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Sea-ice decline exposes Antarctic ice shelves to unprecedented retreat by wave erosion

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Abstract

Recent mass loss from the Antarctic Ice Sheet has been driven by a reduction in ice-shelf buttressing. Reliable sea-level projections therefore require models to represent the processes governing ice-shelf loss; where key mechanisms are missing, retreat and sea-level rise will be underestimated. Here, we investigate the impact of emerging summer sea-ice decline on ice shelves in the Amundsen Sea Polynya, which already experiences persistent sea-ice-free summers, and the Ronne Entrance, which in 2025/2026 became one of the first regions outside a polynya to remain sea-ice-free for an entire summer. We show that the year-on-year retreat of sectors of the Getz Ice Shelf bordering the Amundsen Sea Polynya is the result of summer sea-ice loss, whereby wave erosion by seasonally heated warm surface waters undercuts the calving front, forming buoyant underwater feet whose flexure drives frontal ablation rates of 680 m a⁻¹. Even higher rates exceeding 2,000 m a⁻¹ from this mechanism are observed in the Ronne Entrance. If comparable rates occur more widely in response to expected sea-ice decline, nearly all Antarctic ice shelves, including those considered stable, will undergo sustained year-on-year retreat and will be unviable in their present configurations. This mechanism linking sea-ice decline and ice shelf retreat represents a potent driver of continent-wide ice-shelf loss currently omitted from sea-level projections.

Main

The Antarctic Ice Sheet is fringed by ice shelves whose buttressing restrains the flow of grounded ice from the ice sheet interior into the ocean^{1,2}, and whose deterioration has driven the ice sheet's accelerating contribution to sea-level rise. Outside the Antarctic Peninsula, where final ice-shelf

collapse has been linked to surface melt³⁻⁵, weakening is driven by intrusions of warm modified Circumpolar Deep Water (mCDW) melting the undersides of ice shelves⁶. This ocean-induced thinning causes ice shelves to unanchor from pinning points⁷, initiating a feedback in which thinning, acceleration and damage become mutually reinforcing⁸. Efforts are underway to incorporate these processes into next-generation sea-level projections⁹. However, projections may remain biased low if additional mechanisms capable of driving ice-shelf retreat have yet to be identified or represented¹⁰.

Over the coming decades, Antarctic sea ice is expected to continue to decline, resulting in extended sea-ice-free summers adjacent to the coastline^{11,12}. The decline of protective landfast and pack ice has already been shown to increase ice-shelf vulnerability to flexure-induced failure^{13,14}. But the seasonal loss of sea ice could also expose ice shelves to enhanced wave erosion at the ice-front waterline, a process not accounted for in any ice sheet model. In Alaska, warm near-surface water masses have been observed to drive seasonal melt rates up to 8 m d⁻¹ at the waterline of the calving face¹⁵. When combined with melt at depth, this can drive even greater frontal ablation rates¹⁶ on the order of 30 m d⁻¹ through buoyancy-driven calving multiplier effects^{17,18}. These high frontal ablation rates occur at glaciers without floating ice shelves and in warmer climatological settings than those currently in Antarctica. However, Antarctic ice shelves flow at much slower speeds, meaning even frontal ablation rates an order of magnitude lower could drive widespread frontal retreat. Here, to investigate this possibility, we analyse the calving dynamics of ice shelves flowing into the Amundsen Sea Polynya (ASP) and the Ronne Entrance. Over recent decades, the ASP has been the only Antarctic location where sea-ice decline has resulted in consistent, annual open-water conditions throughout the summer^{19,20}. In the 2025/2026 austral summer, the Ronne Entrance became one of the first regions to reach the same condition outside a polynya environment. Consequently, the calving processes operating in these areas serve as critical sentinels for how other sectors of the ice sheet will respond to projected widespread summer sea-ice loss.

Ice front retreat in the Amundsen Sea Polynya

The ASP forms each spring as prevailing southeasterly coastal winds blow over Bear Ridge peninsula (Extended Data Fig. 1), where numerous small icebergs ground to form a longitudinal barrier to drifting sea ice^{19,20}. Over the past five decades, increased iceberg deposition along Bear Ridge, driven by the disintegration of the Thwaites Ice Tongue²¹, together with strengthening southeasterly winds²⁰, has made the ASP more persistent, with the polynya now forming one to

two months earlier in spring (Extended Data Fig. 1). As a result, the ice fronts of the Dotson Ice Shelf and the easternmost outlet of the Getz Ice Shelf, here termed Getz A, are almost entirely sea-ice-free each summer (Fig. 1a). Further west, Getz B and C are sea-ice-free in most summers, depending on the lateral extent of the polynya. Getz D–G lie beyond the direct influence of the ASP, and although sea-ice cover there is declining, sea ice has remained immediately adjacent to these ice fronts throughout most summers.

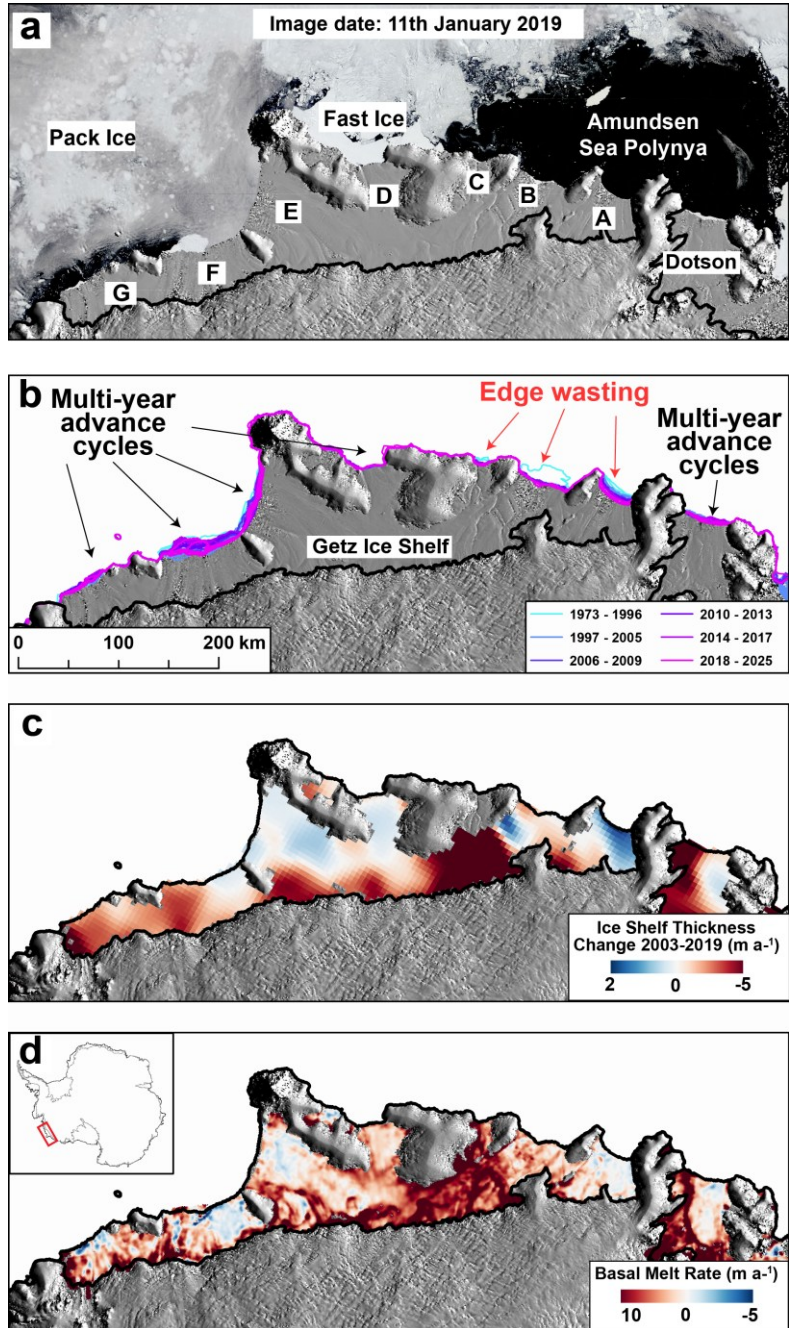


Figure 1: The Dotson Ice Shelf and sections of the Getz Ice Shelf that flow into the ASP. **a)** MODIS image showing the typical extent of the ASP in January. **b)** Ice shelf frontal extent change 1973-2025. Dotson and Getz A-C, that flow into the ASP have all retreated. **c)** Ice-shelf thickness change 2003-2019⁶. **d)** Basal melt rates 2010-2020²². Note the thickening and low

basal melt rates at Getz A. The underlying image in all panels is the REMA DEM²³ and the black line is the MODIS-derived 2009 grounding line²⁴.

We use the Landsat archive to quantify ice-front change from 1972 to 2025 on annual to decadal timescales and to derive decadal frontal ablation rates for Getz A and B, the only outlets where surface features can be reliably tracked (Methods). Seasonal calving behaviour between 2014 and 2025 is resolved using Sentinel-1 SAR imagery for all outlets. All ice-shelf outlets draining into the ASP exhibit unambiguous frontal retreat since 1972 (Fig. 1b, Fig. 2). The Dotson Ice Shelf has retreated by 7 km and produced infrequent tabular icebergs, with multi-year periods of uninterrupted advance (Fig. 2a). No consistent seasonality is observed, as the ice front advanced without major calving events between 2016 and 2020 (Fig. 2e). Getz A has retreated by 13.3 km and is characterized by persistent year-on-year retreat via edge wasting²⁵ (Fig. 2b), with frontal ablation rates increasing from $380 \pm 30 \text{ m a}^{-1}$ in 1973–1989 to $680 \pm 30 \text{ m a}^{-1}$ in 2013–2025 (Extended Data Fig. 2), driving accelerated ice-front retreat. Getz B produced large tabular icebergs in the early 1970s before transitioning to a similar edge-wasting regime from the 1980s onward, resulting in a net retreat of 16.1 km (Fig. 2c). At Getz B, frontal ablation rates increased from $490 \pm 30 \text{ m a}^{-1}$ in 1973–1989 to $540 \pm 30 \text{ m a}^{-1}$ in 2013–2025, with a transient reduction to $380 \pm 30 \text{ m a}^{-1}$ during 2000–2013 (Extended Data Fig. 2). Getz C has retreated by 6.8 km and exhibits comparable edge-wasting behaviour (Fig. 2d). Getz A and B display pronounced seasonal calving, in which winter ice-front advance is offset by numerous small summer calving events that drive net recession; this seasonal signal is weaker at Getz C (Fig. 2f–h). In contrast, at Getz D–G, which do not drain into the ASP, we observe only minor overall retreat, multi-year periods of uninterrupted frontal advance, no edge wasting and no clear seasonal calving pattern (Extended Data Fig. 3).

Frontal retreat at the Dotson Ice Shelf coincides with intense channelized basal melting and associated thinning²⁶, leading to pinning-point loss⁷ and grounding-line retreat²⁷. Similar patterns of channelized basal melt and thinning occur at Getz C^{28,29}, consistent with its concurrent ice-front retreat. This explanation does not apply to Getz A and B, where basal melt rates have remained near zero (Fig. 1d). At these outlets, ice-shelf thickness changes are small, with slight thickening at Getz A and slight thinning at Getz B (Fig. 1c). Moreover, neither outlet exhibits appreciable flow acceleration³⁰, detectable changes in structural integrity between 1972 and 2025 (Extended Data Fig. 4) or substantial summer surface melt³¹. The absence of these commonly invoked drivers of ice-shelf retreat indicates that frontal retreat at Getz A and B is driven by a distinct mechanism, operating independently of basal and surface melting.

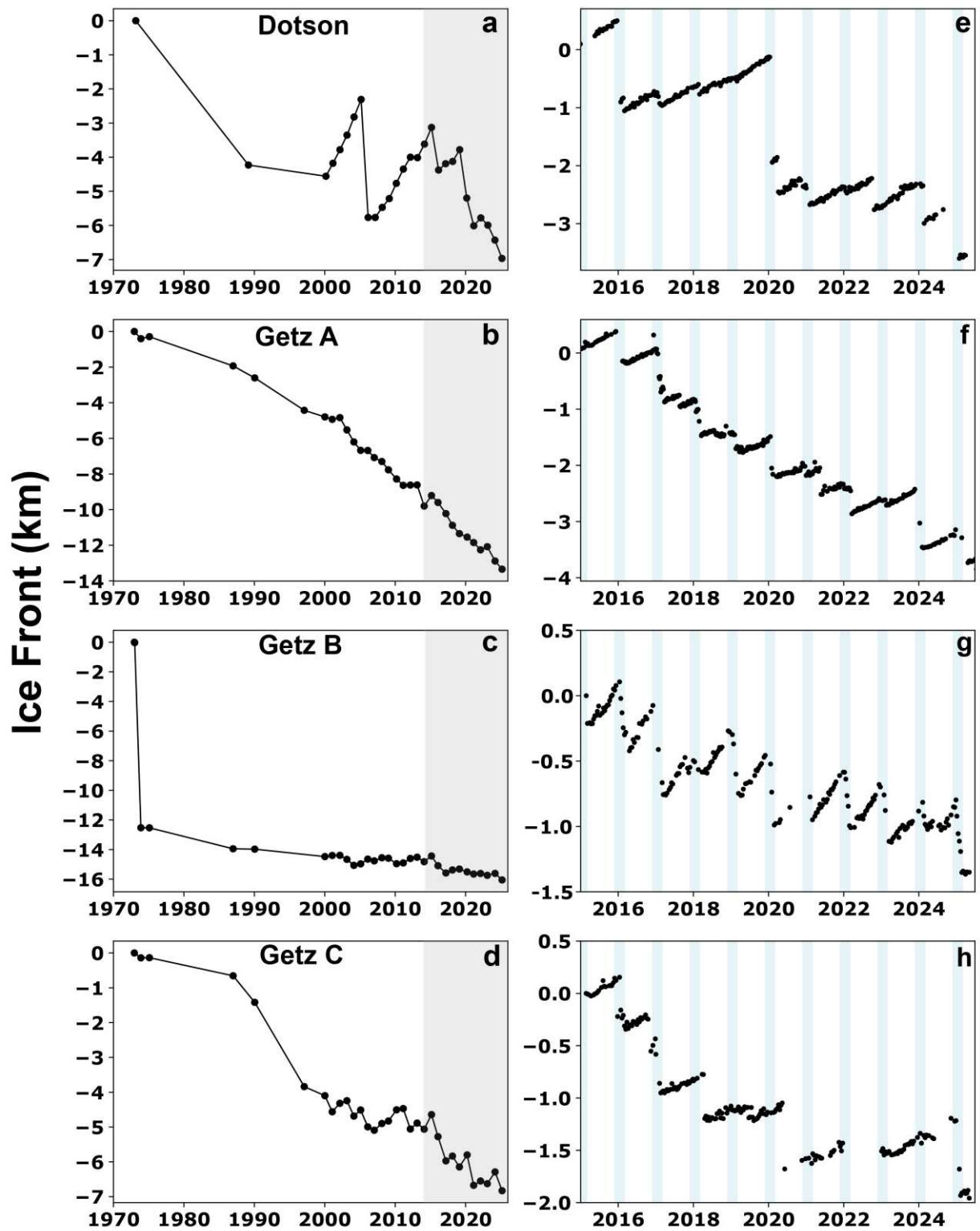


Figure 2: Timeseries of ice-front change for Dotson Ice Shelf and Getz A-C outlets. **a-d)** Annual ice-front change from 1972-2025 from Landsat and MODIS imagery. The grey shadings represent the period covered via Sentinel-1 shown in plots e-h. **e-h)** Weekly observations of ice-front change from Sentinel-1 2014-2025. The blue shadings represent the austral summer months December – February.

Warm sea surface temperatures drive frontal retreat

Repeat ICESat-2 surface profiles reveal widespread, rapidly developing rampart–moat structures along the Getz A–C ice-shelf fronts prior to calving (Fig. 3). These structures form when melting at the waterline undercuts the cliff front. This uneven erosion leaves behind a highly buoyant, submerged ice foot²⁵ whose upward flexure of the ice front produces the rampart–moat profile. As this foot enlarges, its upward buoyant force generates tensile stresses that eventually exceed the strength of the ice, triggering calving via the 'footloose mechanism'³² (Extended Data Fig. 5). While the footloose mechanism is known to trigger tabular iceberg break-up in warm waters^{25,32,33} and drive calving at Northern Hemisphere tidewater glaciers^{34,35}, it has previously been tied to only a single ice-shelf calving event at the Ross Ice Shelf³⁶. Here, our observations indicate that footloose-style calving is widespread along the Getz A–C outlets (Fig. 3) and acts as the dominant mechanism of frontal ablation, consistent with observed seasonal calving (Fig. 2f-h) and ice-front retreat. This mechanism is directly supported by ICESat-2 altimetry, which, under optimal conditions for subsurface detection, resolved an underwater foot protruding ~60 m seawards at Getz A (Extended Data Fig. 6). In contrast, rapidly developing rampart–moat structures are absent at the Dotson Ice Shelf, matching its lack of seasonal calving (Fig. 2e).

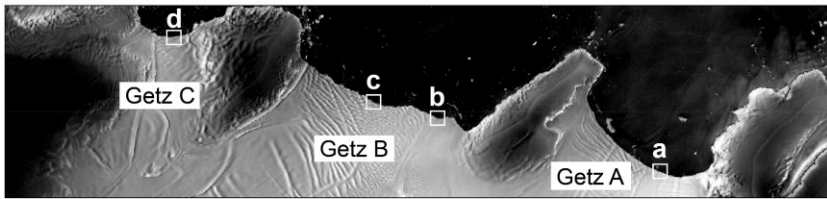
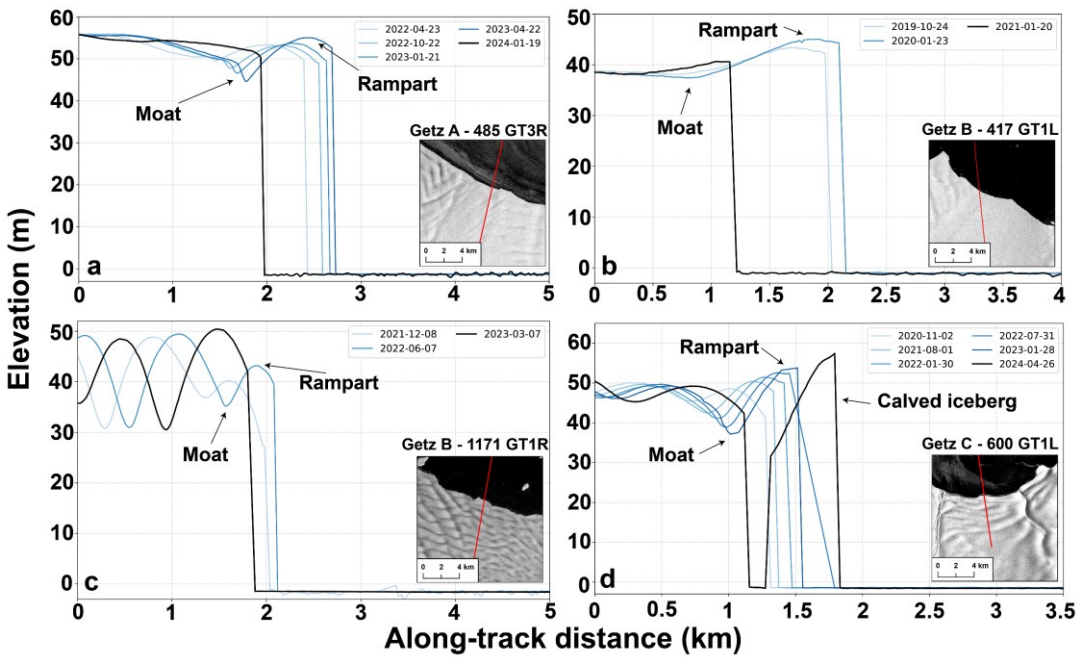


Figure 3: ICESat-2 repeat profiles showing the development of rampart-moat structures (blue lines) followed by calving events (black lines). Location of the ICESat-2 track are shown in the inset images in each panel and overview image at the bottom. The insets show Sentinel-1 SAR images with the closest date to the final ICESat-2 profile. The overview image at the bottom is a Sentinel-1 SAR image from February 2023.

The spatial extent of the ASP and its complex coastline impose fetch-limited conditions that inhibit wave growth and suppress mechanical wave energy at ice fronts³⁷, compared to open-ocean conditions. Therefore, the thermal component of warm SSTs is crucial for waterline erosion within the ASP, which is enhanced by up to 0.3 °C in some years via phytoplankton blooms³⁸. Ship-based observations consistently document SSTs at the Dotson Ice Shelf approximately 1.5 °C lower than those recorded at the Getz outlets³⁹⁻⁴¹, a pattern corroborated by monthly top-of-atmosphere brightness temperatures derived from Landsat-8 and Landsat-9 thermal infrared imagery acquired near the ice-shelf fronts (Extended Data Fig. 7). The physical mechanism underlying this SST contrast remains unclear, though one plausible contributor is meltwater input from seasonally ablating grounded icebergs on Bear Ridge. To estimate summer melt rates at the waterline of the Dotson and Getz ice-shelf outlets, we apply a thermodynamic ocean-wave melt-rate parameterization developed for ice cliffs⁴². Adopting ship-based SST values of 0.5 °C for Getz and -1 °C for Dotson, and assuming wave heights of 0.5 m and periods of 2–4 s (Methods), we obtain melt-rate estimates of 0.2–1.2 m d⁻¹ for Getz and 0.07–0.40 m d⁻¹ for Dotson. While these values are subject to uncertainties in local wave conditions, the systematically lower temperatures and resulting melt rates at Dotson align with its lack of widespread rampart-moat structures. This suggests that cooler frontal ocean temperatures suppress ice-foot formation.

In the absence of three-dimensional in situ observations, estimating the underwater foot length, perpendicular to the ice front, required to drive footloose-style calving is challenging. Idealised modelling suggests the critical foot length is only weakly dependent on ice-shelf thickness, with calving estimated to be triggered by feet ~50–100 m in length^{25,32,33}, which is within the range of our ICESat-2 observation (Extended Data Fig. 6). Because waterline erosion occurs almost exclusively during the summer, reaching this critical threshold typically requires two successive melt seasons (Fig. 3), implying an erosion rate of 25–50 m per summer at the Getz Ice Shelf outlets. Elastic beam modelling predicts calving length that is strongly controlled by ice-shelf frontal thickness^{33,36} (Methods). Mean ice-shelf frontal thicknesses of ~350 m and ~270 m correspond to predicted calving lengths of ~880 m and ~725 m, matching observed means of 900 m and 730 m for Getz A and Getz B respectively (2014–2025; Extended Data Fig. 8). Despite greater crevassing and damage at Getz B compared to Getz A, this close agreement

indicates thickness, not pre-existing crevasses, primarily controls calving length. The observed increase in frontal ablation at Getz A over the past 50 years (Extended Data Fig. 2) may therefore reflect not only more persistent sea-ice-free conditions (Extended Data Fig. 1) but also a positive feedback in which retreat into thicker ice increases calving length, and hence frontal ablation, promoting further retreat.

The sensitivity of frontal ablation rates to interannual variability in sea-ice conditions provides further evidence that retreat of the Getz Ice Shelf outlets is driven by the relatively warm surface water environment in the polynya. Since 2000, when annual imagery became available, the Getz A outlet failed to retreat in only four intervals: 2001-2002, 2011-2013, 2014-2015, and 2022-2023 (Fig. 2b). The first two periods coincide with anomalously short sea-ice-free summers (Extended Data Fig. 1). The third corresponds to the only year in which a narrow band of landfast sea ice persisted into January, while the fourth coincides with the breakout of an armada of icebergs from Bear Ridge Peninsula, that drifted into the ASP and likely cooled SST (Extended Data Fig. 9), associated with the ungrounding of iceberg B-22A. If the seasonal absence of summer sea ice continues, Getz A will retreat deeper into its embayment. This would reduce sidewall contact and could trigger acceleration and damage feedbacks that drive self-sustaining retreat⁸ and possible collapse of the ice shelf. In the following section, we assess the potential for the footloose mechanism to become widespread along the circumference of the Antarctic Ice Sheet as summer sea ice declines, thereby increasing the vulnerability of ice shelves to larger-scale collapse.

Widespread ice shelf loss in response to sea-ice decline

We quantify the mean maximum number of consecutive sea-ice-free days (Methods) at all Antarctic ice fronts during December and January, when solar insolation is at its annual maximum, for the periods 1979–2016 and 2016–2024. We relate these values to the frontal velocities of the corresponding ice shelves, because sustained retreat ensues wherever frontal ablation outpaces ice delivery to the front. From 1979-2016, only the Dotson, Getz and Ross Ice Shelf outlets exceeded 40 mean sea-ice-free days; all others remained below 30 days, with only 14% above 20 days and a median of 7 days (Fig. 4). These conditions indicate that nearly all Antarctic ice shelves have been shielded from enhanced frontal ablation associated with the footloose mechanism because they are encased by sea ice throughout summer. Since 2016, however, Antarctic sea ice has declined markedly¹¹ and the number of sea-ice-free days at ice fronts has increased, with 41% above 20 days and a median of 16 days (Fig. 4). This decline is

expected to continue, resulting in widespread sea-ice-free summers across coastal Antarctica by the late twenty-first century^{11,12}. This would expose most ice shelves to seasonally warm SSTs and, along coastlines where fetch is not limited as it is within the ASP, to substantially greater wave energy than the Getz outlets currently experience.

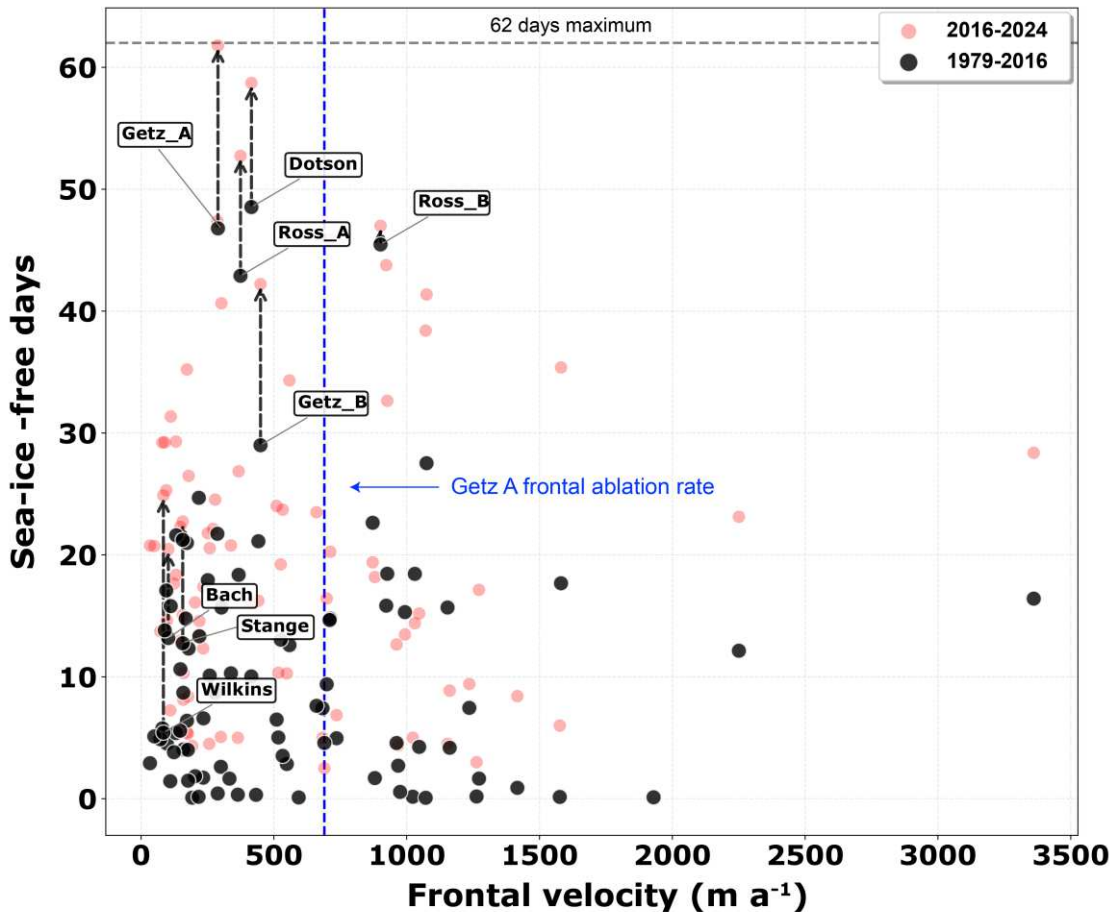


Figure 4: Mean consecutive sea-ice-free days in December and January adjacent to each major ice shelf outlet in Antarctica relative to mean frontal velocity from the MEaSURES ice velocity mosaic of Antarctica⁴³. The black dots represent the long-term average 1979-2016, the pink dots represent 2016-2024 showing an increase in sea-ice free days.

Over the past decade, the most pronounced sea-ice reduction has occurred in the Bellingshausen Sea⁴⁴; despite strong interannual variability, sea ice was entirely absent from the Ronne Entrance embayment during the 2025–2026 austral summer (Fig. 5). All floating ice shelves within the embayment retreated during this period, with retreat characterised by edge wasting and the presence of rampart-moats structures, both hallmarks of the footloose mechanism (Fig. 5). The sole exception was the western front of the Wilkins Ice Shelf, which remained fringed by sea ice throughout the summer and exhibited no detectable calving. Across the eight ice-shelf sections within the embayment, frontal ablation rates averaged 1,300 m a⁻¹, with a peak at 2,100 m a⁻¹ at

Wilkins Inlet and a minimum of 900 m a^{-1} at the Verdi Ice Shelf. Though based on a single year of data, these ablation rates are substantially higher than those observed at the Getz Ice Shelf. These enhanced frontal ablation rates are likely driven by relatively warmer SSTs, with ship-based observations recording maximum sea surface temperatures of 2.2°C within the Ronne Entrance during 2022-2023⁴⁵, a summer that also experienced below average sea ice conditions. In all cases, frontal ablation rates far exceed the rate of ice front advance, meaning that if sea-ice-free conditions persist, these ice shelves will undergo similar year-on-year retreat to Getz A, but at greater rates given the larger difference between frontal velocity and frontal ablation rates. In the case of the Wilkins Ice Shelf, collapse could occur imminently if retreat were to erode the ice bridge connecting it to Latady Island^{46,47}. Elsewhere, the Bach Ice Shelf will steadily retreat into its embayment, priming it for hydrofracture-induced collapse should this retreat generate crevassing, given the extensive summer ponding already present⁴⁸.

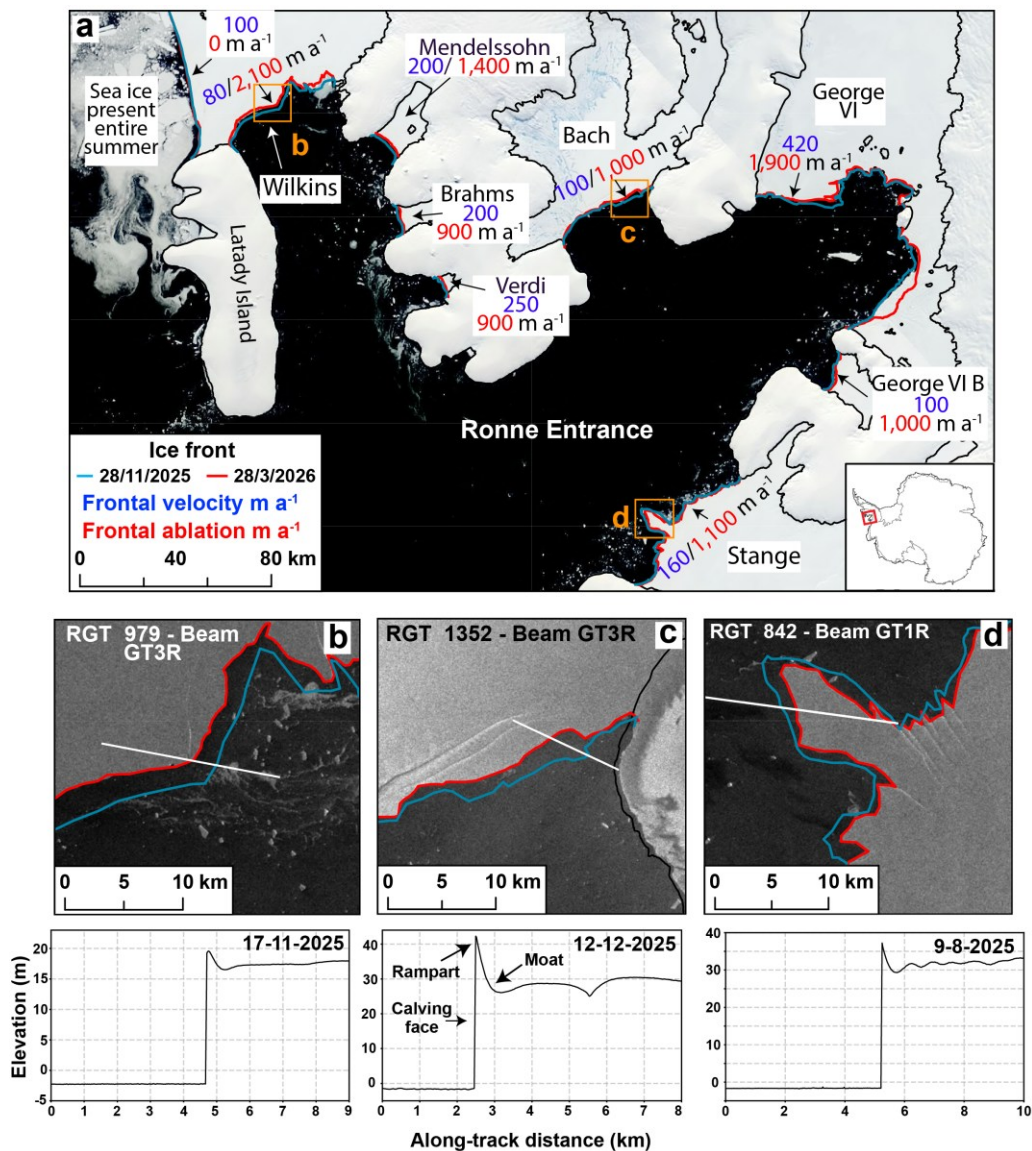


Figure 5: Retreat of ice shelves in the Ronne Entrance during austral summer 2025/2026. **a)** MODIS image from 9 February 2026 showing sea-ice-free conditions; the Ronne Entrance had been sea-ice-free since early December. Overlain are ice front positions from 28th November 2025 and 28th March 2026, along with frontal velocity rates (blue text) calculated from the MEaSUREs Antarctic velocity mosaic⁴³ and frontal ablation rates (red text) calculated for 2025/2026. Note that the only section with zero frontal ablation is fringed by sea ice. **b–d)** Sentinel-1 images from 28th March 2026 showing characteristic edge-wasting retreat, with overlain ICESat-2 tracks (white lines) and corresponding profiles containing rampart-moat structures shown below.

Frontal ablation rates at the Getz Ice Shelf and Ronne Entrance surpass the pan-Antarctic median ice shelf frontal velocity of 370 m a^{-1} , suggesting that the majority of ice shelves are primed for sustained retreat following summer sea-ice-loss. This includes historically stable regions like Dronning Maud Land, where widespread low frontal velocities ($\sim 100 \text{ m a}^{-1}$) leave ice shelves especially susceptible to rapid retreat via the footloose mechanism. Local environmental conditions, however, have the potential to limit this impact. For instance, the Ross Ice Shelf front is free of sea ice throughout most summers (Fig. 4), yet footloose-style calving and frontal ablation remain limited³⁶ despite the presence of pervasive rampart-moats⁴⁹. Local winds force solar-heated surface waters beneath the ice front, driving strong localised basal melt^{50,51}. This subsurface warming may suppress footloose-style calving through two pathways. First, by eroding the buoyant subaqueous foot at depth during its formation. Second, via a wedge mechanism in which cool, fresh meltwater rising from depth shields the upper ice front from thermal erosion⁵². Consistent with this, the characteristic wavelengths of the Ross rampart-moat structures indicate formation by internal bending stresses rather than buoyant undercutting⁵³. Consequently, localised oceanographic dynamics can insulate specific regions from buoyancy-driven calving even amid widespread sea-ice loss.

Overall, our results indicate that the potential for sea-ice decline to fundamentally alter the configuration of pan-Antarctic ice shelves through persistent edge-wasting has been overlooked. In the current generation of numerical models used to project sea-level rise, ice-shelf loss is driven only by surface melting and basal melting associated with intrusions of warm mCDW⁵⁴. This restricts projected ice-shelf loss to warmer northern latitudes and to regions where bathymetry permits mCDW access. By the end of the century, however, the seasonal loss of coastal sea ice is expected to expose most Antarctic ice shelves to warmer SSTs and elevated wave energy, supercharging wave-erosion, and subsequent buoyancy-driven calving. This would promote pan-Antarctic ice shelf retreat, rendering the majority of ice shelves unviable in their current configurations. Any grounded-glacier calving faces exposed after ice-shelf loss would also be vulnerable to this process. Representing iceberg calving and incorporating the full effects

of sea-ice decline into future projections, including enhanced wave erosion and ice-shelf flexure^{13,14}, is therefore urgently required; otherwise, projections will systematically underestimate Antarctic mass loss and its contribution to global sea-level rise.

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Author contributions: B.W.J.M led the development of the study, with input from T.J.W.W and A.J.C. B.W.J.M led the data collection and analysis with input from N.H and R.G.B for the ice front positions, input from T.J.W.W and A.J.C on calving statistics and rampart-moats, input from R.N.I on Landsat thermal images and from N.G on basal melt rates. B.W.J.M led the manuscript writing with input from all authors.

Data Availability: All Landsat data used in this study are freely available to download from USGS Earth Explorer (<https://earthexplorer.usgs.gov/>). Sentinel-1 data is freely available from the Copernicus Browser (<https://browser.dataspace.copernicus.eu/>). MODIS data is freely available from NASA WorldView (<https://worldview.earthdata.nasa.gov/>). Sea ice concentration data⁵⁵ is available at <https://doi.org/10.5067/MPYG15WAA4WX>. ICESat-2 ATL03⁵⁶ data is available at <https://doi.org/10.5067/ATLAS/ATL03.007> and ATL06⁵⁷ at <https://doi.org/10.5067/ATLAS/ATL06.007>. *Nathaniel B. Palmer* NBP2002 cruise data⁴¹ is available at <https://doi.org/10.7284/154878>. *Polarstern* PS134 cruise data⁴⁵ are available at 10.1594/PANGAEA.964292. All shapefiles of the Getz and Dotson ice shelves will be made freely available following acceptance of this manuscript.

Methods

Ice-front extent change and frontal ablation

We mapped ice-front positions from 1972 to 2025 using imagery from the Landsat archive and MODIS. Prior to 2000, observations were intermittent and were derived from Landsat 1 (60 m resolution) and Landsat 4 and 5 (30 m resolution). From 2000 onward, consistent annual coverage was achieved using Landsat 7 and, subsequently, Landsat 8 and 9 (all 30 m resolution), supplemented by coarser MODIS imagery (250 m resolution). Because calving at Getz is strongly seasonal, imagery was preferentially selected from late February to March each year to maintain inter-annual consistency. Ice-front change was quantified using the box method⁵⁸, which accounts for spatial variability in advance and retreat along the ice front.

We developed a workflow that combined the Segment Anything Model (SAM)⁵⁹ with the annually mapped ice fronts described above to automatically track ice-front positions from Sentinel-1 IW HH imagery acquired between November 2014 and May 2025. These data have an approximate spatial resolution of 40 m and a repeat interval of about 7 days. Each Sentinel-1 scene was divided into 15×15 km tiles that overlapped by 500 m and were positioned along the corresponding annual ice front, so that every scene from a given year was tiled relative to that year's front (Fig. S1). SAM was then applied to each tile to segment the ice shelf (Fig. S2). Because the tiles overlapped, the true ice front was sampled repeatedly and could be identified as the region with the highest density of detections; this suppressed false positives caused by variable environmental conditions, such as frazil streaks or drifting sea ice adjacent to the ice shelf (Fig. S3 and S4). Even so, small gaps remained in the record, particularly at Getz F and G, where recently calved icebergs persisted close to the ice front and obscured its position.

We estimated mean decadal frontal velocity and frontal ablation rates for the Getz A and B ice shelf outlets by manually tracking stable surface features, such as crevasses, near the ice front between satellite image pairs from the periods 1973–1989, 1989–2000, 2000–2013, and 2013–2025. For each pair, we manually displaced the first image until features near the ice front aligned with the same features in the second image. The resulting width-averaged displacement was taken as the frontal velocity, and the width-averaged offset remaining between the two ice fronts after this displacement was taken as the frontal ablation rate. We estimated uncertainty as ± 30 m a^{-1} ; for 1973–1989, this equates to a positional uncertainty of ± 480 m and therefore can be considered conservative for feature tracking alone. However, we used this higher value to qualitatively take into account factors such as subtle changes in the shape of features and ice

motion trajectories through time. We could not estimate frontal ablation rates at the Dotson and Getz C ice shelf outlets because of a lack of stable features near the ice front through time.

ICESat-2 profiles

We used the ICESat-2 ATL06 product⁵⁷, corrected for tides, the dynamic atmosphere and the geoid, to manually examine the development of rampart-moat structures at the ice fronts of the Getz and Dotson ice shelves between 2018 and 2025. All reference ground track profiles crossing the ice fronts of the Dotson Ice Shelf and the Getz A–C outlets were inspected. We additionally used the ATL03 product⁵⁶ to manually assess all beam profiles along reference ground track 485 over the same period for the presence of underwater ice feet. Under optimal conditions, ICESat-2 can penetrate water to depths of up to 38 m⁶⁰.

SSTs and melt rate

We used sea surface temperature observations collected in the Amundsen Sea during a research cruise aboard the RVIB *Nathaniel B. Palmer* between 12 January and 25 February 2022⁴¹. The cruise reached the Dotson Ice Shelf on 22nd January, Getz B on 14th February and Getz A on 22nd February. The Getz outlets were therefore sampled later in the austral summer than Dotson, so the temperatures recorded there are likely further from the seasonal maximum (Extended Data Figure 7b). On this basis, the maximum sea surface temperature of 0.5 °C recorded at Getz B is likely an underestimate of the seasonal maximum during the 2021–2022 austral summer.

Seasonal surface thermal patterns adjacent to each ice shelf outlet were investigated using Top of Atmosphere (TOA) Brightness Temperature derived from the Landsat 8 and 9 Collection 2 Level 1 product⁶¹ accessed via the Google Earth Engine platform⁶². Thermal infrared Band 10 was selected for thermal analysis due to its superior radiometric calibration and lower susceptibility to stray light artifacts compared to the other thermal infrared Band 11⁶³. We compiled all Landsat scenes acquired between 2013 and 2025 during the austral summer period from November to March that cover the ice fronts of the Dotson and Getz A to C outlets. Pixels affected by cloud, sea ice, and icebergs were excluded by masking values below 269 K as no data. For each scene, we extracted the median brightness temperature within predefined regions adjacent to the ice shelf outlets (white boxes, Extended Data Figure 7a), retaining only those scenes where at least 75 percent of pixels were valid. Brightness temperature was not converted to sea surface temperature because no calibrated product is currently available for Antarctic waters. Instead, we analyse the seasonal cycle of brightness temperature to compare relative conditions adjacent to each ice shelf⁶⁴, which indicate cooler surface waters near the Dotson Ice Shelf compared to the Getz ice shelf outlets.

To estimate the waterline melt rate at the ice fronts we adapt the parametrization of Wolterman et al⁴²:

$$\frac{dm}{dt} = \left(\sqrt{\frac{6\alpha c_p}{\pi g L}} \right) a \omega^{5/2} \Delta T$$

Where $\frac{dm}{dt}$ is the horizontal melt rate (in m s⁻¹), $\alpha = 1.4 \times 10^{-5}$ m² s⁻¹ is the thermal diffusivity of ice, $c_p = 4.2 \times 10^3$ J kg⁻¹ K⁻¹ is specific heat capacity of water, g is the acceleration due to gravity, and $L = 3.34 \times 10^5$ J kg⁻¹ is the latent heat of fusion of water. Here, wave amplitude is a , angular frequency is ω and the thermal driving potential is $\Delta T = |T_w - T_f|$, where the freezing temperature of sea water is $T_f = -1.8$ °C and T_w corresponds to SST.

We used SST values of 0.5° C for Getz and -1° C for Dotson based on the observations by the Nathaniel B. Palmer. We do not extract significant wave height and wave period from coarse global wave models because they do not resolve the complex headlands and embayments along the ice shelf coastlines. Instead, we estimate significant wave height of 0.5 m and wave periods of 2-4 s based on the likely fetch limited conditions, where prevailing along-shore easterly winds are blocked by headlands. This gives melt rates of 0.2-1.2 m d⁻¹ for Getz and 0.07-0.4 m d⁻¹ for Dotson.

Sea ice and ice front velocity

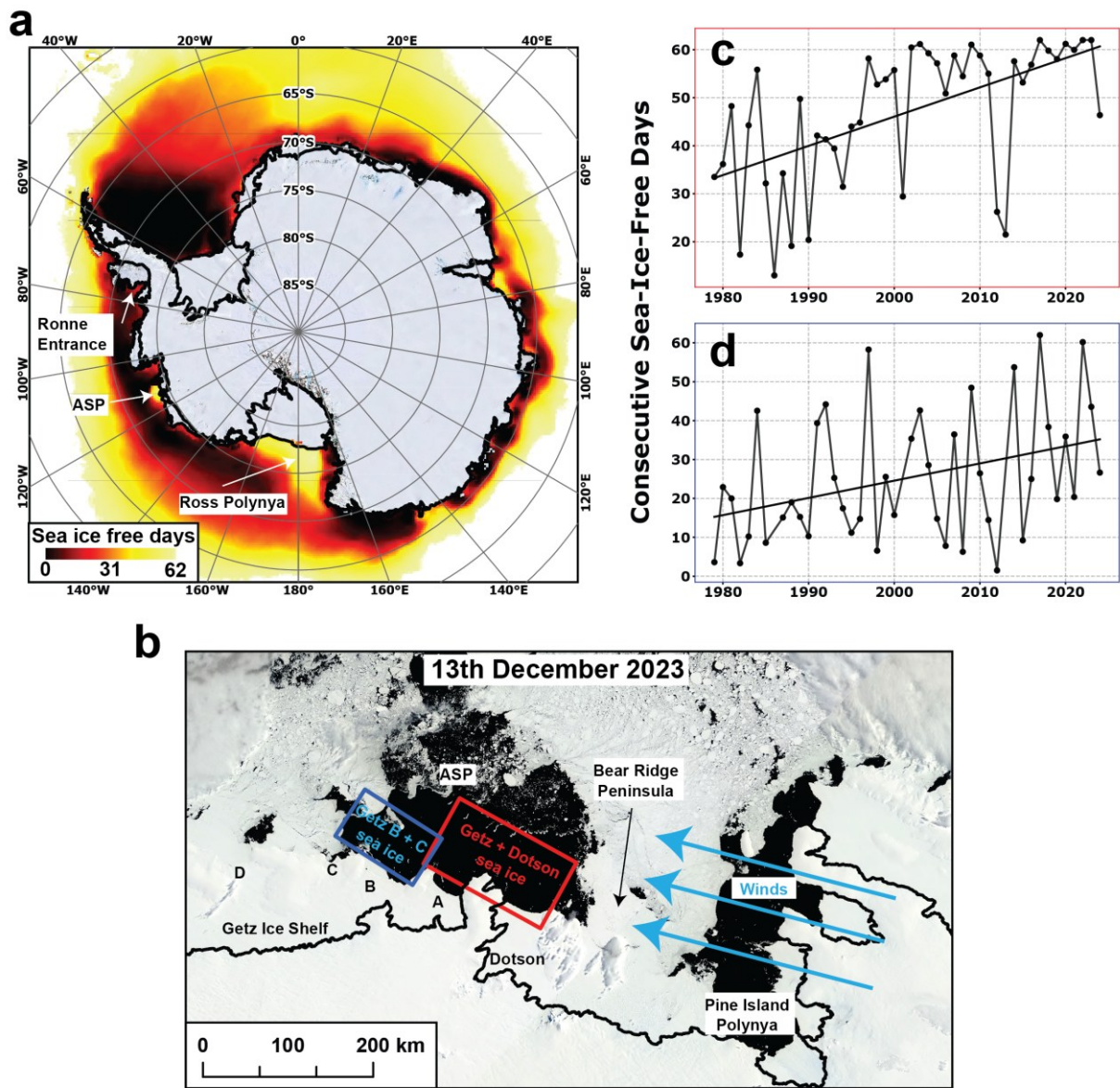
Gridded sea ice concentrations were obtained from passive microwave records spanning 26 October 1978 to 31 December 2024, utilizing the Nimbus-7 SMMR, DMSP-F8/-F11/-F13 SSM/I, and DMSP-F17 SSMIS sensors at a 25 km spatial resolution⁵⁵. Following established convention, individual pixels were classified as sea-ice-free when concentration fell below 15%⁶⁵. From this baseline, we quantified the number of consecutive sea-ice-free days during the core austral summer window (1 December to 31 January) for each year. This temporal window was selected because early-season sea ice retreat exposes open water directly to near-maximum solar insolation, which is required for meaningful surface water warming⁶⁶. We extracted the number of sea-ice-free days in front of each ice shelf section by calculating the mean pixel value within a bounding box extending approximately 50 km seaward from the ice front and spanning the width of the section. To calculate ice front velocity, we extracted the mean frontal speed for each ice shelf section using the MEaSURES ice velocity mosaic of Antarctica⁴³.

Calving and ice shelf thickness

We measure the maximum length and area for all calving events that were covered by Sentinel-1 imagery between November 2024 and June 2025 for the Getz A and B ice shelf outlets and extract mean ice shelf thickness across the ice front via BedMachine⁶⁷. We then compare these observations to theoretical estimates from elastic beam theory. The key assumption is that the foot-induced upward bending can be approximated as a homogeneous elastic floating beam that is subject to an upward shear force at the ice front (where the shear force is determined by the buoyancy of the foot). Calving is then estimated to occur at a given distance from the ice front where the bending is at a maximum and when the bending stress surpasses the yield strength of the ice. This framework has been shown to compare well to observed foot-induced bending and calving in a range of settings^{32,33,36}. To a first order approximation, the location of maximum bending stress (i.e. of maximum curvature) is located at a distance L_{calve} from the ice front, where $L_{\text{calve}} = \pi/(2\sqrt{2})l_w$ ^{32,33}. Here, l_w is the buoyancy wavelength which is given by:

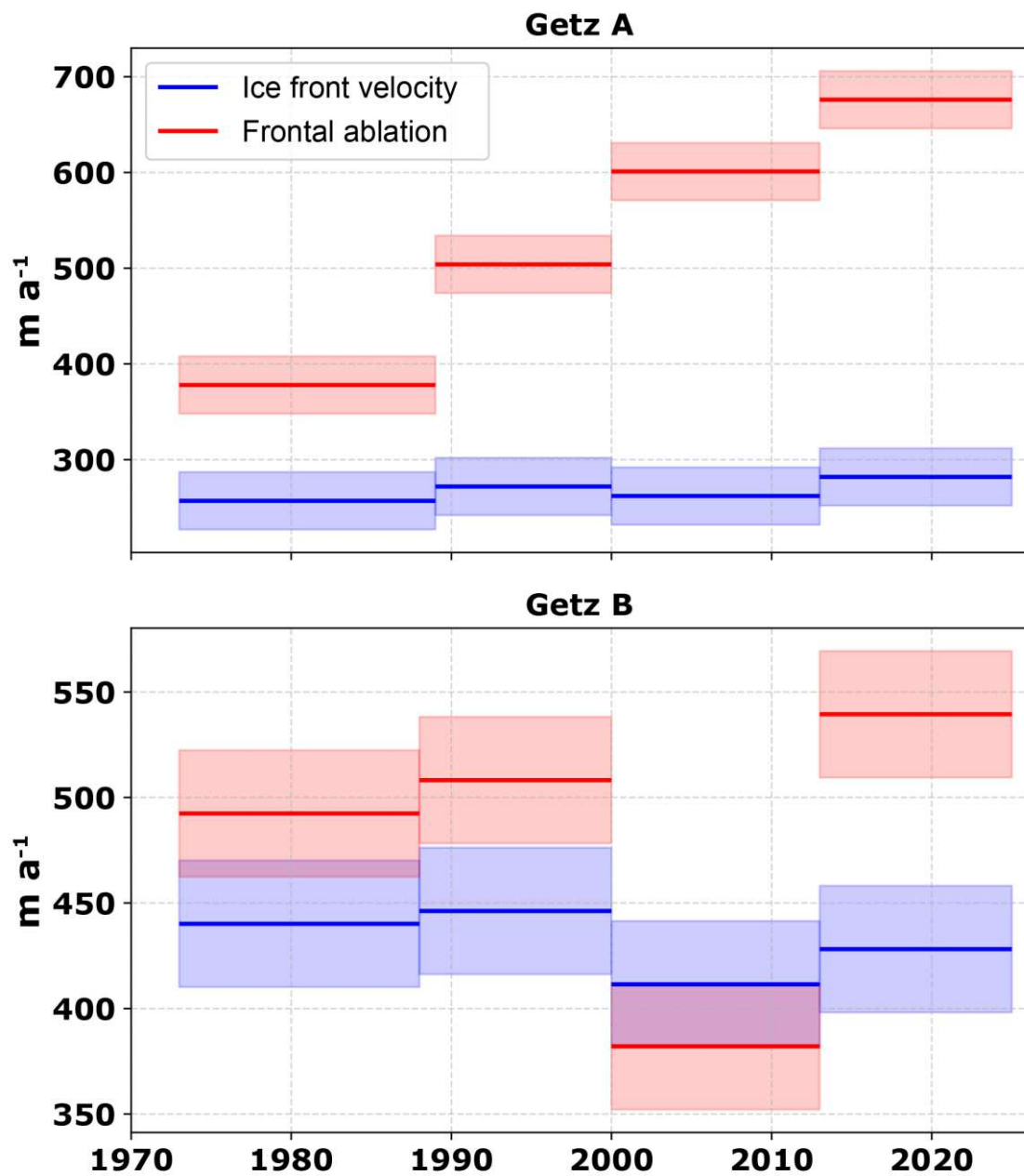
$$l_w = \left(\frac{Eh^3}{12(1 - \nu^2)g\rho_w} \right)^{1/4}$$

This describes the balance of bending stiffness of the beam versus restoring buoyancy of the water. Here, E is the elastic modulus, h the ice thickness, ν the Poisson ratio, and ρ_w the density of water. This entails that $L_{\text{calve}} \sim h^{3/4}$, in other words, the size of calving is expected to scale somewhat close to linearly with ice shelf frontal thickness (Extended Figure 8). The main unconstrained parameter is the elastic modulus E since the material properties of ice shelves are strongly impacted by crevasses, smaller-scale damages, and temperature effects, among other factors. Previous work^{13,68,69} has discussed how this may result in lowered “effective moduli” relative to that of pure ice (which is ~ 10 GPa). We assign the commonly used value from observational estimates of $E = 1$ GPa⁷⁰. Extended Figure 8 shows the strong agreement between our simulated calving widths and observations and illustrates the uncertainty in simulated calving widths arising from the assignment of the elastic modulus. Finally, we note that the elastic beam framework, combined with a simple yield strength criterion, suggests that the critical foot length that produces enough bending to drive calving scales with $h^{1/4}$, i.e., is only weakly dependent on the ice shelf thickness, as mentioned in the main text.

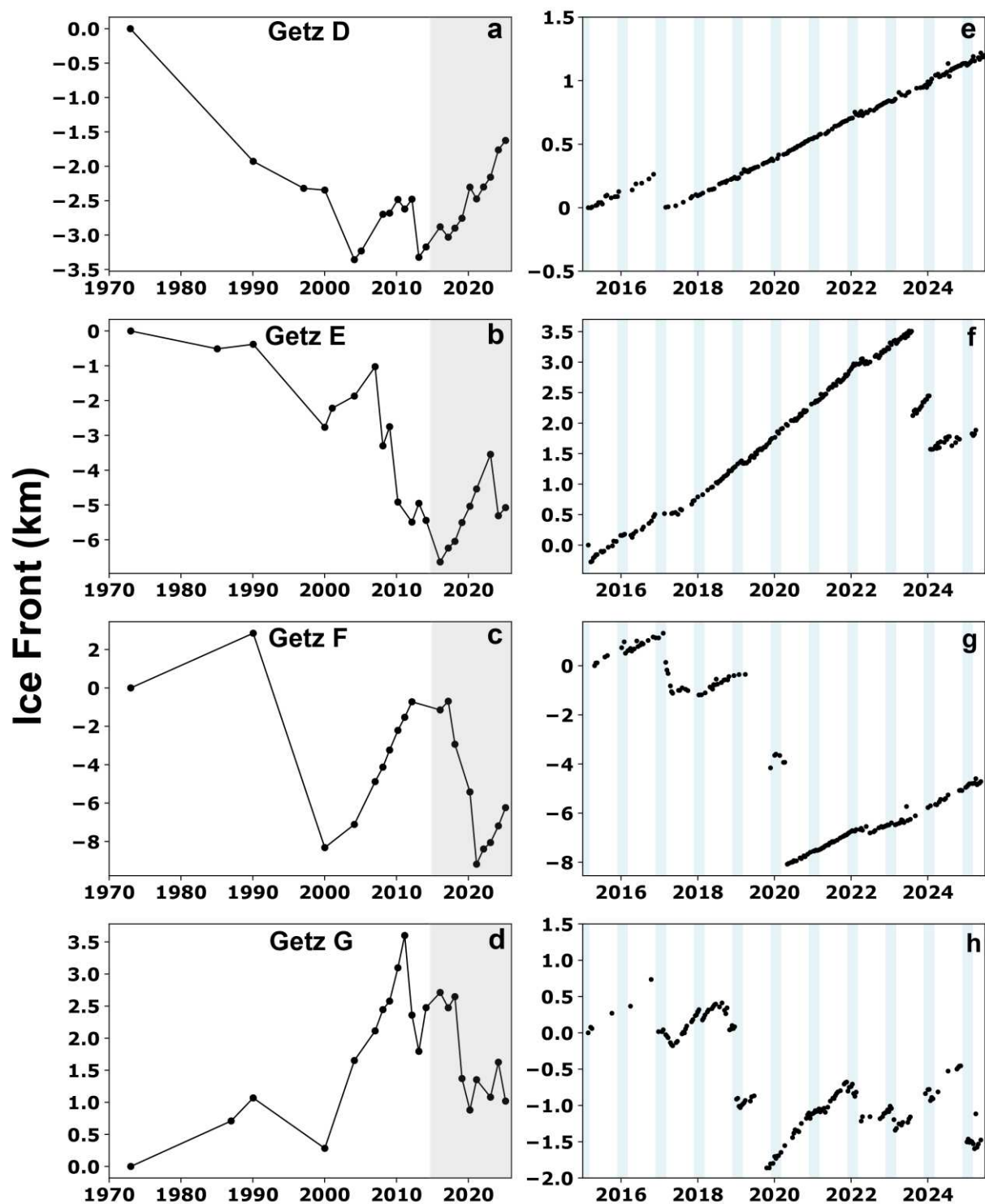


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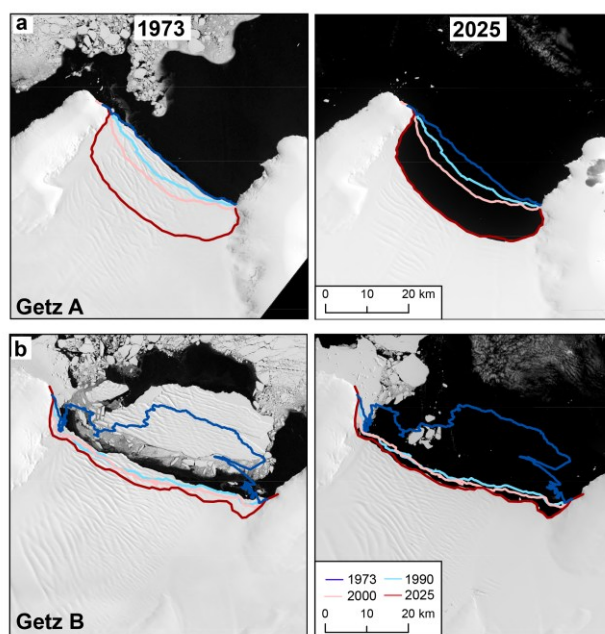
452 **Extended Data Figure 1:** a) Mean number of consecutive sea-ice-free days in December and
453 January across Antarctica 1979-2024⁵⁵. b) Schematic showing the wind direction over Bear
454 Ridge Peninsula that drives the formation of the ASP. The background MODIS image is from
455 NASA WorldView. The black line is the MODIS 2009 grounding line²⁴. c) Timeseries of
456 consecutive sea-ice-free days in December and January adjacent to the Getz A and Dotson ice
457 outlets (red box) with the black line showing the linear regression and d) As for c) but for the
458 Getz B and C ice shelf outlets (blue box).



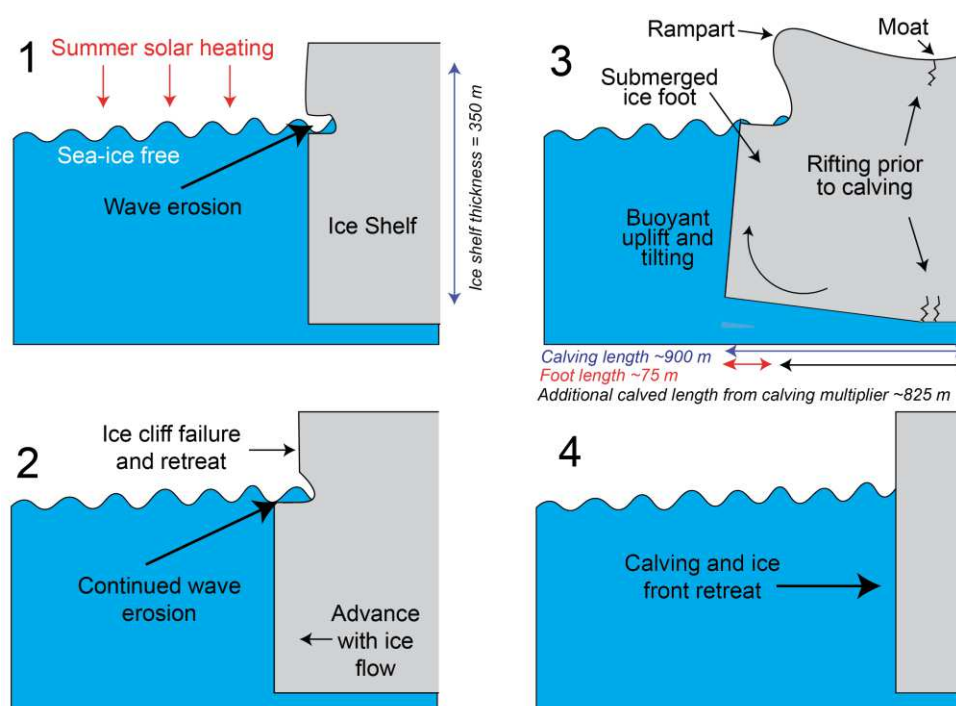
Extended Data Figure 2: Frontal ablation rates and frontal advance rates for the Getz A and Getz B ice shelf outlets.



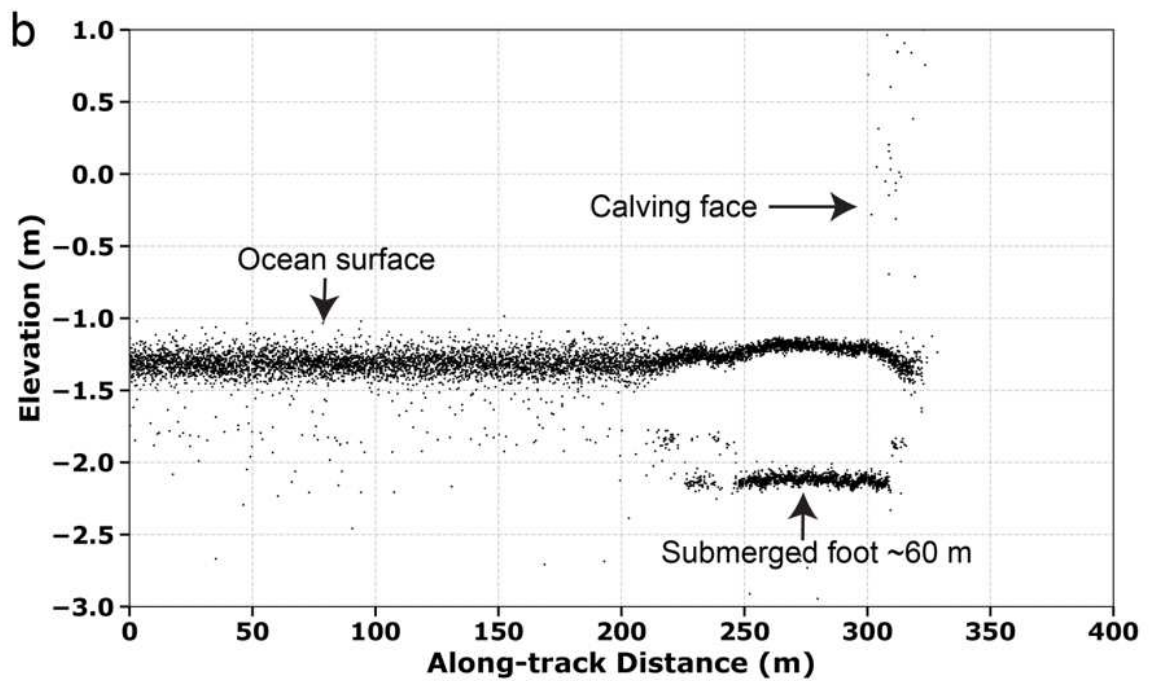
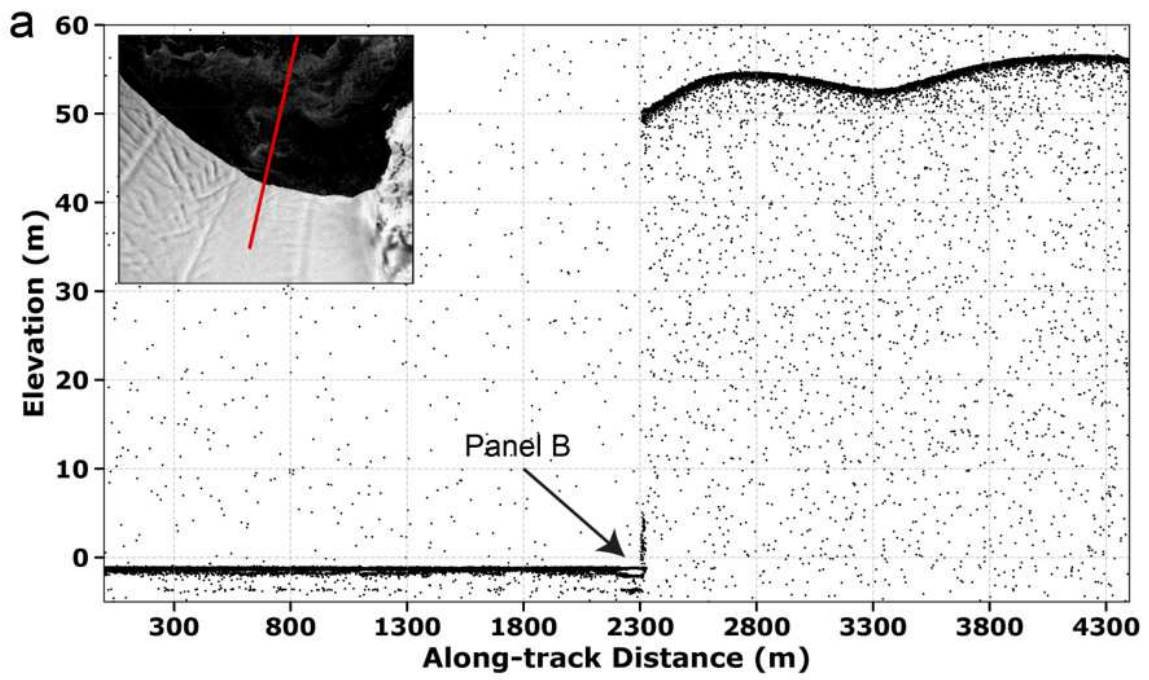
Extended Data Figure 3: Timeseries of ice-front change for Getz D-G outlets. **a-d)** Annual ice-front change from 1972-2025 from Landsat and MODIS imagery. The grey shadings represent the period covered via Sentinel-1. **e-h)** Weekly observations of ice-front change from Sentinel-1 2014-2025. The blue shadings represent the austral summer months December – February.



Extended Data Figure 4: No change in ice-shelf structure is observed between 1973 and 2025, with ice-front outlines overlain for a) Getz A and b) Getz B ice-shelf outlets. The underlying Landsat images are courtesy of the U.S. Geological Survey.



Extended Data Figure 5: Schematic of the footloose mechanism (not to scale). **1)** Seasonal loss of sea ice allows for heating of the ocean surface which in turn drives wave erosion at the waterline. **2)** Notching leads to ice cliff failure and retreat. **3)** Retreat of the ice cliff forms a buoyant underwater foot that lifts up and tilts forming rampart-moat structures, as picked up on ICESat-2 profiles (Fig. 3). **4)** The ice shelf calves a 900 m long iceberg once the foot reaches a critical threshold of around 75 m, assuming ice shelf thickness of 350 m. Thus around 75 m of erosion at the waterline drives a 900 m calving event. This is termed the calving multiplier effect.

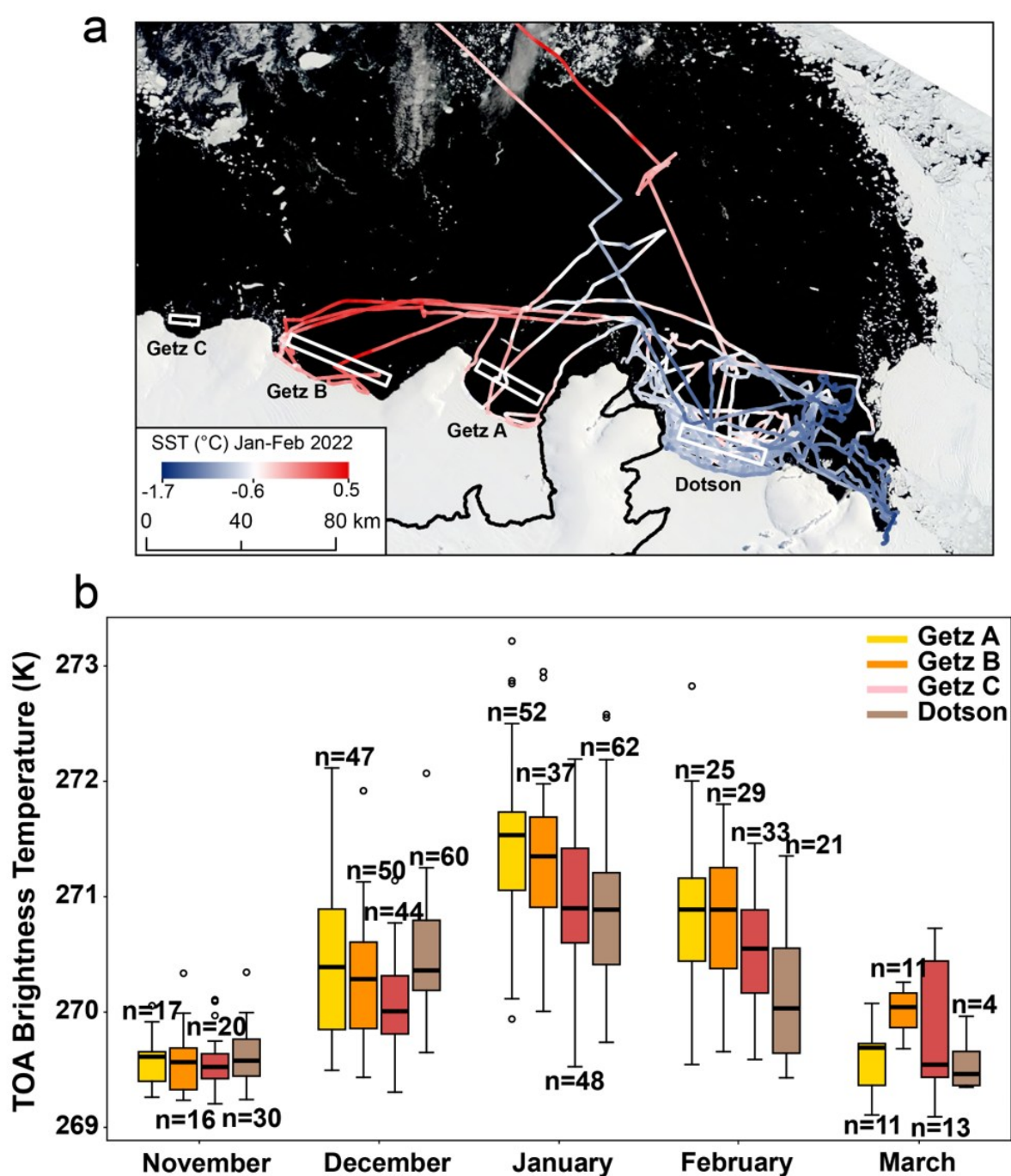


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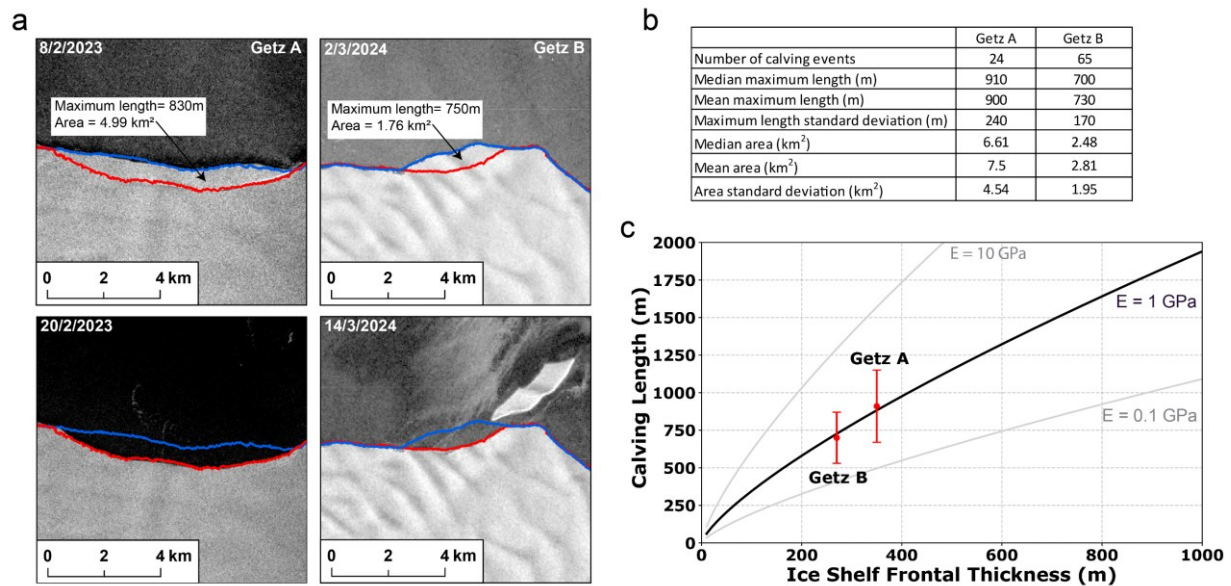
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484 **Extended Data Figure 6:** ICESat-2 observations of the submerged ice foot at Getz A. **a)** ATL03
 485 RGT 485 beam GR2R profile from 22nd October 2022 showing the ice front. **b)** Zoomed in
 486 profile at the waterline showing the submerged, 60 m long ice foot.

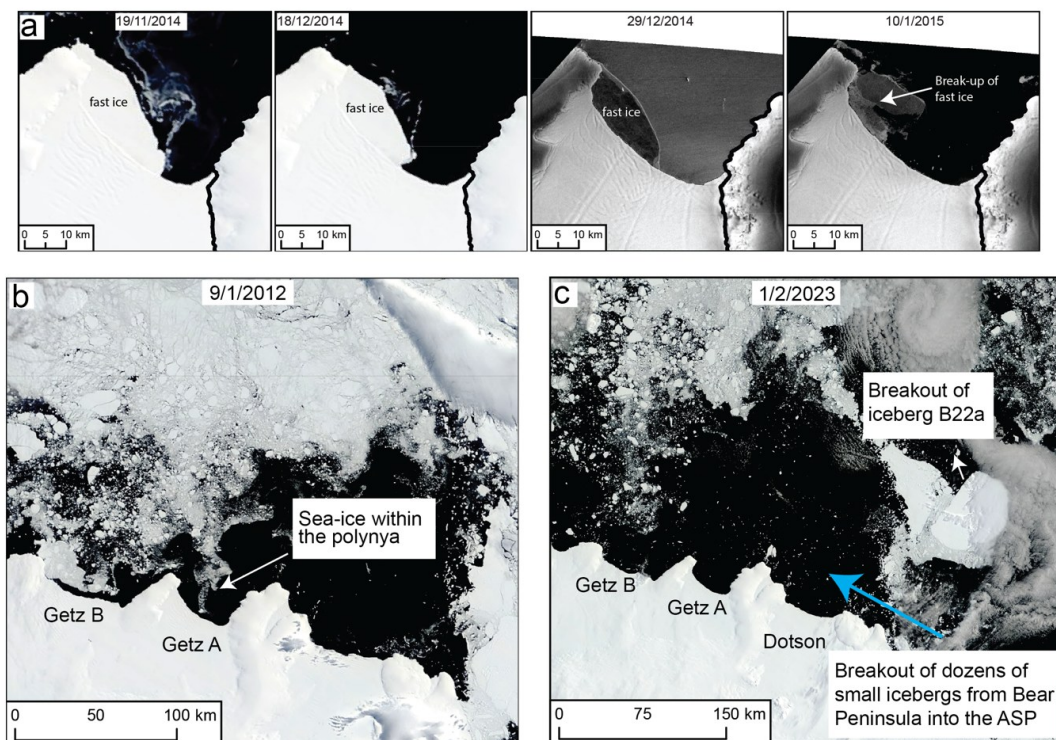
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Extended Data Figure 7: SST characteristics in the ASP. a) Observed SSTs from the Nathaniel B. Palmer NBP22-02 research cruise in January-February 2022⁴¹. The background MODIS image is from NASA WorldView. **b)** Mean monthly top of the atmosphere (TOA) Landsat 8 and 9 brightness temperatures from 2013-2025 extracted from the white boxes in panel a adjacent to each ice shelf. Top of the atmosphere brightness temperature is a proxy for the differences in SSTs. Note the cooler temperatures extracted adjacent to the Dotson Ice Shelf.



Extended Data Figure 8: Calving statistics for Getz A and Getz B 2014-2025. **a)** Sentinel-1 imagery showing example calving events from Getz A and Getz B. **b)** Calving statistics for each ice-shelf outlet. **c)** Predicted calving length in relation to ice shelf frontal thickness via the footloose mechanism for the elastic modulus, E , ranging from 0.1-10 GPa^{32,33}. The red points show strong agreement between predicted and observed calving length for $E = 1$ GPa.



Extended Data Figure 9: Anomalous sea ice conditions adjacent to ice fronts explaining the years where Getz A does not retreat. **a)** Landfast sea-ice persisted late into the summer of 2014/2015, before breaking up around 10th January, 2015. **b)** The infiltration of sea ice into the polynya during January 2012. **c)** The breakout of fast ice and icebergs from Bear Peninsula in January/February 2023 resulted in dozens of small icebergs drifting into the ASP and likely cooling SSTs.

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