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1 Trends and connections across the 2 Antarctic cryosphere

3

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10 Abstract

11 After a century of polar exploration, recent satellite observations have painted an altogether new
12 picture of how Antarctica is changing.

13 Satellite observations have transformed our understanding of the Antarctic cryosphere. The
14 continent holds the vast majority of Earth's freshwater, and blankets huge swathes of the Southern
15 Hemisphere in ice. Reductions in the thickness and extent of floating ice shelves have disturbed
16 inland ice, triggering retreat, acceleration, and drawdown of marine terminating glaciers. The waxing
17 and waning of Antarctic sea ice is one of Earth's greatest seasonal habitat changes, and although its

18 extent has increased modestly since the 1970s, variability is high, and there is evidence of longer-
19 term decline.

20 Introduction

21 At the height of austral winter, Antarctica and the surrounding ocean are covered in a 31.6 million
22 km² cap of ice (Fig. 1). Of this, ~18.5 million km² is formed as sea ice when the ocean freezes ¹, 11.9
23 million km² is a near-permanent ice sheet resting on land or the sea floor ², and 1.6 million km² is
24 contained within long-lived ice shelves that are the floating extensions of the continental ice ³. All of
25 Antarctica's ice is mobile, driven by gravity and, where it is afloat, by the atmosphere and the ocean
26 (Fig. 1). Each element plays a unique role in the climate system, for example the grounded ice is
27 Earth's primary freshwater reservoir ⁴, the ice shelves are a major source of ocean fresh water ⁵, and
28 the sea ice is an important factor in the planetary albedo ⁶.

29 The greatest fluctuation in the extent of ice cover in the Southern Hemisphere is due to the seasonal
30 cycle of sea ice formation which is less than a metre thick on average ⁷, and reduces to one sixth of
31 its peak area in summer ⁸. The decadal trend in Antarctic sea ice extent has, nevertheless, been
32 modest ⁹, and the most striking contemporary changes have occurred in other elements of the
33 regional cryosphere. For example, the grounded ice sheet is estimated to have lost 2720 ± 1390 Gt
34 of its mass between 1992 and 2017 ¹⁰, and its peripheral ice shelves are thinning in numerous
35 sectors ^{3,11-13} and collapsing ^{14,15} at the Antarctic Peninsula. These trends reflect global and regional
36 environmental forcing and are related through a variety of processes, each of which is now better
37 understood thanks to the array of satellite observations that have been acquired over recent
38 decades.

39 Here, we analyse the satellite record to examine continental and regional scale trends in the
40 Antarctic cryosphere, including fluctuations in the extent, thickness, and movement of sea ice, ice
41 shelves, and the grounded ice sheet. We show that spaceborne measurements have allowed key

42 events in Earth's recent climate history to be charted in remarkable detail - including the collapse of
43 ice shelves at the Antarctic Peninsula and the drawdown of glacier ice from West Antarctica - and
44 have illuminated the key processes that are driving contemporary change.

45 Grounded ice

46 Fluctuations in the mass of the Antarctic ice sheet arise due to differences between the net snow
47 accumulation and ice discharge. In recent decades, a variety of techniques have been developed to
48 measure changes in the speed, elevation, and weight of the grounded ice. Airborne radar
49 measurements show that the Antarctic ice sheet is up to 4897 m thick, and has the potential to raise
50 global sea level by 58 metres were it to be rapidly discharged ⁴. It overlies terrain of variable geology
51 and relief, and this has influenced both its formation and its contemporary dynamics. The
52 continental-scale pattern of ice flow was first inferred from cartographic ¹⁶ and, more recently,
53 satellite altimeter ¹⁷ records of the ice sheet surface elevation. On this basis, it has been determined
54 that most of Antarctica's ice is routed into the Southern Ocean through around 30 glaciers and ice
55 streams (Fig. 1), each draining a substantial inland catchment ².

56 Grounded Ice Imbalance

57 The stability of Antarctica's ice can be assessed by tracking the movement of its principal glaciers
58 and ice streams. Although few in number, this task is nevertheless beyond the scope of ground
59 surveys because they are vast. The first remote measurements of ice motion were made possible by
60 repeat satellite optical imagery ¹⁸ and, subsequently, by synthetic aperture radar interferometry ¹⁹.
61 Thanks to step increases in the quantity of satellite image acquisitions over time, systematic surveys
62 of ice flow across and around the continent have now been completed ²⁰, revealing anomalous
63 behaviour in much of Marie Byrd Land ^{21,22} and also at isolated sites at the Siple Coast ²³, at the
64 Antarctic Peninsula ²⁴⁻²⁶, and in East Antarctica ^{27,28}. In most of these places, the pace of ice flow has

65 increased during the satellite era, and, when considered as a whole, the rate of ice discharge from
66 Antarctica exceeds inland snow accumulation ²⁹.

67 In addition to changes in ice discharge, fluctuations in ice sheet mass can be detected through
68 satellite measurements of their volume ³⁰⁻³² and gravitational attraction ^{33,34}. Although all three
69 methods lead to similar results at the continental scale, each approach has its merits, and they are
70 now viewed as being complementary. To date, there have been over 150 individual assessments of
71 ice loss from Antarctica based on these approaches ³⁵ and, when collated ^{10,36}, these studies show
72 that the continent has contributed 7.6 ± 3.9 mm to global sea levels since 1992. Almost half ($3.0 \pm$
73 0.6 mm) of this loss occurred during the last 5 years ¹⁰. While the rate of ice loss from the entire
74 Antarctic ice sheet has changed little during the satellite record, the speedup of glaciers in the
75 Amundsen Sea sector has led to accelerated losses from this region ^{37,38}.

76 Satellite radar altimetry is an especially powerful tool for ice sheet glaciology, because the technique
77 can be used to resolve the detailed pattern of imbalance across individual glacier catchments (e.g.
78 ³⁹), around the much of the continent, with monthly sampling, and over multi-decadal periods (Fig.
79 2). This allows signals of short-term variability to be separated from longer-term trends. Although
80 most of Antarctica has remained stable over the past 25 years, there are clear patterns of imbalance
81 in many coastal sectors - for example thickening of the Kamb Ice Stream and thinning of glaciers
82 flowing into the Amundsen Sea and at the Antarctic Peninsula. These changes reflect imbalance
83 between ice flow and snow accumulation within the surrounding catchments. While the pace of ice
84 flow at the Kamb Ice Stream is unusually low ⁴⁰ and has not altered in recent decades, analysis of ice
85 penetrating radar measurements ⁴¹ shows that it stagnated over a century ago. Elsewhere, inland
86 glacier thinning is almost exclusively coincident with contemporaneous ice speedup ^{21,42,43}, indicating
87 that it is dynamic in nature, and with perturbations at their marine termini ⁴⁴, indicating that it has
88 resulted from ocean forcing.

89 Active subglacial Lakes

90 A surprising application of satellite observations has been monitoring the movement of water
91 beneath the Antarctic Ice Sheet. Over three hundred subglacial lakes - bodies of liquid water at the
92 ice sheet base - have been discovered in Antarctica (Fig. 1) using ice penetrating radar ⁴⁵, and these
93 were at first considered to be isolated and stable reservoirs. However, localised and episodic rises
94 and falls of the ice sheet surface were then spotted in satellite interferometric ⁴⁶ and altimetric
95 records ^{47,48}, suggesting otherwise. These fluctuations, amounting to 1 to 10 m height-changes over
96 sub-decadal timescales, are interpreted to be the surface expressions of water transferring between
97 active subglacial lake networks. More than a hundred active lakes have now been identified using
98 this approach ⁴⁹, and monitoring of their evolution has led to improved understanding of how
99 Antarctic subglacial water systems evolve, and the consequences of this variability ⁵⁰. At the
100 Whillans, Mercer, and Recovery ice streams, the Crane and Byrd Glaciers, and in eastern Wilkes
101 Land, for example, more than a decade of satellite measurements have been acquired ⁵¹. Thanks to
102 these data, we now know that in addition to periodically flushing subglacial cavities, the presence of
103 ⁵² and fluctuations in ²⁷ subglacial lake water can lubricate ice flow in parts of the continent.

104 Ice Shelves

105 When Antarctic glacier ice reaches the ocean it often remains intact, forming floating ice shelves in
106 sheltered embayments. Altogether there are more than 300 Antarctic ice shelves, fringing three
107 quarters of the continent and extending the grounded ice area by some 13 % ⁴. Their average
108 thicknesses range from 300 m to 2500 m ⁵³, and peak at the grounding line where they are fed by
109 inland glaciers. Ice shelves can provide mechanical support for the grounded ice sheet upstream,
110 through contact with confining side walls or sea mounts ⁵⁴. Downstream, they thin as the ice
111 spreads, and they gain and lose additional mass primarily through snow accumulation, iceberg
112 calving, and basal ice melting. Basal melting is driven by several processes ⁵ including the formation
113 of high-salinity water during winter sea ice growth, tidal mixing of seasonally warm water, and the

114 intrusion of warm ocean currents into sub-shelf cavities. Meteorological and oceanographic
115 conditions can also lead to surface melting and basal ice freezing. In some cases, it can take more
116 than a thousand years for ice to travel through Antarctic ice shelves from the grounding line to the
117 calving front ⁵⁵, and geological records show that they have been a persistent element of the climate
118 system throughout the Holocene period (e.g. ⁵⁶). Their dependence on a wide range of factors
119 makes ice shelves a sensitive indicator of environmental change ⁵⁷.

120 Ice Shelf Imbalance

121 Trends in ice shelf area, thickness, and flow can be detected using a wide range of satellite sensors,
122 and a host of other properties can be inferred from these measurements. Ice shelf area can be
123 measured using optical and radar satellite imagery, and this has been used, for example, to chart
124 long-term changes in their extent ¹⁴. A series of satellite radar and laser altimeter missions have
125 provided near-continuous observations of ice shelf surface elevation for several decades, and these
126 have formed the basis of ice shelf thickness ⁵⁸ and thickness change ^{3,11,59} estimates on the
127 assumption that the ice is buoyant within the surrounding ocean. These estimates require careful
128 treatment of fluctuations in ocean tide ¹³ and of changes in the firn column thickness ⁶⁰. When
129 combined, measurements of ice shelf area and thickness change allow their volume and mass trends
130 to be derived. Ice shelf flow can be monitored with repeat pass satellite optical ⁶¹ and radar ⁶²
131 imagery and, if contemporaneous changes in both the flow and thickness of ice shelves are available,
132 the rate of steady state ⁶³ and net ^{12,64,65} basal ice melting can be determined.

133 Analysis of ice-shelf surface elevation measurements derived from multi-mission satellite altimetry
134 (Fig. 2) has allowed their decadal mass change and principal environmental forcing mechanisms to
135 be identified ^{3,11,59,66}. While the major Ross, Filchner-Ronne, and Amery ice shelves have remained
136 stable since the 1990s, many ice shelves in West Antarctica have experienced long-term thinning
137 during the same period. In the locations where retreat or thinning have occurred, the grounded ice
138 inland has also been destabilised. The dominant control on this pattern is believed to be the

139 presence (or absence) of warm ocean currents offshore ⁵⁹. Altogether, the volume of Antarctic ice
140 shelves has declined through net overall thinning ($166 \pm 48 \text{ km}^3 \text{ yr}^{-1}$ between 1994 and 2012; ¹¹) and
141 through progressive calving-front retreat of those at the Antarctic Peninsula ($210 \pm 27 \text{ km}^3 \text{ yr}^{-1}$
142 between 1994 and 2008; ³). Combined, these losses amount to less than 1 % of their volume.
143 However, the highest ice shelf thinning rates have occurred in the Amundsen and Bellingshausen
144 Seas ¹², where five have lost between 10 to 18 % of their thickness ¹¹ due to ocean-driven melting at
145 their bases ⁶⁷. The effects of the wider El Niño-Southern Oscillation have also been detected
146 alongside these longer-term trends ⁶⁶.

147 Ice shelf collapse

148 Ice shelves at the Antarctic Peninsula (Fig. 3) are especially vulnerable, because they are situated at
149 the most northerly latitude on the continent, where temperatures are relatively high and
150 summertime melting is common. In recent decades, a number of these ice shelves have
151 disintegrated in part or entirely ¹⁴. Notable examples include the substantial (>70 %) retreat of the
152 Larsen B ^{15,68}) and Wilkins ⁶⁹ ice shelves, and the effective collapse (>90 %) of the Prince Gustav
153 Channel ⁷⁰, Larsen-A ⁷¹, and Wordie ⁷² ice shelves. Since the 1950s, the combined area of Antarctic
154 ice shelves lost through retreat and collapse has been $33,917 \text{ km}^2$ ^{14,73}, or 22 % of their original
155 extent. Analysis of the geological record ^{56,74} has confirmed that the collapse events are unique
156 during the Holocene period.

157 The retreat and collapse of Antarctic Peninsula ice shelves has occurred in tandem with a rapid
158 regional atmospheric warming happening at several times the global trend ⁵⁷. These events have
159 been linked; warmer air temperatures lead to intensified surface melting ⁷⁵ which is believed to
160 cause hydraulic fracture of surface crevasses followed by ice shelf collapse ⁷⁶. Several Antarctic
161 Peninsula ice shelves have also thinned in the decades leading up to their collapse ^{11,13,59}, primarily
162 through ocean driven melting at their base. This thinning may contribute to instability by weakening
163 ice shelf lateral margins prior to fracture ⁷⁷, and by enhancing rates of iceberg calving ⁷⁸. The

164 relationship is, however, not universal; for example, although the Wilkins ice shelf collapsed in 2009,
165 it did not thin in the preceding five years ⁷⁹. Indeed, recent satellite altimetry ⁸⁰ shows that the
166 surface elevation of Larsen C increased in the preceding decade, in response to cooler and not
167 warmer summertime temperatures. And although the observational evidence suggests that stability
168 of Antarctic ice shelves depends on both their thickness, geographical location or setting, the extent
169 to which those that are partially or wholly intact will continue to resist collapse remains uncertain.

170 The collapse of ice shelves does not contribute directly to sea level rise, because they are afloat.
171 However, there is an indirect effect: observations show that the grounded tributaries to the Larsen A
172 ⁸¹ and B ^{15,68} ice shelves did speed in response to the removal of the floating ice, which is presumed
173 to have offered resistance. To date, ice shelf retreat and collapse has been restricted to those
174 situated at the Antarctic Peninsula, in relatively warm climates, and has not threatened those farther
175 South which fringe the East and West Antarctic Ice Sheets. The largest recorded reduction in ice
176 shelf area at the Antarctic Peninsula to date was the calving of the 11,095 km² A-68 tabular iceberg
177 (Fig. 3) from the Larsen-C in 2016 ⁷³. However, this berg represented just 7 % of the ice shelf area,
178 and was similar in size to one (A-20) that broke free in 1986 ⁸². It is, therefore, not without precedent
179 - even during the relatively short satellite era - and there is as yet no evidence that either breakaway
180 disturbed the remaining ice shelf, or was anything other than routine iceberg production.

181

182 [Buttressing of Grounded Ice](#)

183 The term “buttressing” is used to describe the resistive forces imparted to a grounded ice sheet by
184 its peripheral ice shelves. In its absence, rates of ice sheet discharge increase non-linearly with ice
185 thickness, making the grounding lines of marine-based ice sheets difficult to stabilise because their
186 bedrock tends to deepen inland ⁸³. Floating ice shelves can, however, exert drag as they flow over
187 and around seamounts (pinning points) or as a result of their lateral confinement, and the extent to
188 which this drag can mitigate unstable retreat has long been the subject of glaciological debate (e.g.

189 ⁵⁴). In recent years, the rapid response of Antarctic glaciers to the collapse and thinning of ice shelves
190 at their termini has led to a reassessment of their resistive properties. At the Antarctic Peninsula, for
191 example, glaciers flowing into the former Larsen-A, Larsen-B, and Wordie ice shelves have surged
192 ^{24,68,84} after their collapse ¹⁴, as too have ice streams flowing into the Amundsen and Bellingshausen
193 Seas ^{21,43} in the wake of ice shelf thinning ^{3,12,79} and grounding line retreat ^{85,86}. While the former
194 events have been attributed to the destabilising effect of increased melting at the surface of ice
195 shelves, following regional atmospheric warming ^{15,68}, the latter are now firmly linked to enhanced
196 ocean-driven melting at their base, due to the intrusion of warm circumpolar deep water into the
197 cavities beneath them ^{67,87}.

198 Although the reservoir of grounded ice at the Antarctic Peninsula is relatively modest ¹³,
199 destabilisation of Amundsen Sea sector glaciers is a matter of considerable concern, because the
200 pace of ice drawdown during the satellite era has been swift ⁸⁸, and because they contain enough ice
201 to raise global sea levels by more than a metre ⁴. Over the past two decades, for example, surface
202 lowering has spread inland across the drainage basins of the Pine Island and Thwaites Glaciers at
203 speeds of between 5 and 15 km yr⁻¹, and the majority of their catchments are now in a state of
204 dynamical imbalance (thinning due to accelerated flow). This rapid spreading is a consequence of
205 several connected processes ⁸⁹ (Box 1); ice shelf thinning leads to initial reductions in sidewall and
206 basal traction at glacier termini, which then causes increased strain rates (flow) within the glacier ice
207 upstream, followed by further grounding line retreat due to the associated ice thinning – especially
208 in marine-based sectors of the continent. Glacier speedup may also lead to further reductions in
209 sidewall traction through rifting and fracture, and grounding line retreat can expose more ice to the
210 ocean melting responsible for the initial imbalance.

211 Glaciers flowing into the Amundsen Sea sector of West Antarctica are particularly susceptible to
212 climate forcing, due to their geometrical configuration and the absence of any significant ice shelf
213 barrier ⁹⁰, and today the pace of ice sheet retreat along parts of this coastline dwarfs that during the

214 Holocene period. The region's ice shelves have thinned by 3 to 6 m/yr ^{11,12}, and its glacier grounding
215 lines have retreated by 10 to 35 km since 1992 ^{85,86} – 20 to 30 times the rate since the last glacial
216 maximum according to analysis of the marine geological record ⁹¹. In response to these
217 perturbations, the grounded glaciers inland have sped up ^{21,43} and thinned ^{32,37} at accelerating rates.
218 For example, since the early 1990s, rates of ice flow at the Pine Island Glacier terminus have
219 increased by ~1.5 km/yr ⁴³ and rates of ice thinning have risen to over 5 m/yr ⁸⁸, and the sector
220 overall contributed 4.5 mm to global sea level rise between 1992 and 2013 ³⁸.

221 The forcing for these events is now widely regarded to lie in the surrounding ocean, because ice
222 drawdown has originated at and evolved from the terminus of neighbouring but distinct ice flow
223 units ⁴², and because warm ⁶⁷ and warming ⁴⁴ water is present within the cavities beneath their
224 peripheral ice shelves. According to numerical simulations, the Pine Island and Thwaites Glaciers ^{92,93}
225 may contribute a further ~4 mm to global sea levels over the 21st century in response to continued
226 forcing, and it has been concluded ⁹⁴ that the region is now undergoing marine ice sheet instability,
227 with no geometrical obstacles to prevent irreversible decline. However, satellite observations have
228 revealed that retreat of the Pine Island Glacier halted around 2011 ^{95,96}, and that ice thinning inland
229 abated in the following years ⁸⁸. This suggests that the situation is more complicated than a
230 consideration of the glacier geometry alone, and may involve changes in the degree of ocean
231 forcing, as has occurred in the recent past ^{87,97}.

232 Sea ice

233 Antarctic sea ice forms as the Southern Ocean surface freezes, and interacts with the neighbouring
234 ice shelves and grounded ice in many ways (e.g. Box 1). Satellite observations (e.g. Fig. 1) have
235 allowed us to map its extent ^{8,98}, thickness ^{99,100} and drift ^{101,102}, providing insight into the role it plays
236 in climate ⁶ and how it impacts on ecosystems ¹⁰³. In winter, Antarctic sea ice extends from an inner
237 zone of consolidated pack ice surrounding the continent, to the marginal ice zone near to the

238 powerful Antarctic Circumpolar Current, where floes are less concentrated (Fig.1 and Fig. 4). In
239 summer, the sea ice pack retreats to isolated pockets fringing the continent. As it forms, Antarctic
240 sea ice produces high-salinity shelf water when brine is rejected, which then sinks to the seabed.
241 This water drives buoyant plumes within the cavities beneath floating ice shelves, which melt glacier
242 ice at the grounding line, before returning to the open ocean along their base. Antarctic sea ice is
243 also characterized by local polynyas – persistent gaps in the ice cover (e.g. Fig. 4) that are sustained
244 by upwelling warm water, winds, tides, and ocean currents. These polynyas are a source of bottom
245 water (dense water occupying depths typically below 4000 m), and provide a link between the ocean
246 and atmosphere that affects weather and wildlife ¹⁰³. Land-fast sea ice can also act to stabilise ice
247 shelves and glacier tongues ¹⁰⁴, and to suppress ¹⁰⁵ or – upon its breakup - enable ¹⁰⁶ iceberg calving.

248 [Sea ice extent and drift](#)

249 Fluctuations in the area of Antarctic sea ice have been routinely charted since the late 1970s using
250 passive microwave satellite imagery ⁸. Annually, its average extent ranges from 3.1×10^6 km² in
251 February to 18.5×10^6 km² in September ¹ (e.g. Fig. 1). In contrast to the Arctic, where the area of
252 sea ice has declined progressively ¹, there has been a small, positive increase (1.6 ± 0.4 % per decade
253 between 1979-2016) in the hemispheric sea ice extent of the Southern Ocean ⁹. This trend runs
254 counter to the projections of most climate models ¹⁰⁷, and has occurred alongside a slow warming
255 (0.02 °C per decade since the 1950s) of the Southern Ocean ¹⁰⁸. Despite the trend, there is some
256 evidence of longer-term decline; reanalysis of early satellite records ^{109,110} and historical whale catch
257 positions ¹¹¹ suggest there may have been more ice cover in the 1960s and early 1970s than there is
258 today. In recent years, however, extreme changes have occurred - the extent of Antarctic sea ice
259 reached record maxima in three successive winters (2012, 2013, and 2014; ¹¹²), followed by a record
260 summertime minimum in March 2017 ^{113,114}.

261 Changes in both atmospheric and oceanic forcing affect the extent of Antarctic sea ice ^{115,116}, and
262 although total hemispheric extent has shown little overall change, there have been considerable
263 regional variations ^{117,118}. While the Weddell Sea, Indian Ocean and Western Pacific Ocean have all
264 seen modest trends (1.7 ± 0.8 %, 1.7 ± 0.99 % and 1.8 ± 1.2 % decade⁻¹, respectively) in sea ice
265 extent during the satellite era (1979-2016), there have been more substantial trends (3.3 ± 0.9 %
266 and -2.9 ± 1.4 % decade⁻¹, respectively) in the Ross Sea and Amundsen and Bellingshausen Seas ⁹.
267 The periods during which the Western Ross and Bellingshausen Seas are ice free in summer have
268 also changed, decreasing and increasing by two and three months, respectively, between 1979 and
269 2011 ¹¹⁸. Seasonal and decadal trends in the Weddell and Ross Seas (positive) and the Amundsen-
270 Bellingshausen Sea and Western Pacific Oceans (negative) reflect the influence of atmospheric
271 forcing ¹¹⁹. Although these fluctuations in sea ice extent are strongly correlated with the dominant
272 modes of Southern Hemisphere climate variability ^{117,119}, other factors are involved ¹²⁰. A range of
273 mechanisms have been explored, including changes in oceanic variability ¹²¹, atmospheric circulation
274 ^{122,123}, stratospheric ozone depletion ¹²⁴, meridional wind forcing ¹⁰², and freshwater input from ice-
275 shelf melt ^{125,126}.

276 Understanding the role sea ice plays in the Antarctic climate system also requires a consideration of
277 its dynamics ¹²⁷. In the Southern Ocean, sea ice drifts northwards and diverges under the influence of
278 winds and ocean currents (Fig. 1), and the fraction of open water is higher than in the Arctic ¹²⁸.
279 Satellite observations have illuminated both local-scale ^{129,130} and hemisphere-wide ^{101,102,131}
280 Antarctic sea ice dynamics. Strong, circumpolar, westerly winds drive sea ice eastwards in the outer
281 zonal band, while a nearly-continuous westward circumpolar flow exists along the coastal boundary
282 ¹⁰¹. Persistent atmospheric lows centered at the boundaries of major ocean basins are the dominant
283 drivers of sea ice motion, and these sustain large-scale gyres in the Weddell and Ross Seas ¹³². The
284 speed of sea ice drift in the eastern Weddell and Ross Seas has increased, in contrast to the western
285 Weddell Sea where it has decreased ^{102,132}, though these signals are still small compared to the inter-

286 annual variability ¹⁰¹. The general northward trajectory of the Antarctic sea ice pack also impacts on
287 its age and thickness; rarely does it survive for more than two years, and the average thickness of
288 floes (typically in the range 0.6 to 1.2 m) and pressure ridges are smaller than in the Arctic Ocean ⁷.
289 Locally, katabatic winds, tides, and ocean currents sustain coastal polynyas through sea ice drift
290 around the continent (e.g. Fig. 4), providing a link between the sea ice pack and the ice sheet
291 through their initiation of plumes beneath floating ice shelves ¹³³.

292 Summary and Outlook

293 In just three decades, satellites have transformed our appreciation of the extent and pace of change
294 in the Antarctic cryosphere. Despite being remote, fluctuations in its ice cover have a global impact.
295 The continent holds Earth's primary freshwater reservoir ⁴ and, together with its surrounding ice
296 shelves ³ and sea ice ¹, blankets 6 % of the planet in ice during austral winter. Although persistent ice
297 shelves have fringed Antarctica for thousands of years ⁵⁶, there is now widespread evidence of
298 changes in their extent ¹⁴ and thickness ^{3,11}. Altogether, their volume has decreased by more than
299 300 km³ yr⁻¹ since 1994 ^{3,11}, notably due to collapse and calving at the Antarctic Peninsula and rapid
300 thinning of those in the Amundsen and Bellingshausen Seas. These events have triggered retreat ^{85,86}
301 and acceleration ^{21,43} of marine terminating glaciers and ice streams around the continent, leading to
302 the drawdown of ice from their inland catchments ^{39,42}. Since 1992, the grounded ice sheet has lost
303 1350 ± 1010 Gt of ice, causing a net 3.8 ± 2.8 mm contribution to global sea level rise ³⁶. The waxing
304 and waning of Antarctic sea ice influences the planetary albedo, oceanic circulation, marine
305 productivity, and ecosystems ^{6,103}. Although its extent has increased by 1.6 ± 0.4 % per decade since
306 1979 ⁹, there are significant regional variations ^{9,117,118}, and there is evidence from historical records
307 ¹⁰⁹⁻¹¹¹ of a longer-term decline. These discoveries, and many more, have transformed our
308 understanding of the state of Antarctic ice.

309 Even though considerable progress has been made during the satellite era, key questions remain
310 unanswered. For example, the detailed pattern of glacier change at the Antarctic Peninsula is not
311 well known, because the rugged terrain poses a challenge for traditional remote sensing methods.
312 Though modest, the mass balance of the East Antarctic Ice Sheet nevertheless remains uncertain,
313 because its detection is complicated by uncertainties in rates of snowfall and glacial isostatic
314 adjustment. And the evolution and impacts of abrupt subglacial lake drainage events is poorly
315 defined, because frequent measurements of ice elevation and flow changes are often lacking at the
316 local scale. But understanding the thickness of sea ice across the Southern Hemisphere and the
317 nature of ice shelf collapse and retreat are pressing concerns. While the range of parameters that
318 can now be measured on grounded ice and on ice shelves may be considered comprehensive,
319 available satellite observations are insufficient to fully understand the nature and evolution of the
320 processes that are driving contemporary imbalance. Although the satellite altimeter record has been
321 used to resolve Southern Ocean dynamics, determining sea ice thickness - a key measure of its
322 volume and longevity - from measurements of its freeboard (the portion protruding above the ocean
323 surface) are hampered by poor knowledge of snow loading and its impact on the satellite retrieval.

324 In the case of Antarctic sea ice, uncertainties in the degree of radar penetration into the snowpack
325 ^{134,135} has so far limited the use of the 25-year radar altimeter record for measuring its thickness.
326 Some advances have been made using laser altimetry ^{99,100}, which scatters from the surface of the
327 overlying snow, but continental-scale trends in Antarctic sea ice thickness and volume nevertheless
328 remain elusive due to the paucity of in situ measurements. One way to tackle this problem is to
329 exploit the relationship between the amount and roughness of snow on sea ice and its total
330 thickness ¹³⁶, an approach that may be realised with the launch of ICESat-2 which has a laser capable
331 of detecting surface roughness and thickness ^{137,138}. Another possibility is to combine freeboard
332 measurements retrieved from different scattering horizons to estimate the snow load directly, for
333 example using observations acquired by the CryoSat-2 Ku-band and ALtiKa Ka-band radar altimeters

334 ^{139,140}, and, in the future, ICESat-2. New techniques are also emerging to map the extent, type, age,
335 drift, and roughness of sea ice with fine resolution using synthetic aperture radar imagery.

336 Over land ice, the record of ice sheet motion data is too sparse to determine whether changes in
337 flow have occurred on sub-annual timescales across much of the continent. For example, the record
338 of ice sheet motion data is too sparse to determine whether changes in flow have occurred on sub-
339 annual timescales across much of the continent. On this point, the outlook is promising thanks to the
340 systematic acquisition plans of the Sentinel-1 synthetic aperture radar and Landsat-8 optical imager
341 missions. A key unanswered science question is how long it will take for the ice shelves that are
342 currently thinning to reach a point whereby they are no longer providing effective buttressing for
343 the grounded ice inland. To address this, observations are required with sufficient frequency to track
344 the events themselves which, in the case of ice shelf collapses, have taken place over months or
345 even days ¹⁴. Although it is possible to monitor grounding line migration with high precision using
346 synthetic aperture radar interferometry ^{85,86}, the revisit period of satellite missions is currently too
347 long for the technique to be effective over rapidly deforming ice, and so other methods - such as
348 repeat satellite altimetry ¹⁴¹ - will need to be exploited to track this precursor to ice sheet dynamical
349 imbalance.

350 The past decade has been a golden era for satellite glaciology, with a host of different sensors in
351 orbit, simultaneously. However, measuring ice loss from Antarctica at the continental scale is today
352 heavily reliant upon a single ageing mission - CryoSat-2 - which, at 8 years old, is now more than
353 double its planned lifetime, and the continuity of passive microwave observations of sea ice
354 concentration and extent remains uncertain. Given the societal importance of changes in ice cover
355 and global sea level, this situation carries considerable risk, and the next generation of satellite
356 sensors are eagerly awaited.

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364 Competing Interests

365 The authors declare no competing interests.

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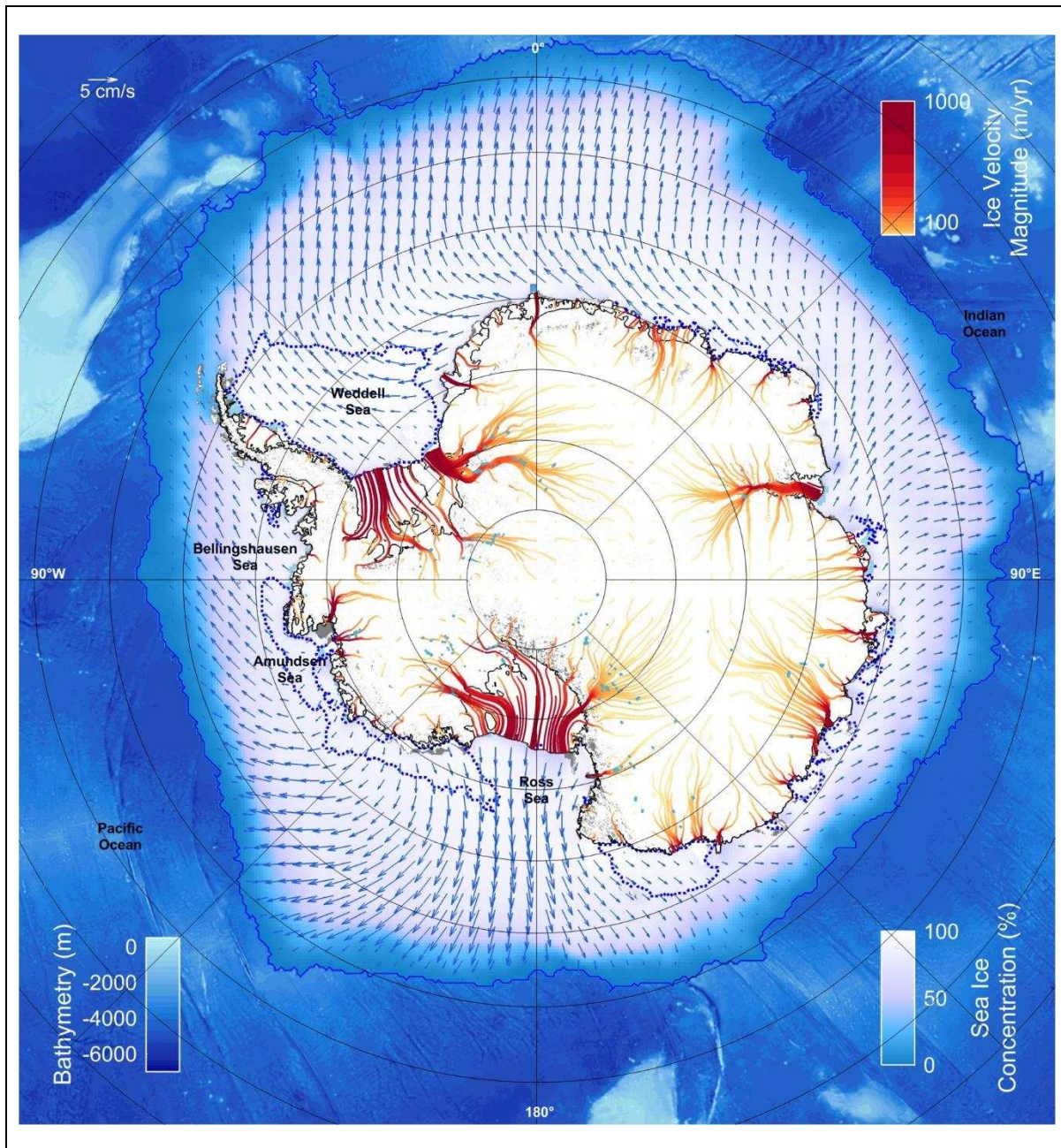


Figure 1. Average annual motion of the Antarctic ice sheet and ice shelves, and of the surrounding sea ice in winter. The ice sheet is drained by around 30 principle flow units and the sea ice transport is generally northwards, with gyres in the Ross and Weddell Seas. Grounded ice and ice

shelf motion are derived from multiple satellite interferometric synthetic-aperture radar data acquired between 2007 and 2009²⁰. Ice sheet motion flowlines are superimposed on the MODIS mosaic of Antarctica¹⁴². Sea ice motion is the mean of daily gridded Polar Pathfinder radiometry obtained during peak winter (September) of each year in the period 1990 to 2016¹⁴³. Sea ice motion vectors are superimposed on a map of mean sea ice concentration derived from passive microwave brightness temperatures in September between 1990 to 2016¹⁴⁴. Also shown (blue dashed boundaries) are the average minimum extent of sea ice recorded between 1990 to 2016¹⁴⁴, the grounded ice sheet and the floating ice shelves (black boundaries), and the bathymetry of the surrounding ocean¹⁴⁵. Active subglacial lakes (light blue) were mapped using satellite radar and laser altimetry⁵¹.

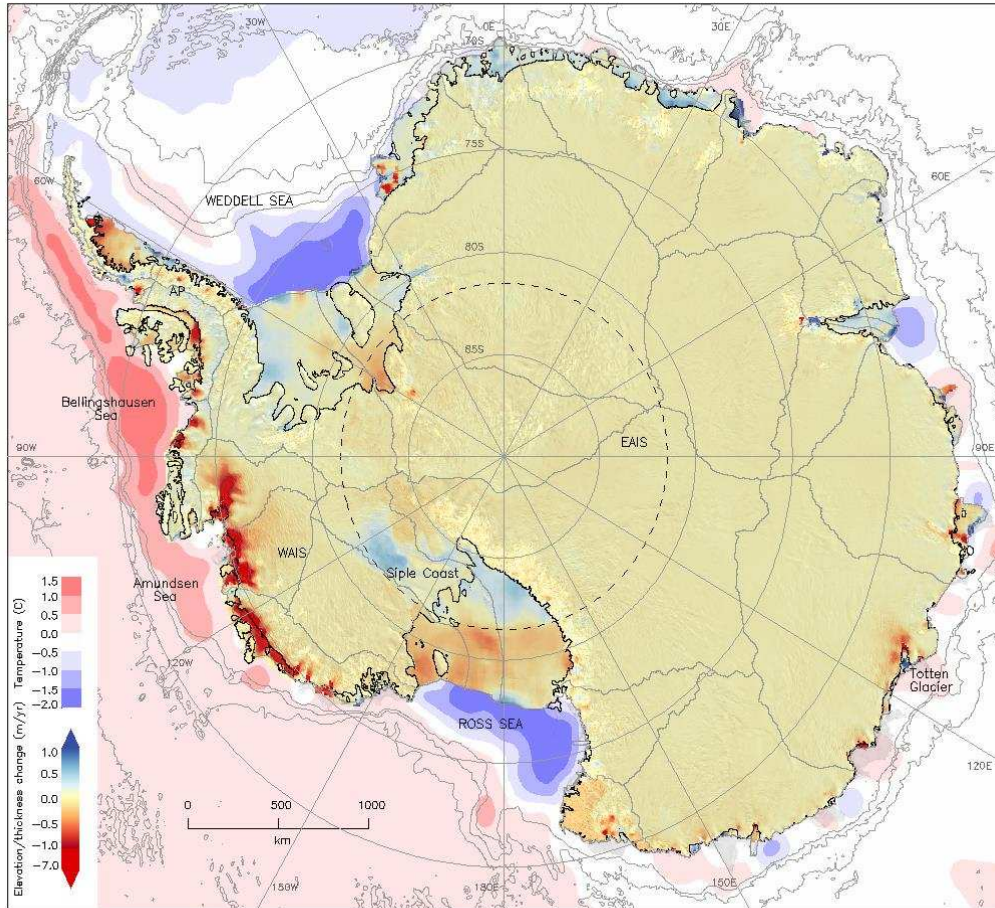


Figure 2. Average trend in the elevation and thickness of Antarctic grounded ice and ice shelves, respectively, determined between 1992 and 2017 North of 81.5°S (dashed grey circle), and between 2010 and 2017 elsewhere. Also shown is the depth of ¹⁴⁶ and estimated ocean temperature at ¹⁴⁷ the sea floor around the continent. Changes in grounded ice and ice shelf thickness were estimated using repeat satellite altimetry following the methods of ³⁶, ¹⁴⁸, and ³. Thickness trends are superimposed on an optical image mosaic of the floating and grounded ice ¹⁴², and is divided (grey lines) into the principle ice drainage catchments ². Since 1992, the grounded ice sheet and its peripheral ice shelves have thinned in locations adjacent to warm ocean currents. Although the East Antarctic ice sheet is mostly stable, there have been marked

changes in West Antarctica, including accelerated thinning of glaciers draining the Amundsen Sea sector and constant thickening in southerly catchments of the Siple Coast. While the former is a response to ocean-driven melting of ice shelves at glacier termini ¹², the latter is associated with stagnation of ice flow due to a loss of basal lubrication ¹⁴⁹. At the Antarctic Peninsula, ice shelf collapse (Cook and Vaughan, 2010) has triggered inland glacier acceleration ^{25,68} and thinning ¹⁵.

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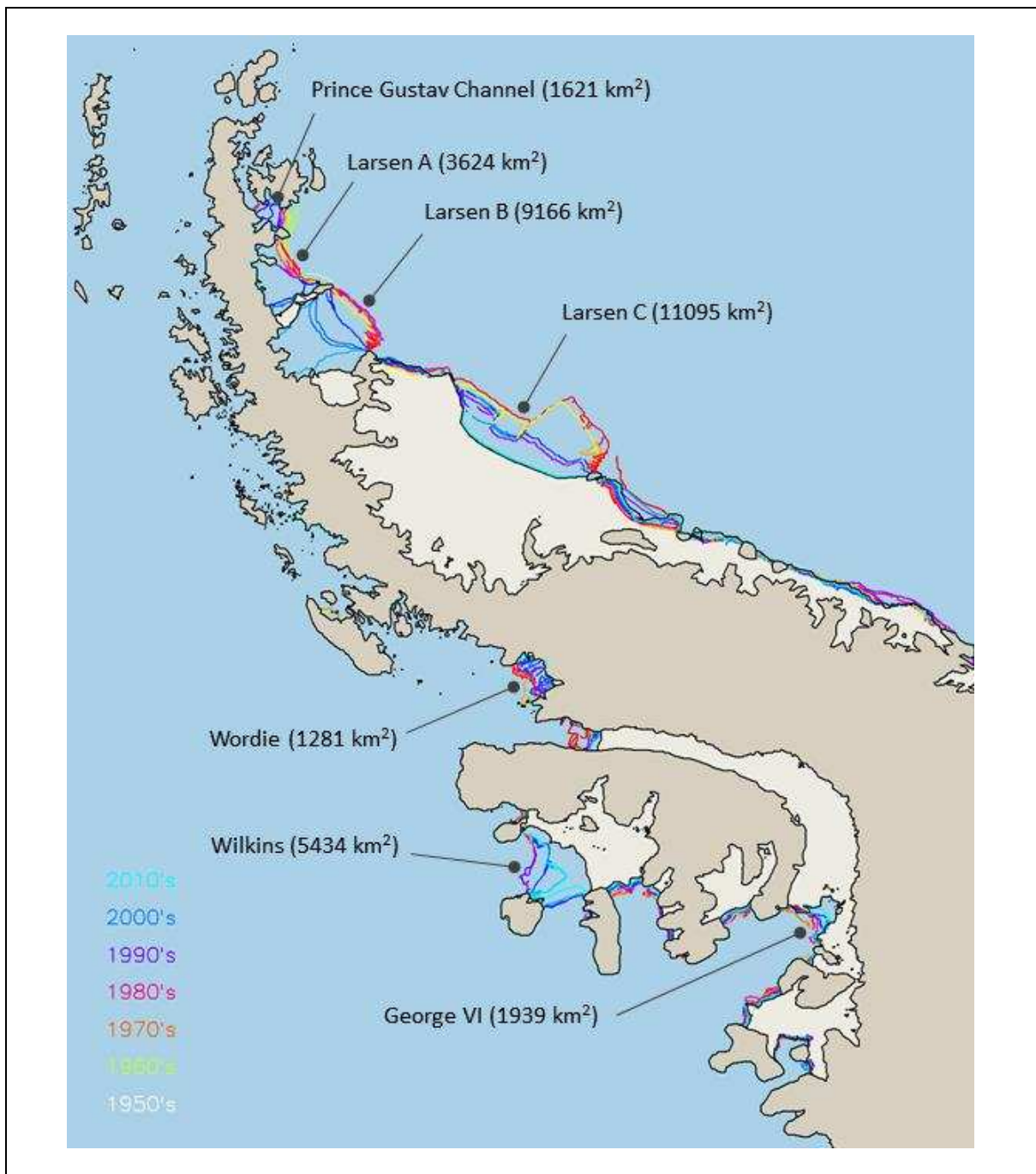


Figure 3. Temporal changes in the location of ice shelf barriers at the Antarctic Peninsula (coloured lines) as determined from satellite imagery since the 1950s, and their net reduction in area over the same period (numbers in brackets) ^{14,73}. The reduction in area at Larsen-C includes the recent calving of the A-68 tabular iceberg. In the 1950s, the total area of Antarctic Peninsula ice shelves was estimated to be 152,246 km². Since then, an area of 33,917 km² has been lost during episodic calving events.

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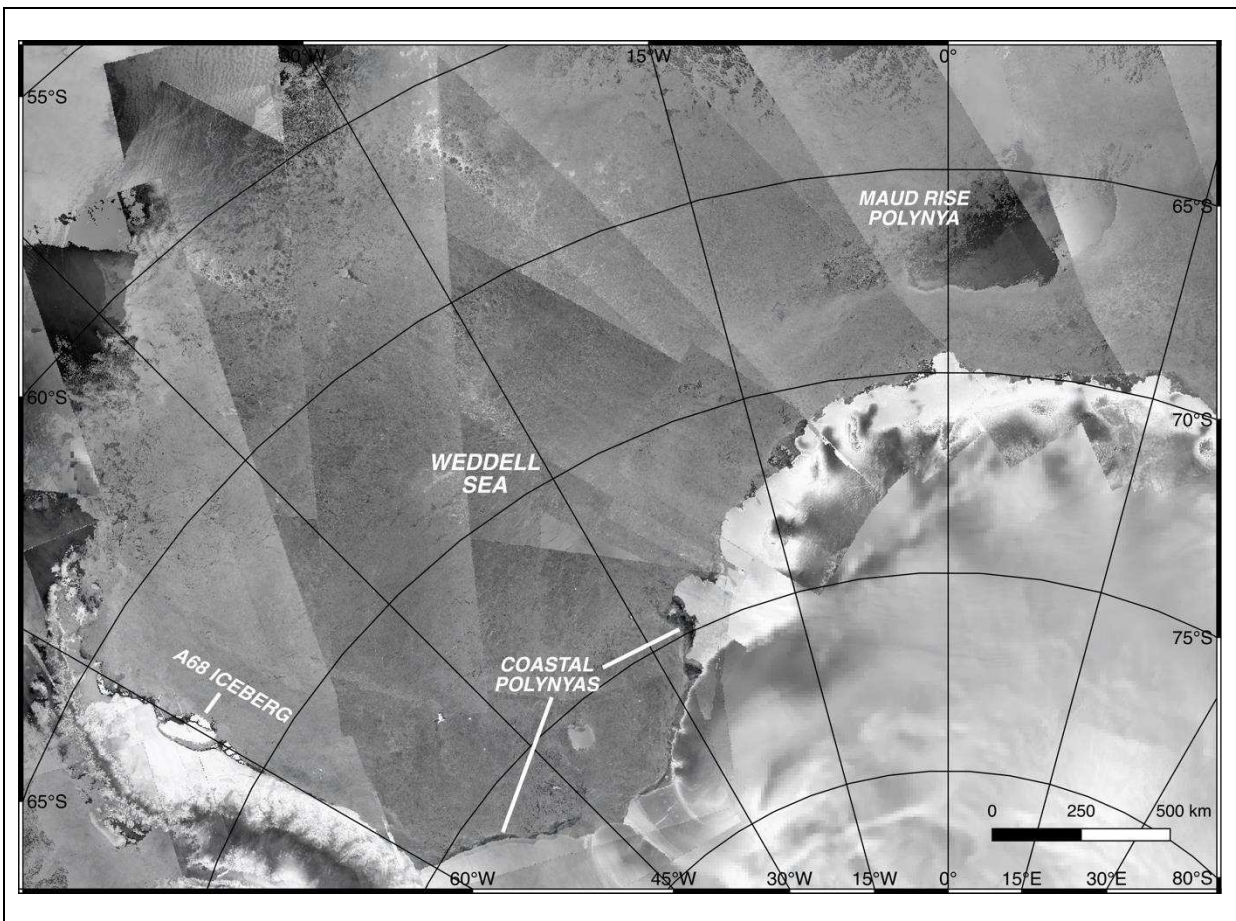


Figure 4. Sea ice in the Weddell Sea in early November 2017, based on a composite of Sentinel-1

synthetic aperture radar imagery, MODIS optical satellite imagery, and ASCAT scatterometer data. The satellite data were acquired during Austral winter when the ice cover is close to its maximum extent, stretching beyond the tip of the Antarctic Peninsula into Drake Passage. The composite reveals details of the diffuse ice cover in the marginal ice zone along the boundary with the open ocean, and the more compact, consolidated ice cover farther south. A large, $\sim 35,000 \text{ km}^2$ ice-free area is visible near Maud Rise (66S, 5E); this polynya is formed by thermally-driven convection in the water column, due to the interaction of ocean currents and sea floor topography. Coastal polynyas are visible along the edges of the Filchner-Ronne, Brunt and Stancombe-Willis Ice Shelves, where openings in the ice cover are driven by katabatic wind, tides, and ocean currents. New sea ice, which rejects brine during formation, is continually produced in these polynyas, making polynyas a source of dense, high salinity deep water which plays an important role in the global thermohaline circulation. Also visible is the large, tabular A68 iceberg (located $\sim 68\text{S}$, 60.5W) which calved from Larsen C Ice Shelf on 12 July 2017, and is now adrift in the western Weddell Sea.

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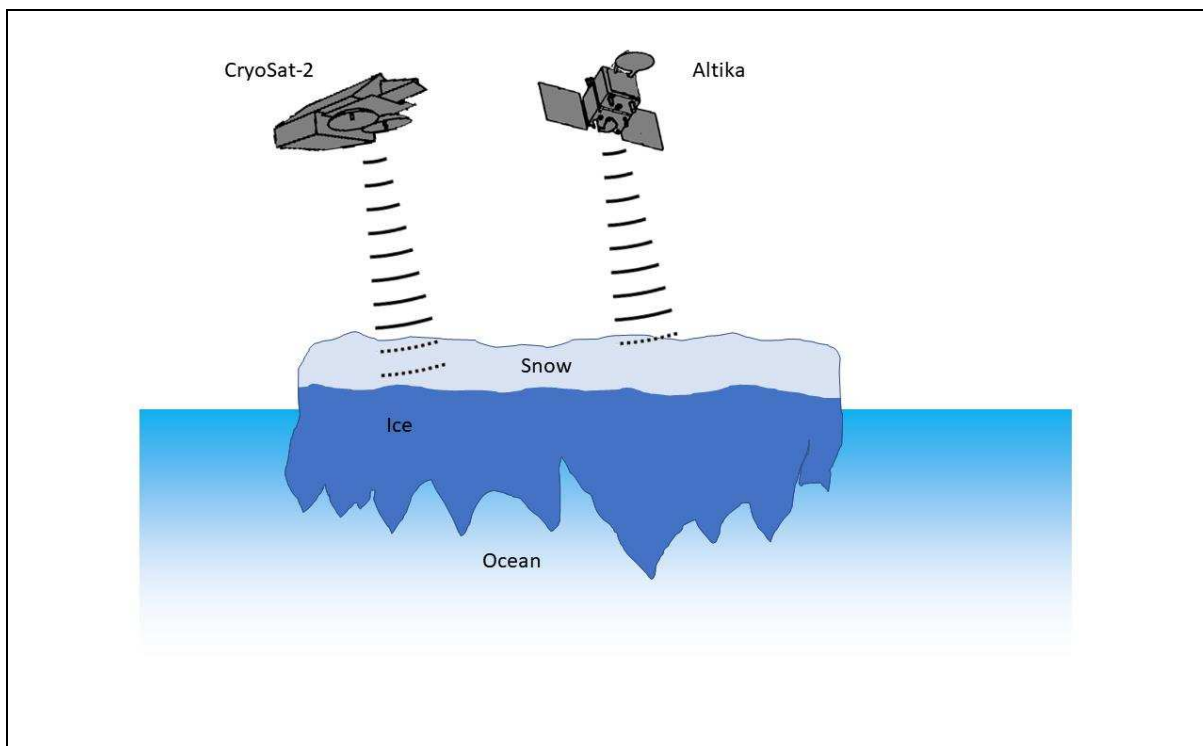
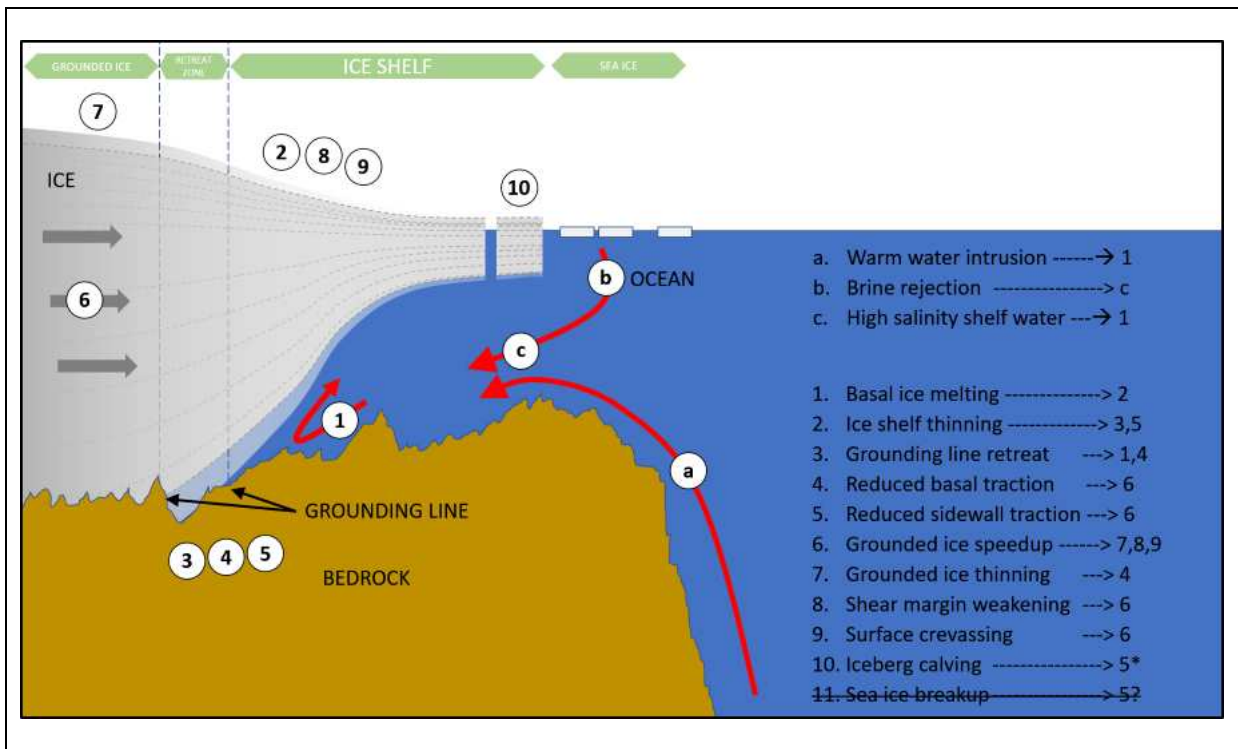


Figure 5. Schematic of a sea ice floe as observed by CryoSat-2 and AltiKa. Floe thicknesses are typically derived from measurements of their freeboard (the portion protruding above the ocean surface), an estimate of the snow loading, and the principal of buoyancy⁹⁹. CryoSat-2 and AltiKa operate at different radar frequencies, and their echoes scatter from locations near to the lower and upper boundaries of the snow layer, respectively^{139,140}. Measurements acquired at both frequencies could provide a direct measurement of the snow loading, improving the certainty of sea ice floe thickness estimates.

379 Box 1



Box 1. Ice shelf buttressing processes (points 1-10) associated with external forcing (points a-c), and their connectivity. Ice shelves are floating sheets of ice that form as glaciers spread out into the ocean, typically within confined embayments. They are permanently attached to the grounded ice sheet resting on land, and accumulate snow at their surface and, occasionally, frozen ocean water at their base. If warm water enters the ocean cavity beneath an ice shelf (a) it can drive increased basal

ice melting (1) and ice shelf thinning (2), which in turn leads to retreat of the grounding line (3) – the junction between grounded and floating ice on the seafloor. Ice shelf thinning reduces sidewall (lateral) traction (4) and grounding line retreat reduces basal traction (5). Both processes lead to speedup of the grounded ice (6), which causes grounded ice thinning (7). Glacier speedup can also lead to weakening of lateral shear margins (8) and increased crevassing (9). Iceberg calving (10) can also lead to reduced sidewall traction. Sea ice (frozen sea water) can play a role, through brine rejection (b) which drives the production of warm high salinity shelf water (c), and by effectively reducing traction on ice-shelf breakup (1).

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