Even if this assumption proved
420 dubious in places, it affected our results only lightly as respective low surface-elevation rates
421 mostly did not translate to large rates of grounding-line migration. An absolute surface-

422 elevation rate above 1 m/yr, as well as all the area along the Amundsen Sea Embayment and
423 Getz Ice Shelf, was assumed to stem from ice-dynamical imbalance mainly, so that we set
424 $\rho_m = 850 \frac{\text{kg}}{\text{m}^3} \pm 50 \frac{\text{kg}}{\text{m}^3}$. Anywhere else, we acknowledged that both processes could happen at
425 a similar extent by defining $\rho_m = 625 \frac{\text{kg}}{\text{m}^3} \pm 175 \frac{\text{kg}}{\text{m}^3}$. Such a superposition has, for example,
426 been observed along the English Coast where both ice-dynamical imbalance and decreasing
427 snow fall lead to thinning³⁴. The density uncertainties also accommodate the error arising
428 from assuming a hydrostatic equilibrium at the grounding line where in fact elastic flexure of
429 the stiff ice body leads to a local deviation from this equilibrium.

430

431 *Direction of grounding-line motion*

432 We had to make a relatively strong assumption about the direction of grounding-line motion \vec{n}
433 because there is only one equation for the two-component vector \vec{v}_{GL} . Here, we implemented
434 three different assumptions:

- 435 1. The grounding line advanced in direction of ice flow (positive values of v_{GL}) and
436 retreated in opposite direction (negative values). The same assumption has implicitly
437 been made in other studies by evaluating grounding-line retreat along flow^{11–13}. Flow
438 directions were obtained by bilinearly interpolating surface velocities⁴⁵ to the
439 grounding-line coordinates.
- 440 2. The grounding line advanced (positive) and retreated (negative) perpendicular to the
441 grounding line (represented by the normal vector \vec{n}_{GL} obtained from finite differences
442 of the grounding-line coordinates).
- 443 3. Where the bedrock gradient points seawards from the grounding line ($\vec{n}_{\text{GL}} \cdot \vec{\nabla}B > 0$),
444 the grounding line was assumed to advance (positive) in the direction of this gradient
445 or retreat (negative) in the opposite direction. By that, grounding lines would have

446 migrated towards shallower ocean bathymetry or retreated towards deeper bathymetry.
447 This would be in accordance with the so-called Marine Ice Sheet Instability
448 hypothesis², which considers grounding lines on retrograde slopes inherently unstable
449 in the absence of lateral stresses. Where the bedrock gradient points inwards, a similar
450 argument would not have held anymore, which is why we then opted for the normal
451 vector \vec{n}_{GL} as in option 2.

452 We note that the results from any two of the three options agree within errors for 88.4% of the
453 considered grounding-line sections between options 1 and 3, and for 98.3% between options 2
454 and 3. We consider option 3 to have the strongest physical basis and thus present mainly these
455 data. The only exception is the ‘further retreat’ scenario on Pine Island and Thwaites Glaciers
456 (see below) for which, in the absence of an actual grounding-line position and thus normal
457 vectors upstream of the 2011 position, we chose option 1 as it supplied us with a continuous
458 field of directional vectors.

459

460 *Grounding-line locations and data editing*

461 We evaluated all respective fields (available on the Bedmap2 1 km x 1 km grid, see above)
462 along the grounding line⁸ where it was determined from InSAR, i.e. where the respective
463 sections are also present in the MEaSURES data set²⁵ (46% of the total grounding line), using
464 bilinear interpolation and then solved for the rate of grounding-line migration (Eq. (8)). We
465 note that the grounding-line positions in fast changing areas like the Amundsen Sea
466 Embayment were also among the most recently updated (observations from 2011). At some
467 locations, the last observations were from the 1990s. As no other region showed an equal
468 extent of imbalance as the Amundsen Sea Embayment, we consider respective observations to
469 be sufficiently up-to-date for a well-informed result from our approach.

470 Our assumption of a hydrostatic equilibrium only makes sense where the ice flows into an ice
471 shelf rather than forming grounded ice cliffs; therefore, we rejected data points which do not
472 separate grounded and floating ice as identified using the respective ice sheet/ice shelf/ocean
473 mask in the Bedmap2 data set (29% of all data points) or at which the Bedmap2 bedrock
474 topography is above sea level (12%). Areas which proved to be highly sensitive to surface-
475 elevation change (absolute propensity above 500) were also discarded (15%, Figure 1, and
476 Figure S1. This latter condition excludes, for example, sections of the Siple Coast and Möller
477 and Institute Ice Streams flowing into the Ronne-Filchner Ice Shelf which, though stagnant
478 today, are very lightly grounded¹⁹ and may therefore merit dedicated InSAR monitoring. It is
479 possible that a better resolved glacier geometry could improve the results in these areas.
480 Because we required grounding-line retreat to be caused by thinning and advance to be caused
481 by thickening, we also discarded data points at which $\frac{\partial S}{\partial t}$ and the resulting v_{GL} have a negative
482 relation (i.e. negative propensity) caused by local errors in the assumption of migration
483 direction or the input data (22%). Additional gaps occur where CryoSat-2 does not sample the
484 surface elevation and respective changes (9%). In summary, we discard the solution in about
485 two thirds of the Antarctic margin, manifesting in data gaps which are 12 km wide on
486 average, with 95% of them below 185 km.

487

488 *Determining portions in retreat and in advance*

489 In order to determine the advancing (retreating) fraction of each region (East Antarctic Ice
490 Sheet, Antarctic Peninsula, West Antarctic Ice Sheet, and – as subsets of the latter – West
491 Antarctica’s sectors along the coasts of the Weddell Sea, Ross Sea, Amundsen Sea,
492 Bellingshausen Sea), we summed up the number of points at which we had retained a solution
493 for the rate of grounding-line migration, which were above (below) +25 m/yr (-25 m/yr), and
494 at which the associated uncertainty did not exceed the actual rate. The threshold of 25 m/yr

495 was introduced *ad hoc* based on modelled and geologically derived retreat rates of a West
496 Antarctic paleo ice stream system^{26,27} so that the impact of small rates on these numbers was
497 limited. Detailed numbers are provided in Table S1.

498

499 *Coincidence of grounding-line migration and fast flow*

500 We evaluated how fast grounding-line migration and fast ice flow are spatially related: The
501 histogram in Figure S2 shows how slow-flowing regions as given by MEaSUREs ice
502 velocities⁴⁵ saw less grounding-line migration, and how faster flowing regions were more
503 often experiencing grounding-line migration, also at higher rates. It is also obvious that
504 grounding-line advance was minor compared to retreat.

505

506 *Glacier identification and glacier-wide averages*

507 In order to be able to discuss rates of grounding-line migration averaged on glaciologically
508 meaningful regional scales, we used 65 glacial entities³ and extended them to the recent
509 grounding line by adding area downstream of their defined area using MEaSUREs surface
510 velocities⁴⁵. Both the rates of surface elevation and grounding-line migration were averaged
511 for each of these basins in areas where surface velocities⁴⁵ exceed 25 m/yr and 800 m/yr
512 respectively (Table S2). Additionally, we report respective average uncertainties and – as a
513 measure for extreme values – the 5- and 95-percentiles within these velocity classes.

514 Depending on the magnitude of surface velocities and availability of rates of grounding-line
515 migration according to the above description, some of the 65 glacier basins are not
516 represented by an average value (e.g. Kamb Ice Stream between Whillans (WHI) and
517 Bindschadler (BIN) Ice Streams), leaving us with 61 basins which actually contain results.

518

519 *Comparison with InSAR-derived rates at Pine Island and Thwaites Glaciers (Figure 2) and*
520 *consideration of ‘further retreat’*

521 Published results of grounding-line retreat at Pine Island Glacier (1992–2011) are given as the
522 average along a central section and the standard deviation across that section by Park et
523 al¹¹. To allow comparison, we computed the same quantities for a previously defined cross
524 section on Thwaites glacier from the MEaSURES grounding-line locations from 1996 and
525 2011^{12,25}.

526 We also consider a ‘further retreat’ scenario, which is designed to account for potential inland
527 migration of the grounding line since 2011 and thus to provide an upper bound on retreat rates
528 since 2011. It should be noted, however, that a recent survey confirmed that substantial
529 further retreat has not occurred.³⁸ The ‘further retreat’ scenario is designed as follows: The
530 coordinates of the 2011 grounding-line observation are advected upstream over the time from
531 its acquisition (2011) to the end of our observational period (2016); the direction is chosen to
532 be opposite of the flow direction according to the MEaSURES velocity observations; the
533 magnitude of advection speed is chosen to be 1500 m/yr as this roughly equals the maximum
534 rates obtained from the InSAR analysis in the Amundsen Sea Embayment^{11,12}. Finally, the
535 average rate of grounding-line retreat in the ‘further retreat’ scenario was determined using all
536 Bedmap2 grid cells that lie in the area between the 2011 and the inland advected grounding
537 lines, as well as in the respective cross sections on Pine Island and Thwaites glaciers. Here, it
538 was necessary to choose option 1) for the assumed direction of grounding-line motion, i.e. the
539 direction of the flow velocity (see above). The ‘further retreat’ scenario allows us to assess the
540 maximum impact that an inaccurate grounding-line position (e.g. due to considerable but
541 unmapped retreat since 2011) has on our results.

542 Our estimated uncertainties of the average altimetry-derived retreat rates along these cross
543 sections (at the 2011 grounding line and upstream of it in the ‘further retreat’ scenario)

544 include both the standard deviation and the average propagated uncertainties of the single
545 locations.

546 An overview over the grounding-line situation and the ‘further retreat’ scenario at Thwaites
547 Glacier and Pine Island Glacier can be found in supplementary Figure S3.

548

549 *Fitted empirical relationship between rates of surface elevation and grounding-line migration*
550 *(Figure 3)*

551 We investigated the general relationship between rates of surface elevation and grounding-
552 line migration by focusing on the glacier-wide averages from applying the 800 m/yr threshold
553 on ice flow. The empirical relationship of 110 metres of migration for each metre of thickness
554 change was obtained from a linear total-least-squares fit⁵¹ to these data forced through the
555 origin, for which the average uncertainties had been re-weighted according to the square root
556 of the number of data points going into the averaging of the rates, i.e. the width of the
557 surveyed section, in each basin, divided by their overall mean. The surface-elevation rates
558 were not corrected for vertical displacement of the Earth’s surface due to GIA, see above.
559 However, with present-day rates usually estimated to be below 1 cm/yr⁴⁸, we expect them to
560 have only a minor impact on our analysis and neglected them here.

561

562 *Data availability statement*

563 The rates of grounding-line migration results that support the findings of this study
564 are available from the CPOM data portal, <http://www.cpom.ucl.ac.uk/csopr/>. We
565 acknowledge the authors of all the data sets which we used in this study and which are freely

566 available online. These are the Bedmap2 bedrock topography²⁴, the MEaSUREs Antarctic
567 velocity map^{45,52} and the MEaSUREs Antarctic grounding-line locations^{25,53}.

568

569 **References in Methods section**

570 46. Ligtenberg, S. R. M., Horwath, M., van den Broeke, M. R. & Legrésy, B. Quantifying
571 the seasonal ‘breathing’ of the Antarctic ice sheet. *Geophys. Res. Lett.* **39**, L23501
572 (2012).

573 47. Shepherd, A. *et al.* A Reconciled Estimate of Ice-Sheet Mass Balance. *Science* **338**,
574 1183–1189 (2012).

575 48. Martín-Español, A. *et al.* An assessment of forward and inverse GIA solutions for
576 Antarctica. *J. Geophys. Res. Solid Earth* **121**, 6947–6965 (2016).

577 49. Armitage, T. W. K., Wingham, D. J. & Ridout, A. L. Meteorological Origin of the
578 Static Crossover Pattern Present in Low-Resolution-Mode CryoSat-2 Data Over
579 Central Antarctica. *IEEE Geosci. Remote Sens. Lett.* **11**, 1295–1299 (2014).

580 50. Griggs, J. A. & Bamber, J. L. Antarctic ice-shelf thickness from satellite radar
581 altimetry. *J. Glaciol.* **57**, 485–498 (2011).

582 51. Krystek, Michael and Anton, M. A least-squares algorithm for fitting data points with
583 mutually correlated coordinates to a straight line. *Meas. Sci. Technol.* **18**, 3438–3442
584 (2007).

585 52. Rignot, E., Mouginot, J. & Scheuchl, B. MEaSUREs InSAR-Based Antarctica Ice
586 Velocity Map. (2011). doi:[http://doi.org/10.5067/MEASURES/CRYOSPHERE/nsidc-](http://doi.org/10.5067/MEASURES/CRYOSPHERE/nsidc-0484.001)
587 0484.001

588 53. Rignot, E., Mouginot, J. & Scheuchl, B. MEaSUREs Antarctic Grounding Line from

589 Differential Satellite Radar Interferometry, Version 2. *Boulder, Color. USA. NASA*
590 *Natl. Snow Ice Data Cent. Distrib. Act. Arch. Cent.* (2016).
591 doi:<http://doi.org/10.5067/IKBWW4RYHF1Q>