

1 Basal melting of Ross Ice Shelf from solar heat  
2 absorption in an ice-front polynya

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5 **Abstract**

6 Ice-ocean interactions at the base of Antarctic ice shelves are rarely observed, yet have  
7 a profound influence on ice sheet evolution and stability. Ice sheet models are highly sen-  
8 sitive to assumed ice shelf basal melt rates; however, there are few direct observations of  
9 basal melting or the oceanographic processes that drive this, and consequently our under-  
10 standing of these interactions remains limited. Here we use new in-situ observations from  
11 the Ross Ice Shelf to examine the oceanographic processes that drive basal ablation of the  
12 world's largest ice shelf. We show that basal melt rates beneath a thin and structurally im-  
13 portant part of the shelf are an order of magnitude higher than the shelf-wide average. This  
14 melting is strongly influenced by a seasonal inflow of solar-heated surface water from the  
15 adjacent Ross Sea Polynya that downwells into the ice shelf cavity, nearly tripling basal  
16 melt rates during summer. Melting driven by this frequently overlooked process is ex-  
17 pected to increase with predicted surface warming. We infer that solar heat absorbed in ice  
18 front polynyas can make an important contribution to the present-day mass balance of ice  
19 shelves, and potentially impact their future stability.

## 20 **Main**

21 The ice shelves that fringe Antarctica interact with the Southern Ocean across a basal surface of  
22  $1.56 \times 10^6 \text{ km}^2$  [1]. Melting of this vast surface is the single largest cause of mass loss from the  
23 Antarctic Ice Sheet [1, 2]. Thinning induced by ice shelf basal melting can also modify inland  
24 ice flow, reducing the stabilising effect of sills, shoals and sidewalls [3,4], in some cases driving  
25 instantaneous dynamic responses as far as 900 km inland [4]. Although these processes provide  
26 a primary control on the future evolution of the ice sheet [3, 5], there are still relatively few  
27 direct observations of basal melting and oceanographic conditions within ice shelf cavities [6],  
28 and this paucity of data impedes the development of theory and models.

29 In Antarctic shelf seas, three main water masses are thought to influence ice shelves [7]:  
30 Circumpolar Deep Water (CDW), a relatively warm water mass that surrounds Antarctica at  
31 intermediate depth; high and low salinity Shelf Water (HSSW and LSSW), which is formed as  
32 the sea surface freezes during winter; and Antarctic Surface Water (AASW), a relatively fresh  
33 and buoyant water mass influenced by solar heating and sea ice melting during summer [8].

34 These water masses have contrasting impacts on ice shelves. CDW in the Amundsen Sea  
35 has caused ice shelves in the region to thin over recent decades [9, 10] driving mass loss from  
36 the interior ice sheet [11, 12]. In contrast, the vast Ross and Filchner-Ronne ice shelves appear  
37 to be near equilibrium [13, 14], due to the presence of cold Shelf Waters that limit their exposure  
38 to CDW [8, 15, 16]. The influence of AASW on ice shelves is less clear and seldom considered.  
39 Although buoyant, AASW can enter ice shelf cavities due to wind [17, 18] and tidal forcing  
40 [19–21]; however, observations of AASW beneath ice shelves have only been made recently  
41 [22, 23], and few studies have examined this process in detail.

42 For the Ross Ice Shelf (RIS), which at  $500\,809 \text{ km}^2$  [1], accounts for 32 % of Antarctica's  
43 total ice shelf area, recent satellite observations suggest relatively low shelf-wide mean basal

44 melt rates of  $0.07$  to  $0.11 \text{ m yr}^{-1}$  [1, 2, 24]. However, these studies also indicate rates above  
45  $1 \text{ m yr}^{-1}$  in the north-western sector of the shelf [1,24]. Although remote sensing estimates have  
46 uncertainties of over 100 % [24], earlier glaciological observations [25–27] and oceanographic  
47 models [23, 28–31] also indicate rapid melting in the north-western RIS. These models suggest  
48 that active circulation of frontal water into the cavity during summer and low-frequency flow  
49 variability may influence this region. Isolated observations from beneath the ice shelf support  
50 this picture [23, 31]; however, the details of these processes and the magnitude of their impact  
51 on the ice shelf remain unclear.

52 Here we present new in-situ observations of basal melting and sub-ice shelf oceanographic  
53 conditions from the north-western RIS. The aims of the study are two-fold; to quantify and map  
54 basal ablation in the region surrounding Ross Island, and to examine the role of surface water  
55 in driving this process.

## 56 **Radar mapping of basal melting**

57 To quantify basal melting of the north-western RIS, we used a downward-looking phase-sensitive  
58 radio echo sounder [32, 33] to make precise measurements at 78 sites surrounding Ross Island  
59 (Fig. 1). All sites were resurveyed after one year, allowing annual-mean ablation rates to be  
60 calculated (see Methods and Supplementary Table 1). Further observations were used to deter-  
61 mine short-term summer melt rates near the ice front (Supplementary Table 2). To map the melt  
62 rate field and estimate the total basal mass loss from the region, the melt rate observations were  
63 interpolated onto a regular grid (see Methods).

64 The observations show intense basal melting within 1 km of the ice shelf front, with annual-  
65 mean rates of  $2.4$  to  $7.7 \text{ m yr}^{-1}$  in this zone (Fig. 1). Melt rates reduce exponentially with  
66 distance from the ice front, typically halving within the frontal 3 km (Fig. 1b). This pattern is  
67 consistent with trends inferred from laser altimetry [34]; however, our observations show higher

68 melt rates, and reveal small-scale spatial variability.

69 The melt rate observations indicate strong seasonal melt rate variability, with rapid summer  
70 melting especially pronounced near the ice front and on Transec TC (Fig. 1c). Here, a max-  
71 imum ablation of  $0.714 \pm 0.007$  m was recorded over a 4.89 day period during January 2013,  
72 indicating a melt rate of  $0.146 \text{ m d}^{-1}$  ( $53 \text{ m yr}^{-1}$ )(Supplementary Table 2). This rate is a factor  
73 of 6.9 higher than the annual-mean rate at the same site, suggesting that a large component of  
74 the net ablation occurs in summer.

75 Beyond the frontal 15 km, ablation rates are lower, yet still considerably above the shelf  
76 wide average of  $\sim 0.1 \text{ m yr}^{-1}$  [1, 2, 24]. The pattern of melting implies that net heat flux into the  
77 cavity is strongest  $\sim 20$  km east of Ross Island, near an embayment where frontal ice thickness  
78 is just  $\sim 100$  m [35]. In this area, rapid melting extends further from the ice front (Fig. 1a). The  
79 mean basal melt rate across the  $7782 \text{ km}^2$  interpolated area is  $1.34 \text{ m yr}^{-1}$ , indicating a basal  
80 mass loss of  $9.5 \text{ Gt yr}^{-1}$ . This represents 20 % of the published net basal mass loss from the  
81 entire RIS [1] from 1.6 % of its area.

82 The rapid frontal melting and the large seasonal variations in melt rates suggest that melting  
83 near the ice front may be influenced by an inflow of warm surface water during summer, as  
84 observed beneath McMurdo Ice Shelf [23]. As warm inflowing water is expected to progres-  
85 sively cool through contact with the ice base, this mechanism could also explain the exponential  
86 reduction in melt rate with frontal distance.

## 87 **Oceanographic observations**

88 To examine the oceanographic processes that drive enhanced melting in the north-western sector  
89 of RIS, a sub-ice shelf mooring was deployed 7 km from the ice front on Transect TB (Fig. 1a).  
90 Moored instruments recorded currents, temperature and salinity hourly for up to 4 years. In  
91 addition, an Upward Looking Sonar (ULS) deployed beneath the ice base allowed basal melt

92 rates to be determined (see Methods).

93        Currents at the mooring site show strong seasonal variability, with mean depth-averaged  
94 outflows during winter (April-November) and inflows during late summer (February-March)  
95 (Fig 2a). The strength and duration of this inflow imply ventilation of the outer  $\sim 50$  to 160 km  
96 of the cavity. Comparison of water temperature at the mooring with sea surface temperature  
97 (SST) [36] north of the ice front suggests the inflow has a direct impact on temperatures within  
98 the cavity (Fig. 2c). Averaged over the region within 100 km of the mooring, SST follows an  
99 annual cycle closely linked to mean sea ice concentration in the same region (Fig. 2c). Here  
100 temperatures vary from the surface freezing point during winter, to seasonal maxima of over  
101  $1^\circ\text{C}$  in January when sea ice is absent. Temperatures measured  $\sim 13$  m below the ice base at  
102 the mooring show a similar but delayed and attenuated cycle; with seasonal maxima occurring  
103 in February,  $\sim 1$  month after the SST peak.

104        Further clues to the origin of the warm inflow are provided by a single serendipitous Con-  
105 ductivity Temperature Depth (CTD) cast, sampled 120 km north of the mooring site (Fig. 2a)  
106 one day prior to the mooring deployment [37]. This cast shows a 40 m thick, relatively fresh  
107 upper layer with a surface temperature of  $0.178^\circ\text{C}$  (Fig. 2b). The  $0.8^\circ\text{C}$  temperature range of  
108 the upper 35 m is associated with a salinity range of just 0.009 psu, suggesting solar heating of  
109 a previously homogenous layer. Comparison of the CTD cast with Temperature-Salinity (TS)  
110 observations from the upper moored sensor, located 8 m below the ice base ( $\sim 229$  m below sea  
111 level), suggests that surface water is drawn into the cavity. By extending the extrema of the  
112 offshore CTD cast with  $2.4^\circ\text{C psu}^{-1}$  melt-water mixing lines [16, 38], we define an envelope  
113 of water masses that could be formed from the offshore water by interaction with the ice shelf.  
114 Throughout January and February 2011, all 863 observations from the upper sub-ice shelf TS  
115 sensor fall within this envelope (Fig. 2b). The sub-ice shelf observations are consistent with  
116 a source region in the offshore profile above 55 m, indicating that the surface layer downwells

117 to reach the mooring site. The characteristics of the warmest sub-ice shelf water are consistent  
118 with the mean properties of the upper  $\sim 50$  m of the offshore profile, suggesting that the surface  
119 layer is homogenized before reaching the mooring.

120 The impact of water temperature variability on the ice shelf is illustrated by comparing the  
121 ULS basal melt rate record with water temperature from the upper moored current meter (Fig.  
122 2d). Both records are low-pass filtered (cutoff frequency  $f_c = 0.02 \text{ cycle day}^{-1}$ ) to reduce the  
123 impact of noise in the range observations (see Methods). The dominant feature of the melt rate  
124 record is a strong seasonal cycle that peaks in February. The smoothed temperature and melt  
125 rate records are highly correlated (Pearson's  $r = 0.78$ ), indicating that water temperature is the  
126 dominant driver of low-frequency melt rate variability at the mooring site.

127 Considering the central 3-year period June 2011 - May 2014, the mean ULS melt rate  
128 is  $1.8 \text{ m yr}^{-1}$  (Fig. 1b). Seasonally, melt rates vary between  $1.1 \text{ m yr}^{-1}$  during late winter  
129 (September - November) when sub-ice shelf water temperatures are at or below the surface  
130 freezing point, and  $3.0 \text{ m yr}^{-1}$  during summer (January - March) when surface water is seen at  
131 the site. This rapid melting occurs over a relatively short period, and the difference between  
132 the late-winter and annual-mean rates implies that summer melting accounts for  $\sim 0.7 \text{ m yr}^{-1}$ ,  
133 or  $\sim 40\%$  of net ablation at the mooring site. Determining the contribution of surface water to  
134 net ablation over the wider survey region is more difficult due to the lack of winter melt rate  
135 observations away from the mooring site. However, the higher summer/annual-mean melt rate  
136 ratios seen on Transect TC, 20 km west of the mooring (Fig 1b,c and Supplementary Table 2)  
137 suggest that the influence of surface water is stronger here.

138 Although melt rates peak during February, winter rates are still an order of magnitude  
139 higher than the satellite-inferred shelf-wide average, and contribute significantly to the high  
140 average melt rates. This indicates that winter current speeds or water temperature at the moor-  
141 ing are higher than the shelf-wide average. Whilst localised flow enhancement may contribute

142 to rapid melting in the region, it appears unlikely the flow variability could explain the order-of-  
143 magnitude melt rate enhancement, and we suggest that temperature variability plays the domi-  
144 nant role. While there are few observations from elsewhere in the cavity, CTD profiles from the  
145 central RIS show a thick Ice Shelf Water boundary layer within  $0.03\text{ }^{\circ}\text{C}$  of the in-situ freezing  
146 point [39]. In contrast, at the mooring site water near the ice base is often above the surface  
147 freezing point during early winter indicating remnant heat from the summer inflow. Even dur-  
148 ing late winter, HSSW some  $0.17\text{ }^{\circ}\text{C}$  above the in-situ freezing point is frequently observed, and  
149 this suggests active cross-frontal flow that ventilates the cavity [40, 41].

## 150 **Surface ocean heat**

151 The identification of a warm surface water inflow that drives rapid basal melting raises crucial  
152 questions; what is origin of this heat, and could this process influence other ice shelves? To  
153 address these questions, we examine summer SST and sea ice concentration observations from  
154 coastal Antarctica.

155 Figure 3a shows long-term mean January SST [36] and sea ice distribution, represented here  
156 by the mean 15 % sea ice concentration contour [42] (see Methods). At the largest scale, sum-  
157 mer SST is inversely correlated with sea ice concentration and the coldest waters are typically  
158 found near the coastline (Fig. 3a); however, higher temperatures are observed wherever signif-  
159 icant open water exists, including coastal polynyas near the Ross and Amery ice shelves (see  
160 also [43]).

161 Within the Ross Sea, SST variability is dominated by a warm surface anomaly, previously  
162 identified in CTD observations [44], that closely matches the position of the Ross Sea Polynya  
163 (Fig. 3b). In this region, January-mean SST reaches  $\sim 0.5\text{ }^{\circ}\text{C}$ . This pattern of warming is con-  
164 sistent with atmospheric modelling that indicates Antarctic polynyas absorb solar heat rapidly  
165 during summer [45], and has previously been attributed to summer insolation in the Ross Sea

166 Polynya [23, 46].

167 To assess whether the warm surface pool evident in Fig. 3b could supply the energy re-  
168 quired for elevated melting in the survey region, we calculated the available thermal energy  
169 within its surface waters during January (see Methods). Considering the region within the 0 °C  
170 SST isotherm (Fig. 3b), and assuming a surface mixed layer depth of 10 m implies a sensible  
171 heat content of  $8.3 \times 10^{18}$  J, sufficient to melt 22 Gt of ice shelf. This is approximately twice  
172 the observed ablation within the survey region. Despite significant uncertainty in the mixed  
173 layer depth, surface waters in the Ross Sea clearly represent a glaciologically significant heat  
174 reservoir during summer.

175 Beyond the Ross Sea, coastal SSTs above  $-0.5$  °C are only seen in the north-western  
176 Antarctic Peninsula where sea ice concentration is low, and in the polynya adjacent to the east-  
177 ern Amery Ice Shelf (Fig. 3a). Consequently, while surface layer heat may affect these regions,  
178 this process does not appear to be widespread at present.

## 179 **Drivers and impacts of surface water inflow**

180 Although surface waters have been considered a potential driver of ice shelf basal melting for  
181 some time [7, 22–24], the observations presented here provide the most detailed evidence of this  
182 process to date. These suggest that solar heated surface water contributes significantly to the  
183 basal mass balance of RIS, and that surface water plays a larger role in the mass balance of ice  
184 shelves than previously assumed.

185 In the north-western Ross Sea, the impact of surface water can be attributed to two pro-  
186 cesses; localised solar heating of the surface ocean during summer, and transport of this energy  
187 into the cavity by a seasonal inflow. Surface heating appears closely linked to the consistent  
188 wind-driven expansion of the Ross Sea Polynya during spring [23, 46]. During this period,  
189 sustained southerly winds, guided by the Transantarctic Mountains, preferentially export sea



190 ice from the western ice front [47,48]. As air temperatures and insolation increase throughout  
191 November and December, the polynya expands rapidly (Fig. 2c and Fig. 3b), as illustrated by  
192 the sea ice distribution during this period (Fig. 3c). This process increases solar energy ab-  
193 sorption in the surface layer, and removes the latent-heat sink presented by sea ice, aiding rapid  
194 heating of the surface layer (Fig. 2c).

195 The drivers of the late-summer inflow are less obvious; however, due to the buoyancy of the  
196 surface layer, it appears likely that external forcing is required. In contrast to the wind-driven  
197 downwelling observed elsewhere [18], the inflow observed here is not associated with down-  
198 welling favourable winds. Modelling suggests shelf and sub-ice shelf circulation in the region  
199 is strongly influenced by density gradients caused by seasonal brine release in the polynya [41]  
200 and that these influence seasonal flow variability near Ross Island [28]. Considering these fac-  
201 tors, we conclude that the elevated melt rates in the north-western RIS are linked to the location  
202 of Ross Sea Polynya, and ultimately the mean winds and orography of the region.

203 The identification of surface layer heat as a driver of basal melting on the RIS has several  
204 important implications. Firstly, as heat absorption within the polynya is controlled by atmo-  
205 spheric processes [45], basal mass balance within the frontal zone of the ice shelf is likely to  
206 vary with atmospheric and surface ocean conditions near the ice front on seasonal, inter-annual,  
207 and longer time scales. Considering that summer sea ice concentrations in the Ross Sea are  
208 projected to decrease by 56% by 2050 [49], and the ice free period is also expected to in-  
209 crease [50], it appears likely that ice shelf basal melting within this region will also increase  
210 rapidly. If surface warming and sea ice loss is widespread, this process may also become more  
211 widespread. Secondly, AASW drives a mode of basal ablation that is distinct from that of  
212 denser water masses, and these differences have implications for ice shelf stability. For exam-  
213 ple, whereas meltwater derived from HSSW can re-freeze in shallower regions potentially sta-  
214 bilising ice shelves [51]; due to its relative warmth, meltwater formed from AASW is unlikely

215 to be redeposited. Furthermore, the influence of surface water is greatest in frontal regions.  
216 Although some frontal regions are unimportant to the stability of ice shelves, others contain  
217 critical pinning points that sustain the frontal location [52, 53]. Ross Island appears to be one  
218 such pinning point, and recent modelling shows that the rapid melting identified here influences  
219 a structurally critical region where ice thickness changes can influence the flow speed of the  
220 entire ice shelf [4].

221 The exposure of this sensitive part of the ice shelf to surface ocean heat implies that ground-  
222 ing line flux of the entire ice shelf may be modulated at seasonal to inter-annual scales by the  
223 surface water inflow. This process represents a frequently overlooked but potentially important  
224 factor in regional ice shelf mass balance and should be considered in future assessments of ice  
225 shelf stability.

## 226 **Methods**

### 227 **Radar observations**

228 Basal melt rates were measured using the British Antarctic Survey's Autonomous Phase-sensitive  
229 Radio Echo Sounder (ApRES) [33]. This Frequency Modulated Continuous Wave (FMCW)  
230 radar has a center frequency of 300 MHz and bandwidth of 200 MHz. The instrument uses  
231 Direct Digital Synthesis to generate the linear 1 s chirp, and demodulates the radio frequency  
232 carrier wave by mixing the receive (Rx) signal with an attenuated feed of the transmit (Tx) sig-  
233 nal. The resulting audio frequency signal is digitized at 40 kHz. The same high-stability master  
234 clock is used to drive both signal generation and sampling, ensuring precise synchronization.  
235 Technical details of the instrument are given in [32].

236 The instrument was used in a pseudo-monostatic configuration, with Tx-Rx antenna sepa-  
237 ration of 3.44 m. At each site an ensemble (burst) of typically 100 chirps were recorded. Radar  
238 sites were marked with surface stakes to ensure the same column of ice was sampled on each  
239 visit. Care was taken to ensure precise relocation of the instrument relative to the marks, and  
240 repositioning error was estimated as  $<0.05$  m.

241 Data from each visit to each site were pre-processed as follows. For each burst, noisy chirps  
242 were removed, and the remaining chirps averaged. Each burst-mean chirp was weighted with  
243 a Blackman window, then extended to a multiple of 8 times its original length by appending  
244 trailing zeros. Following this zero-padding, the signal was circularly rotated so that the first  
245 sample of the modified signal was that of the centre of the unpadded chirp (see ref. [54]). Each  
246 chirp was then Fast Fourier Transformed.

247 Due to the frequency-range proportionality in linear FMCW radar observations [32], the  
248 amplitude of the resulting spectrum is analogous to a time-series of echo amplitude recorded by  
249 a time-domain radar. These complex valued spectra are hereafter referred to profiles. Without

250 zero-padding the profiles have a frequency resolution of 1 Hz, and the 200 MHz bandwidth  
251 implies a temporal resolution of 5 ns, corresponding to a range resolution of 0.43 m in ice. With  
252 zero-padding the interpolated range resolution is 0.0537 m. Here and in the following analysis  
253 we assume a nominal propagation velocity of  $1.68 \times 10^8 \text{ m s}^{-1}$  [55].

254 Although changes in the range of the basal reflector are relatively simple to determine,  
255 these are influenced by many factors including changes in radar hardware (e.g. cable length),  
256 compaction of the upper snow layers (firn), and strain within the solid ice and basal melting.  
257 To isolate the component due to basal ablation, observations of internal reflector displacement  
258 were used to tune a displacement model. This model was used to estimate the displacement  
259 of the ice base expected in the absence of basal melting. Net basal ablation is then determined  
260 from the difference between the expected and observed basal displacement. The approach used  
261 here is described in more detail in [54].

262 Vertical displacements between profiles observed at the same site were quantified by cross-  
263 correlating overlapping 15 m segments of the profiles. This provided vertical displacement esti-  
264 mates at 7.5 m resolution throughout the ice shelf. First, the integer range-bin (or coarse) offset  
265 between the two profiles was determined by cross-correlation of the profile amplitude. Fine-  
266 scale offsets were determined from the mean phase difference between the profile segments.  
267 This was evaluated from the angle of the cross-correlation of the complex profile segments,  
268 after applying the coarse offset. The estimates of coarse and fine displacement were added to  
269 provide an estimate of total vertical displacement between the two profiles (Fig. S1).

270 The displacement of the ice base was estimated in a similar manner. Amplitude cross cor-  
271 relation was used to determine the coarse offset, and complex cross-correlation to determine  
272 the phase difference. Assuming the ice base to be the strongest peak in each profile, the coarse  
273 offset was determined using the 10 m segment of the first profile above the peak of the basal  
274 reflector. Phase difference was determined from the leading edge of the basal peak, defined here

275 as the 1 m segment of the profile above the peak of the basal reflector.

276 The resulting estimates of vertical displacement throughout the profile were used to tune  
277 a model of displacement that was used to estimate the expected location of the base. The  
278 displacement model was formed to allow for the major processes expected to influence the  
279 observations, namely; hardware changes, accumulation, vertical strain and firm compaction.  
280 The model was of the form:

$$\delta z(z) = A + Bz + Ce^{(-z/z_0)}. \quad (1)$$

281 Here  $\delta z(z)$  is the modelled vertical displacement of internal reflectors as a function of  $z$ , the  
282 range from the antenna calculated using a nominal propagation velocity of  $1.68 \times 10^8 \text{ m s}^{-1}$ .  $A$   
283 represents a range independent offset allowing for hardware changes and surface accumulation.  
284  $B$  represents linear vertical strain, associated with horizontal convergence/divergence in the ice  
285 shelf. The third term on the right hand side represents an exponential model of firm compaction  
286 with surface compaction  $C$ , and length scale  $z_0$ . As many sites had insufficient depth to deter-  
287 mine  $z_0$  robustly, this was set to 21 m based on sites where this could be determined from the  
288 observations.

289 This model was used at all sites except the frontal site on Transect TB (tb\_00000). Here,  
290 curvature in the deep displacement observations suggested bending, possibly caused by e.g. a  
291 submarine keel at the ice front [56]. To allow for this deformation, a more complex displacement  
292 model which also allows for vertical displacement caused by bending of the ice shelf was used  
293 at this site as follows:

$$\delta z(z) = A + Bz + Ce^{(-z/z_0)} + D(z - z_n)^2. \quad (2)$$

294 The final term here represents the vertical displacement induced by bending as a quadratic  
295 function of distance from neutral depth  $z_n$  following ref. [57].

296 At each site the model free parameters were tuned to minimise the model-observation differ-  
297 ences in a least squares sense, using non-linear optimisation. Following tuning the model was  
298 used to determine the displacement of the base expected in the absence of basal ablation. The  
299 melt rate was then determined from the difference between the observed and expected vertical  
300 displacement of the basal reflector and the observation interval (Supplementary Fig. 1).

301 Formal errors in the melt rate estimates are typically 0.01 to 0.1 m yr<sup>-1</sup> (Tables S1 and S2),  
302 usually dominated by the 1 % uncertainty in the signal propagation speed in ice [55]. Systemic  
303 errors, such as the appropriateness of the firn compaction model are unaccounted for in the  
304 formal error estimate; however, the results are, within the stated errors, robust to reasonable  
305 variations in processing methods.

### 306 **Melt rate spatial interpolation**

307 In order to estimate net basal ablation from the survey region and to aid visualization of the  
308 melt rate field, the point-melt rate observations were interpolated onto a regular 1 km grid using  
309 the geostatistical interpolation method of Kriging [58]. Kriging is typically used where the  
310 underlying value of the assumed stochastic process being sampled is either constant (simple  
311 Kriging) or a linear function of position (Kriging with a trend). However, in this case melt  
312 rate variability is a strong non-linear function of frontal distance. To improve the statistical  
313 properties of the Kriged variable, the observed melt rates ( $m_o$ ) were first decomposed into a  
314 modelled component including a mean melt rate and a component of frontal melt enhancement  
315 ( $m_m$ ), and a melt rate anomaly ( $m_a$ ) (Supplementary Fig. 2), i.e.:

$$m_o(x, y, d) = m_m(d) + m_a(x, y). \quad (3)$$

316 Kriging was performed on the melt rate anomaly, and following interpolation, the frontal  
317 component determined for the grid location was added to the interpolated anomaly. This ap-

318 proach is closely related to regression-Kriging [59]; however, in this case the explanatory vari-  
319 ation is a non-linear function of the auxiliary variable (frontal distance).

320 Following ref. [34], frontal melt enhancement was modelled as an exponential function of  
321 frontal range. To accommodate the non-zero melt rates apparent at sites furthest from the ice  
322 front, a spatially constant term was added, providing a melt rate model of the form:

$$m_m = \alpha + \beta e^{(-d/d_f)}. \quad (4)$$

323 Here  $\alpha$  is a spatially constant background melt rate,  $\beta$  the magnitude of melt rate enhance-  
324 ment at the ice front,  $d$  the distance to the front of the Ross Ice Shelf, and  $d_f$  an e-folding  
325 length scale. While some frontal effect is evident on the McMurdo Ice Shelf, there are insuf-  
326 ficient observations near the McMurdo Ice Shelf front to reliably tune a model of frontal melt  
327 enhancement. Furthermore, differences in SST between the RIS ice front and McMurdo Sound  
328 (Fig. 3b) suggest that differences between frontal effects in the two regions are likely. For these  
329 reasons, we do not model frontal melt enhancement for the McMurdo Ice Shelf.

330 Least squares fitting of the model to the 78 melt rate observations provided parameter es-  
331 timates of  $\alpha = 1.29 \pm 0.09 \text{ m yr}^{-1}$ ,  $\beta = 5.0 \pm 0.4 \text{ m yr}^{-1}$  and  $d_f = 1900 \pm 300 \text{ m}$ . The model  
332 provides a good fit to the observations (Supplementary Fig. 2) with a coefficient of determi-  
333 nation  $R^2 = 0.78$ , confirming that much of the melt rate variability within the network can be  
334 described by this simple model of frontal melt enhancement.

335 Kriging with a linear trend was implemented using the mGstat toolbox [60] using a spher-  
336 ical semi-variogram with a sill of  $0.45 \text{ m}^2 \text{ yr}^{-2}$  and a range of 40 km; properties based on the  
337 observed spatial covariance of the melt rate anomaly (Supplementary Fig. 2). After Kriging,  
338 the modelled melt rates (Supplementary Fig. 2) were added to the interpolated anomaly field  
339 (Supplementary Fig. 2) to produce the final gridded melt rate estimate (Fig. 1a). Kriging  
340 also provides an estimate of the variance of the interpolated value, and the resulting melt rate

341 field was cropped to exclude land, sea, and regions where the standard deviation estimate was  
342  $>0.5 \text{ m yr}^{-1}$ .

### 343 **Mooring**

344 The sub-ice shelf mooring was deployed on 21 January 2011 through a hot-water bored access  
345 hole at  $77^{\circ}29.315' \text{ S}$ ,  $171^{\circ}34.272' \text{ E}$ , approximately 7 km south of the ice front. Ice shelf and  
346 water column thickness at the deployment site were 266.5 m and 578 m, respectively, and the sea  
347 bed depth was 798.5 m. This deployment followed the 2-month deployment discussed in [31].

348 Nortek Aquadop current meters and Seabird SBE-37 temperature salinity instruments were  
349 located at 4 levels throughout the water column from  $\sim 10 \text{ m}$  below the ice base to  $\sim 30 \text{ m}$  above  
350 the sea bed. Further details of the moored instrumentation and processing are provided in [54].

351 The ULS was composed of an upward looking Tritech PA-500 acoustic altimeter, configured  
352 as an external sensor to a Nortek Aquadop current meter, mounted to the mooring wire 15 m  
353 below the ice base. To avoid detecting the deployment borehole, the sensor was tilted  $13^{\circ}$  from  
354 the mooring wire, and its  $5^{\circ}$  acoustic beam width implies a  $\sim 2 \text{ m}$  diameter acoustic footprint  
355 at the ice base. The instrument sampled every second hour from January 2011, until battery  
356 failure in December 2014.

357 The raw range record required significant processing to remove outliers and noise. Pro-  
358 cessing included removing outliers, correcting for mooring swing, sensor tilt, and sound speed  
359 variations. Remaining scatter in the range observations of  $\pm 0.15 \text{ m}$ , attributed to basal rough-  
360 ness within the insonified region, was minimised by low-pass filtering the time series with a  
361 cut-off frequency of  $0.02 \text{ cycle day}^{-1}$ . Radar observations at the mooring site provide a precise  
362 and independent measure of basal melting that can be used to validate the ULS estimate. Be-  
363 tween observations on 16 January 2013 and 10 January 2014, the radar and ULS indicate mean  
364 melt rates of  $1.57 \pm 0.02 \text{ m yr}^{-1}$  and  $1.66 \text{ m yr}^{-1}$ . The  $0.09 \text{ m yr}^{-1}$  discrepancy between the



365 estimates is attributed to error in the ULS melt rate estimate. Further details of the instrument,  
366 processing and validation are given in [54].

### 367 **SST and sea ice concentration**

368 SST data were downloaded from the Group for High Resolution Sea Surface Temperature  
369 (GHRSSST), Multiscale Ultrahigh Resolution L4 archive [36]. The time series in Fig. 2c was  
370 formed from daily foundation SST fields at a nominal resolution of 5 km, averaged over the  
371 region within 100 km of the mooring. The spatial fields in Fig. 3 were derived from daily  
372 foundation SST fields at 25 km resolution, averaged over all January samples within the years  
373 2003-2018. The mean sea ice concentration time series in Fig. 2c and the monthly 15 % sea ice  
374 concentration contours in Figs. 3a and 3b are derived from the 25 km resolution NOAA/NSIDC  
375 Climate Data Record of Passive Microwave Sea Ice Concentration, Version 3, [42, 61]. The  
376 time series was created from daily sea ice fields, averaged over the region within 100 km of the  
377 mooring. Contours were created from monthly-mean fields averaged over the years 2010-2017.  
378 Daily and monthly data sets were downloaded from the National Snow and Ice Data Centre,  
379 `ftp://sidads.colorado.edu/pub/DATASETS/NOAA/G02202_V3/south/`.

### 380 **Polynya surface heat content**

381 To determine the available sensible heat content of the surface ocean layer within the polynya re-  
382 gion during January, we estimated its mean properties as follows. Considering the  $8.3 \times 10^4 \text{ km}^2$   
383 region within the  $0^\circ\text{C}$  SST isotherm (Fig. 3b), the mean January SST is  $0.22^\circ\text{C}$ . Based on the  
384 CTD cast in Figure 2b we assume a surface salinity of 34.4 psu. The freezing point potential  
385 temperature of this water at a depth of 200 m, typical of ice draft in the survey region (Fig 1a),  
386 is  $-2.04^\circ\text{C}$ . This indicates a temperature difference between the surface layer and the ice base,  
387 or thermal driving [62], of  $2.26^\circ\text{C}$ . Using a seawater density of  $1030 \text{ kg m}^{-3}$ , this implies a

388 thermal reservoir of  $8.3 \times 10^{17} \text{ J m}^{-1}$  of mixed layer depth. Archived CTD observations [63]  
389 indicate a wide range of mixed layer depths within this region during summer, typically exceed-  
390 ing 10 m. Assuming a mixed layer thickness of 10 m implies a surface layer sensible heat con-  
391 tent of  $8.3 \times 10^{18} \text{ J}$ . Using an ice density of  $916 \text{ kg m}^{-3}$ , and an initial temperature of  $-25 \text{ }^\circ\text{C}$ ,  
392 indicative of surface temperatures in the region [25], this is sufficient to melt 22 Gt of the ice  
393 shelf.

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563 **Author contributions**

564 CLS and MW designed and deployed the sub-ice shelf mooring. CLS designed the radar survey  
565 and CLS and PC undertook the radar fieldwork. CLS analysed the radar data with advice from  
566 KN. All authors contributed to the manuscript.

567 **Competing interests**

568 The authors declare no competing interests.

## 569 **Data and materials availability**

- 570 • Data for coastlines and ice shelves in Figs 2 and 3 are from the Antarctic Digital Database  
571 (<https://www.add.scar.org/>). The Ross Ice Shelf frontal line was modified to  
572 reflect recent changes in frontal position.  
573
  
- 574 • MODIS images in Figures 1a and 3c were sourced from the *NSIDC Images of Antarctic*  
575 *Ice Shelves* archive ([http://nsidc.org/data/iceshelves\\_images/index\\_](http://nsidc.org/data/iceshelves_images/index_modis.html)  
576 [modis.html](http://nsidc.org/data/iceshelves_images/index_modis.html)).  
577
  
- 578 • SST data is from *The Group for High Resolution Sea Surface Temperature (GHRSSST)*  
579 *Multi-scale Ultra-high Resolution (MUR) SST* record, obtained from the NASA EOSDIS  
580 Physical Oceanography Distributed Active Archive Center (PO.DAAC) at the Jet Propul-  
581 sion Laboratory, Pasadena, CA (<http://dx.doi.org/10.5067/GHGMR-4FJ01>).  
582
  
- 583 • Sea ice concentration data is from the *NOAA/NSIDC Climate Data Record of Passive Mi-*  
584 *crowave Sea Ice Concentration, Version 3* (<https://doi.org/10.7265/N59P2ZTG>).  
585
  
- 586 • Offshore CTD data is from the *Calibrated Hydrographic Data from the Ross Sea acquired*  
587 *with a CTD during the Nathaniel B. Palmer expedition NBP1101 (2011)* dataset ([https :](https://doi.org/10.1594/IEDA/317595)  
588 [//doi.org/10.1594/IEDA/317595](https://doi.org/10.1594/IEDA/317595)), downloaded from [http://www.marine-geo.](http://www.marine-geo.org/tools/search/Files.php?data_set_uid=17595)  
589 [org/tools/search/Files.php?data\\_set\\_uid=17595](http://www.marine-geo.org/tools/search/Files.php?data_set_uid=17595).  
590

591 • Mooring data are archived at <https://figshare.com/s/1c92fe8eb227b878e344>.

592

593 • ApRES radar data are archived at <https://figshare.com/s/1255b7d76ed69e015c3a>.

## 594 **Code availability**

595 Computer code used to process the radar observations is available from the corresponding au-  
596 thor on request.

597 **Figure captions**

598 **Figure 1**

599 Basal melt rates of the north-western Ross Ice Shelf. (a) Annual-mean basal melt rate observa-  
600 tions (coloured dots) and interpolated melt rate field (background colour) with mooring location  
601 (red star) and labeled Transects TA-TC. Grey boxes indicate sites within transects. Also shown  
602 are ice thickness contours in m (grey) and the shear zone (dashed) that separates the Ross and  
603 McMurdo ice shelves (RIS and MIS, respectively). Background is Modis image from 10 Nov  
604 2015 [64]. Panels (b) and (c) show annual-mean (b) and short-term summer (c) melt rates for  
605 frontal Transects TA-TC. Also shown in (b) is the mean melt rate from the Upward Looking  
606 Sounder (ULS) (red star).

607 **Figure 2**

608 Oceanographic conditions and melt rate variability. (a) Depth-averaged flow at the mooring  
609 site for winter (May-November; blue) and late summer (February-March; green) for calendar  
610 years 2011-2014. Also shown are the mooring site (star), the R.V. Palmer offshore CTD cast  
611 [37](triangle), and the region within 100 km of the mooring used to average SST and sea ice  
612 concentration (light blue shading). (b) TS observations from the offshore CTD cast (triangles)  
613 and from the highest sub-ice shelf TS sensor, located  $\sim 8$  m below the ice base (stars). Also  
614 shown are the in-situ freezing point at the ice-base (blue line), representative water properties of  
615 High Salinity Shelf Water (HSSW), modified Circumpolar Deep Water (mCDW) and Antarctic  
616 Surface Water (AASW) [65], and water masses possibly derived from the offshore surface layer  
617 through interaction with the ice shelf (grey shading). Arrows indicate TS evolution associated  
618 with solar heating and ice melt. (c) Mean SST [36] (light blue) and sea ice concentration [42,61]  
619 (grey, right scale) within 100 km of the mooring (see Methods), and temperature measured  
620  $\sim 13$  m below the ice base (dark blue). Also shown is SST from the offshore CTD cast (red

621 triangle). (d) ULS basal melt rates (black) and low-pass filtered sub-ice shelf water temperature  
622 (blue, right scale).

623 **Figure 3**

624 SST and sea-ice concentration around Antarctica. (a) Long-term mean January SST [36] with  
625 15 % sea ice concentration [42, 61] contour (red) (see Methods). (b) Close up of the Ross  
626 Sea additionally showing the January mean 0 °C SST isotherm (white) and the 15 % sea ice  
627 concentration contours for November (blue) and December (green). (c) Modis image during  
628 strong southerly airflow on 12 Nov 2015 [64]. Grey boxes in (a) and (b) indicate the regions  
629 shown in (b) and (c), respectively. The red star indicates the mooring location.







