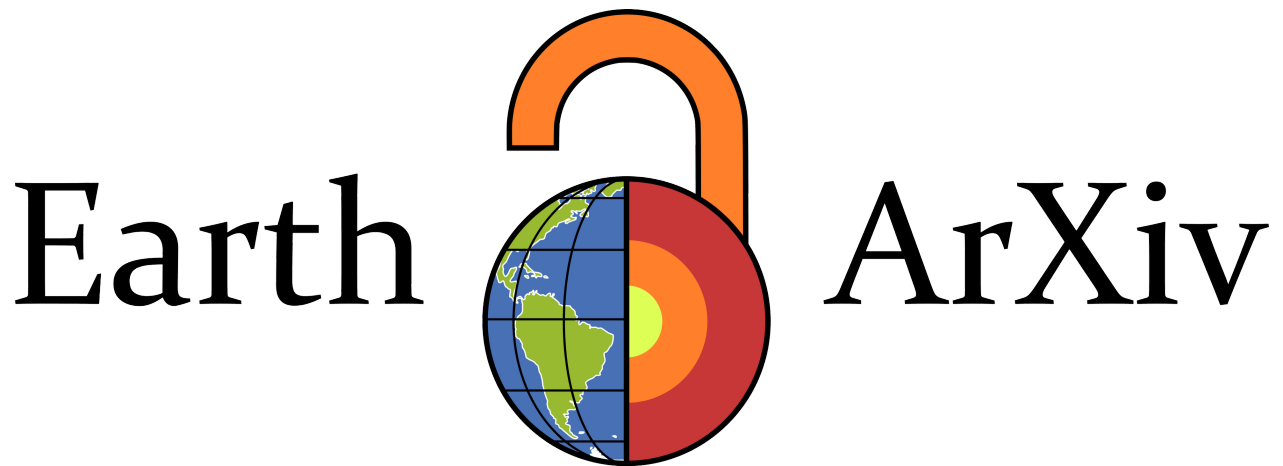




This is a non-peer-reviewed preprint submitted to EarthArXiv.

This manuscript has been submitted for publication in PNAS journal. Please note the manuscript has yet to be formally accepted for publication. Subsequent versions of this manuscript may have slightly different content. If accepted, the final version of this manuscript will be available via the 'Peer-reviewed Publication DOI' link on the right-hand side of this webpage.

Ocean-driven reactivation of basal melt and dynamic instability beneath Pine Island Glacier ice shelf since 2018

Lucille Gimenes^{a,1}, Romain Millan^{a,1}, Jean-Baptiste Barré^a, Amandine Clement^a, Jonas K. Andersen^b, Maud Bernat^d, Kaian Shahateet^{a-c}, and Antoine Rabatel^a

This manuscript was compiled on June 10, 2026

Pine Island Glacier (PIG) is Antarctica's largest contributor to sea-level rise and a major source of mass loss from the West Antarctic Ice Sheet. Its ice shelf is vulnerable to incursions of warm Circumpolar Deep Water (CDW) and has experienced rapid grounding-line retreat, flow acceleration, and increased ice discharge since the 1990s. Here, we reconstruct a 20-year record of basal melt rates beneath the PIG ice shelf using high-resolution multi-sensor stereo-photogrammetric digital elevation models combined with precise laser altimetry measurements. Following a slowdown in basal melting between 2004 and 2010, we document a renewed intensification of melting within the Pine Island cavity since 2018, leading to cavity expansion and coinciding with a 0.5°C increase in ocean temperature below 400 m depth since 2019. Intense basal melting rates exceeding 50 m yr⁻¹ now affect all ice-ocean interfaces deeper than 400 m, representing 40% of the ice-shelf area, and exceed 200 m yr⁻¹ near the grounding line. The recent emergence of localized melt channels on the southern tributary coincides with flow acceleration, increased structural damage, and complete detachment from the main glacier trunk, leaving it largely unbuttressed. Together with the recent re-acceleration of grounding-line retreat and ice discharge, these observations indicate a renewed phase of dynamic instability and mass loss, with important implications for the future evolution of the West Antarctic Ice Sheet.

Antarctica | ocean warming | glaciology | remote sensing

Glaciers and polar ice sheets are the main contributors to sea level rise (1). Antarctica, whose ice volume represents a potential sea level equivalent of 58 meters (2), is one of the largest sources of uncertainty (3). This is closely linked to the future evolution of ice shelves, which exert a significant buttressing effect on upstream ice, thereby regulating the discharge of ice into the ocean. The weakening of these floating ice tongues is mainly driven by the advection of warm circumpolar deep waters (CDW) beneath the shelves, leading to increased basal melt and grounding line retreat, which directly impacts ice discharge into the ocean. In some cases, the complete collapse of ice shelves led to a drastic increase in ice discharge into the ocean, (*e.g.*, Larsen A-B (4, 5) in Antarctica, and Zachariæ Isstrøm in Greenland (6, 7)).

One of the most vulnerable regions is the Amundsen Sea sector, which gathers the majority of Antarctica's mass loss through the Pine Island and Thwaites glaciers (8). Specifically, the Pine Island Glacier (PIG) ice shelf flows at almost 5 km yr⁻¹ at its front, discharging more than 130 Gt of ice per year since 2008 (8). The ice shelf extending from PIG is also undergoing dramatic changes, with some of the highest basal melt rates on Earth, largely exceeding 100 m yr⁻¹ near the grounding line (7, 9–12). Between 1992 and 2011, the grounding line experienced a strong retreat of more than 30 km (13) (Fig. 1a), which was accompanied by a significant increase in ice discharge (13, 14), and enhanced damage (15, 16).

Significance

Pine Island Glacier is Antarctica's largest contributor to current sea-level rise and has changed significantly over the past three decades. Pine Island ice shelf restrains the inland ice flow but its behavior is highly sensitive to ocean-driven melting. Using two decades of high-resolution satellite observations, we document a renewed intensification of basal melt beneath the ice shelf since 2018, following a period of relative slowdown. High basal melt now affects all deep ocean-exposed regions, reaching several hundreds meters per year near the grounding line, among the highest basal melt rates quantified on Antarctic ice shelves. This intensification of basal melt coincides with recent ocean warming, grounding-line retreat, glacier flow acceleration and damage propagation, warning on a new phase of instability and mass losses.

Author affiliations: ^aUniv. Grenoble-Alpes, CNRS, IRD, INRAE, Grenoble-INP, Institut des Géosciences de l'Environnement, Grenoble, France; ^bDepartment of Geosciences and Natural Resource Management, University of Copenhagen, Copenhagen K, Denmark; ^cAeronautics Institute of Technology, São José dos Campos, Brazil; ^dUniversité de Toulouse, LEGOS (CNES/CNRS/IRD/UT3), Toulouse, France

L.G. and R.M. designed and performed the research. L.G., J.B., J.A., A.C., M.B. processed and analyzed data. All authors contributed in the paper writing.

The authors declare no competing interests.

²To whom correspondence should be addressed.
E-mail: lucille.gimenes@univ-grenoble-alpes.fr and romain.millan@univ-grenoble-alpes.fr

In-situ oceanic campaigns and modeling of the interaction between the ice and the ocean have revealed the particular sensitivity of PIG ice shelf to oceanic forcing, mainly due to its vulnerability to the inflow of warm CDW in the cavity beneath the ice shelf (17–19). Some studies showed that PIG likely underwent an irreversible retreat phase that started in the 1970s and persisting until the early 1990s (20). Tipping-point and hysteresis behavior characteristic of marine ice sheet instability (MISI) were also detected (21), suggesting the potential for self-sustained retreat under present-day forcing. Although basal melt rates quantified from satellite data appear to have declined or stabilized since the 2010s (9), a continuous monitoring of this sensitive sector is essential to understand past and future dynamic changes of the Antarctic ice sheet and sea level rise. Recent studies have indeed documented a retreat of PIG grounding line up to 7 km since 2011, using Sentinel-1 data acquired at daily intervals (22, 23). These recent changes raise the question of potential shifts in oceanic conditions and a further intensification of basal melt beneath the ice shelf.

Historically, melt rate estimates have heavily relied on radar and laser altimetry data collected since the 1990s (9–11). This led to major progresses in monitoring basal melt rates, by providing the first continent-wide and spatially distributed estimate (10), but also by enabling the assessment of ice shelves mass balance partitioning (10, 24). Altimetry provides highly precise elevation measurements (on the order of 0.1 m) with dense along-track sampling. However, ground tracks are still spaced by several kilometers, requiring across-track interpolation that can smooth local channelized melt features (25). Recent studies using ice-ocean simulations have highlighted the importance of resolving small-scale basal topography, showing that channelized ice-shelf geometries can increase area-averaged basal melt rates by up to 35% compared to smoothed geometries (26). These results suggest that incised basal channels can trap warm and saline CDW multiplying melt rates by up to an order of magnitude within the channels themselves (26).

Since the early 2000s, numerous satellite optical sensors have captured stereoscopic image pairs over the polar ice sheets. Among these are data from the ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) sensor on NASA's TERRA satellite, as well as acquisitions from the SPOT (Satellites Pour l'Observation de la Terre) satellites operated by CNES/Airbus, some of which were collected during the 2008 International Polar Year (27). Since 2009, the Reference Elevation Model of Antarctica (REMA) program has provided comprehensive coverage of the Antarctic Ice Sheet, delivering time-stamped Digital Elevation Models (DEMs) at 2-m resolution (28). Although optical satellites are sensitive to cloud cover and do not operate during the polar night, these multi-sensor high-resolution DEMs, combined with precise laser altimetry measurements, open the door to an extended and high-resolution mapping of basal melt rates beneath ice shelves during the last 26 years. Existing studies using stereo imagery revealed substantial differences in spatial patterns of basal melting compared to continental-scale altimetry products (25, 29), demonstrating the potential of these datasets to resolve fine-scale basal processes. However, their diversity and temporal depth have yet

to be fully exploited to investigate basal melt variability over the past decades.

In this study, we provide a comprehensive assessment of the evolution of the PIG ice shelf over the last two decades. Using multi-sensor satellite observations, we reconstruct a new high-resolution time series of basal melt rates and investigate their spatial and temporal variability. We further examine concurrent changes in ice-shelf cavity geometry, ice dynamics, grounding-line migration, and damage, and place these observations in the context of evolving ocean conditions and subglacial hydrological activity.

Data and Methods

We combined multi-source satellite and airborne datasets to reconstruct high-resolution elevation changes and basal melt rates over the past two decades. We used REMA DEMs (28) from 2010–2023, which were derived from Worldview (Maxar) stereo-imagery at 2-m resolution (Fig. 1g–h). We also exploited ASTER stereo imagery to generate DEMs over the period 2002–2025 at a resolution of 30 m, using the MMASTER pipeline for both stereo-processing and jitter correction (7, 30) (more details in Material and Methods) (Fig. 1c–d). We complemented our analysis with DEMs derived from SPOT-5 High-Resolution Stereoscopic (HRS) instrument (CNES/Airbus) over 2007–2009 generated at 20 m resolution (31) (Fig. 1e–f). As part of the ERC project IceDaM (32), a Pléiades (CNES/Airbus) acquisition plan targeted more than 70% of PIG ice shelf (Fig. 1i–j) in 2024–2025 (Fig 1). Following the shutdown of the REMA service in this period, this acquisition plan provided a timely means to extend the time series of high-resolution DEMs (generated at 4-m resolution) over PIG ice shelf (see Material and Methods).

Laser altimetry datasets from Operation IceBridge (ATM, LVIS) and ICESat missions were used to generate reference DEMs, after applying corrections for geoid, tides, atmospheric pressure, and ocean dynamic topography (Fig. 1b & S1). Annual summer elevation mosaics were then produced through a two-step co-registration workflow in which DEM strips acquired during the selected year were vertically and horizontally aligned to the altimetry-based reference DEM using overlap-based selection criteria (see Material and Methods). The co-registered DEMs were merged iteratively into the final mosaic, with feathering applied along tile boundaries to minimize edge artifacts and ensure spatial consistency. Remaining gaps and areas of limited coverage were supplemented with altimetry and complementary multi-sensor DEM data to maximize spatial coverage (see Fig. S2).

We calculated surface flow velocities between DEM pairs and combined them with post-processed estimates from the MEaSUREs project (33) to obtain continuous and comprehensive yearly flow fields, enabling Lagrangian tracking of ice parcels and ice flux divergence estimation (Fig. S3, Material and Methods). Basal melt rates were then computed using a mass conservation framework, where the ice thickness (inferred from surface elevation under the hydrostatic equilibrium assumption), Surface Mass Balance (SMB) and Firn Air Content (FAC) estimates were combined with elevation changes and ice flux divergence along flow paths. We used SMB estimates

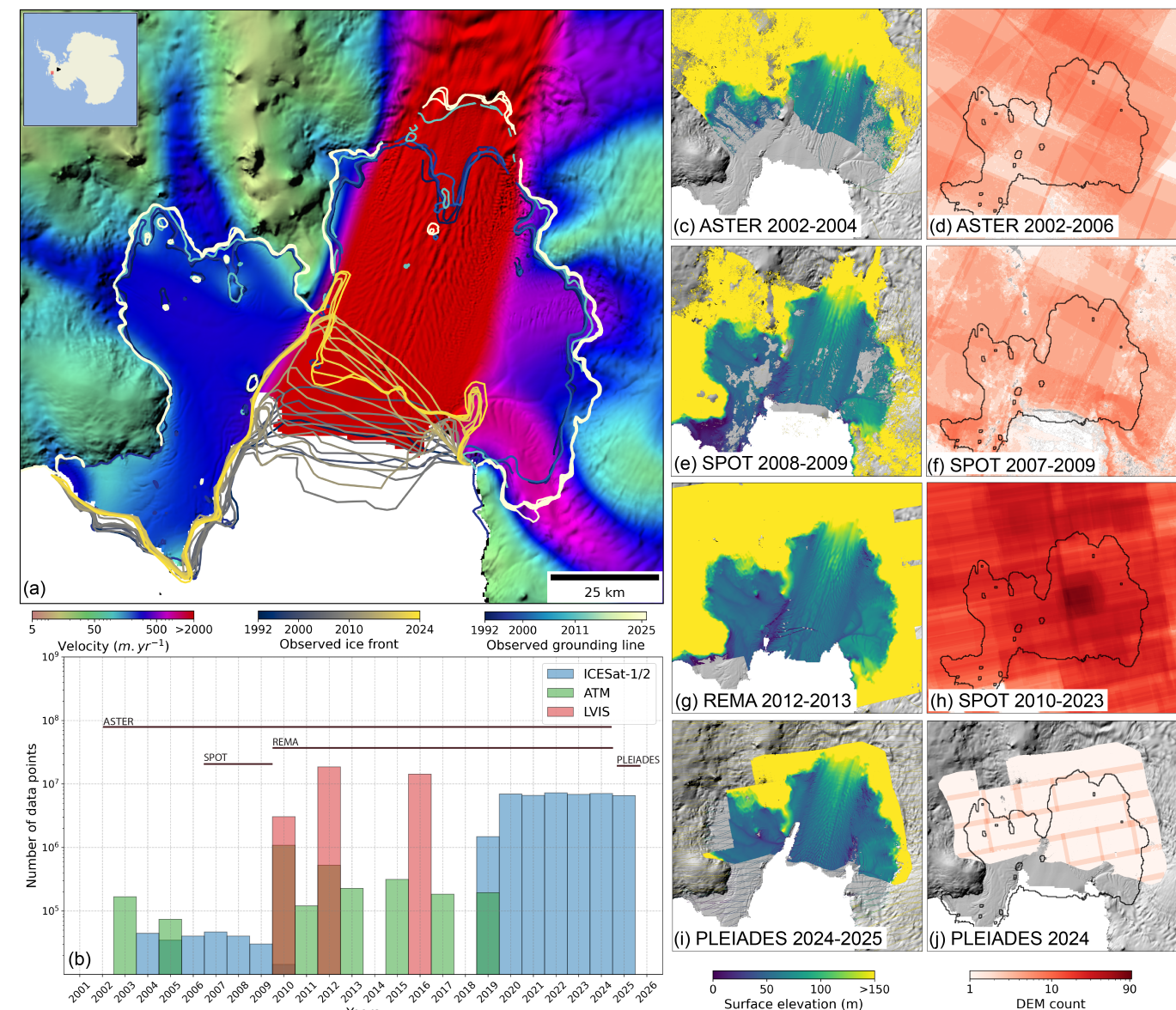


Fig. 1. (a) Average ice velocity during 2000–2024 overlaid on the 200-m x 200-m REMA mosaic, with observed grounding lines and ice fronts (1992–2025), **(b)** histogram of available laser altimetry points on PIG ice shelf, along with the timeline of used DEMs product sources. The other panels **(c–j)** show examples of annual surface elevation mosaics and DEM strip count for different elevation sources: **(c–d)** ASTER DEMs, **(e–f)** SPOT5 DEMs products, **(g–h)** REMA DEMs and **(i–j)** Pléiades DEMs.

from the Glacier Energy and Mass Balance (GEMB) model (34) (Fig. S4). We explore uncertainties in FAC(35) using an ensemble of three FAC products from different regional climate models: the GEMB (34) model, the Institute for Marine and Atmospheric research Utrecht Firn Densification Model (IMAU-FDM v1.2A, (36)) and the Goddard Space Flight Center firn densification model (GSFC, (37)). We calculated melting rates for all three FAC estimates and averaged them to provide an ensemble time-varying product (more details can be found in Material and Methods).

Multi-sensor quadruple-difference interferograms spanning 1991–2025 were used to delineate the flexure zone of PIG ice shelf, where the ice is not expected to be in full hydrostatic equilibrium. Anomalous refreezing patterns are removed and resulting basal melt estimates are spatially smoothed in this flexural zone (Fig. S5; Materials and Methods), where ice thickness estimates

may otherwise be significantly over- or underestimated (38). Basal melt rates were calculated across multiple time intervals and stacked to produce temporally resolved mosaics, providing a comprehensive assessment of spatial and temporal variability in basal melting (see Material and Methods).

Finally, we analyzed the three-dimensional evolution of basal melting and geometrical changes within the PIG ice shelf cavity and placed these results in the context of the glacier's dynamical evolution using a three-decade record of grounding-line migration and ice-flow velocity(22, 23, 33). We further interpreted these changes in relation to the evolution of ocean conditions using a dense compilation of Conductivity–Temperature–Depth (CTD) measurements since the 1990s, along with glacial hydrological activities from InSAR and optical image classification (see Materials and Methods).

Results

Spatial distribution of basal melt rates. The basal melt rates inferred on PIG ice shelf are among the highest recorded in Antarctica, as shown in the median mosaic product (Figure 2a) computed using all melt-rates in 2004–2024. The spatially integrated mass losses due to basal melt over this period reaches 135 ± 33 Gt yr⁻¹. FAC estimates from the three models exhibit discrepancies of up to 30% (Fig. S6b), but resulting variations in basal melt rates remain relatively modest and below 5% (Fig. S6d), which is within the intrinsic uncertainties of DEMs and velocity observations (see also Tab. S1). Locally, these variations can reach 30 m yr⁻¹ near the ice front, shear margins, and grounding line (Fig. S6c), which are zones where velocity divergence is the highest (Fig. S3f–g).

In the vicinity of the grounding line along the fast flowing trunk, values of basal melt rates exceed 100 m yr⁻¹ and can reach up to 250 m yr⁻¹ where the ice thickness is >1000 m (Fig. 2e). In this region, the ice melts along two main hot spots with melt rates >150 m yr⁻¹ which are located on both flanks of the grounding line (features 1 & 2 in Fig. 2a). Basal melt remains extremely high—typically above 50 m yr⁻¹—even 30 km downstream from the grounding line. Two melt channels also seem to stand out in the central part of the ice shelf (Fig. 2a, feature 3), along with a melt hot spot near the southernmost part of the ice front, where rates consistently exceed 30 m yr⁻¹ (feature 4). Another spot of high melt is visible along the northern shear margin (~60 m yr⁻¹, feature 5). These two hot spots are located in regions where the ice shelf completely disintegrated, and where thickness changes were therefore maximal.

In contrast, the rest of the ice shelf experiences much lower basal melt rates. The northern part is thinner (~200–400 m) and shows weaker basal melt, rarely exceeding 5 m yr⁻¹ (see Fig. 2a), which is close to ice thickness changes from flux divergence (Fig. S3g). The southern tributary also exhibits low basal melt rates on average (0–10 m yr⁻¹), but with local exceptions reaching substantially higher values (~60–80 m yr⁻¹). This is particularly true close to the grounding line in the westernmost part (>100 m yr⁻¹) where the ice shelf is thicker (~600–1000 m) and a distinct basal melt channel is also located (~60 m yr⁻¹, feature 6).

We compare our melt rates spatial distribution with five published datasets. Three are based on radar altimetry at resolutions of 250 m (12) (LOW25), 1 km (24) (DAV23) and 2 km (9) (PAO23). We also include products based on stereo imagery and altimetry at 250 m (29) (SHE19) and 50 m (25) (ZIN26) (Fig. S7). At the ice-shelf scale, all products reveal similar large-scale patterns of basal melting, providing confidence in the overall spatial distribution of melt rates. However, our 50 m estimates resolve substantially finer spatial structure and generally exhibit higher melt-rate magnitudes, particularly near the grounding line. In this region, the coarse-resolution altimetry products of PAO23 and DAV23 underestimate basal melt rates by up to 90%, while LOW25 reproduces the main melt hot spots more accurately but still underestimates peak values by approximately 30%. Away from the grounding zone, the high spatial resolution of stereo-photogrammetric DEMs enables the detection of narrow basal channels that are either smoothed or altered

by low signal-to-noise ratio in radar-altimetry products, and more specifically in yearly mosaic (e.g. Fig S8). Overall, our results are most consistent with previous stereo-photogrammetry-based estimates (25, 29), both in terms of spatial patterns and melt-rate magnitude (see detailed comparison in Supplementary Material). When integrated over the ice shelf, our basal melt estimates for 2010–2017 are broadly consistent with previous datasets (Tab. S2). Although PAO23 generally shows lower basal melt rates, this is compensated by locally enhanced melting near the ice front and along the southern shear margin, resulting in the same amount of basal melt (62 Gt yr⁻¹). LOW25 exhibits overall higher basal melt rates (22 % higher integrated melt), except near the grounding line. The higher integrated basal melt in ZIN26 (97 Gt yr⁻¹) likely results from the advection of high basal melt rates close to the grounding line.

Evolution of melt spatial pattern across periods. The analysis of basal melt rates over three sub-periods reveal major spatio-temporal changes beneath PIG ice shelf (Fig. 2b–d). From 2004–2010 to 2011–2017, basal melt weakened in the fast-flowing region near the grounding line, and melt rates exceeding 200 m yr⁻¹ became confined to a small portion of the southern sector (Fig. 2b–c). Over the same time period, rates in the central section of the grounding line decreased to <100 m yr⁻¹. Downstream, the main trunk of the ice shelf generally exhibits higher melt rates, with almost a 100% increase, which is especially pronounced in two channels located in the center (feature 3). Basal melt is about 30% weaker near the southernmost part of the ice front and along the northern shear margin. The south tributary also exhibits lower melt rates near its grounding line (~45 m yr⁻¹), except at its western tip where there is still a strong melting signal (~60–70 m yr⁻¹). During this time period, mass losses due to melting rates decreased from 118 ± 57 Gt yr⁻¹ (2004–2010) to 100 ± 26 Gt yr⁻¹ (2011–2017). Between 2011–2017 and 2018–2024, our results show a major intensification and a spatial expansion of basal melt near the grounding line, with intense melt extending across nearly its entire width (>200 m yr⁻¹). Additionally, basal melt rates a few kilometers downstream in the central part of the grounding zone also increased and became superior to 100 m yr⁻¹. A broad band of high basal melt rates (>60 m yr⁻¹), approximately 6 km wide, extends continuously along the main trunk of the ice shelf from the grounding line to the calving front. High basal melt rates are also present during this last period close to the shear margins where important thickness changes are occurring, partly due to enhanced ice shelf damage (15). It is worth noting that during 2018–2024, the main trunk detached from PIG northern ice shelf after 2016, and also recently from its southern part in 2026 (Fig. S9). A fork-shaped basal melt channel (~10–20 m yr⁻¹) becomes clearly visible on the northern shelf (feature 7 in Fig. 2d & Fig. S10a–b), signs of melting intensification in this region. The ice shelf's southern part is also subject to higher levels of basal melt, with a distinct basal melt channel in the west, where the grounding line has also retreated by a few kilometers compared to the early 2000s (Fig. 2, feature 6). During that time, mass losses from ice shelf integrated basal melt increased by 45% relative to 2011–2017 and reached 145 ± 22 Gt yr⁻¹ (2018–2024).

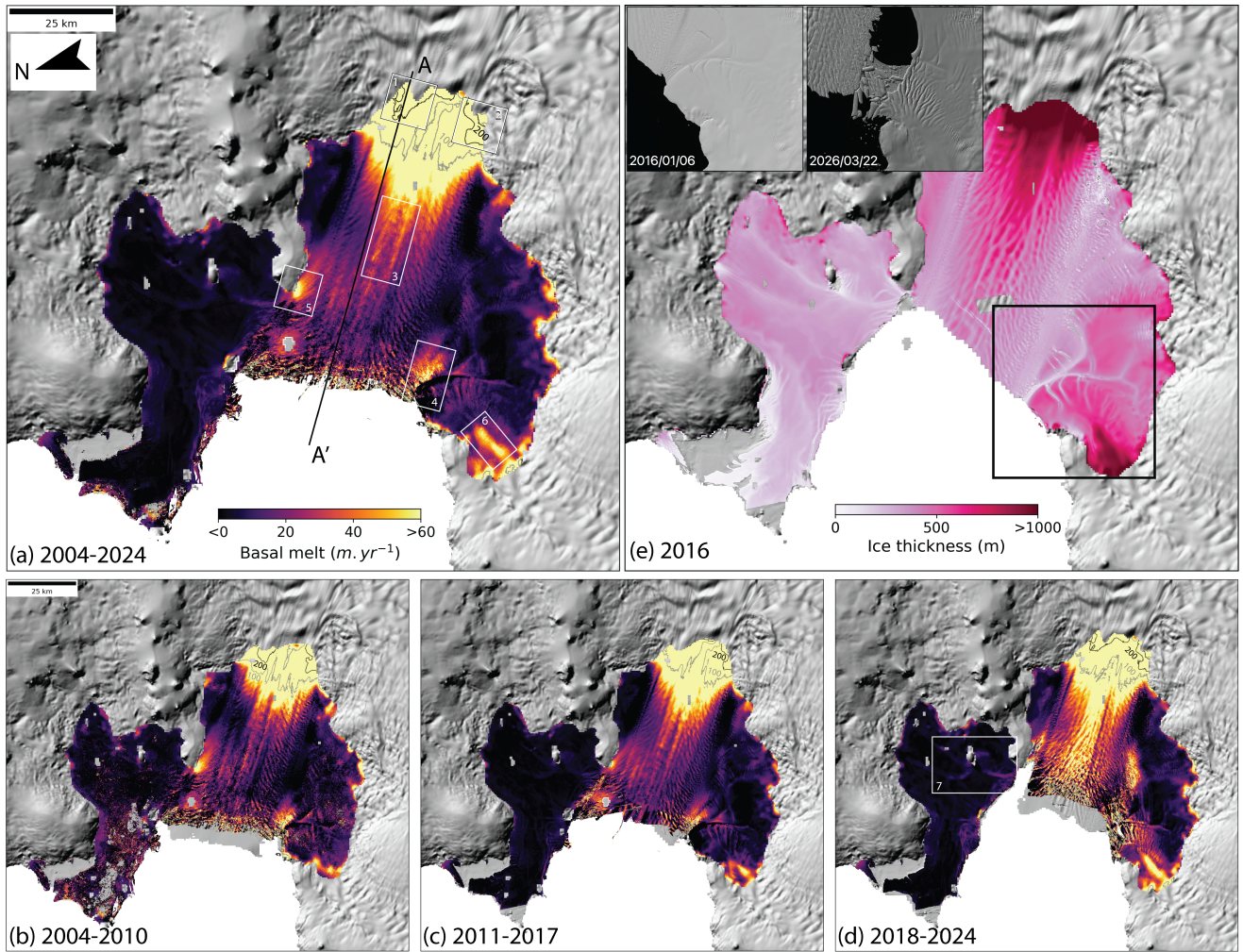


Fig. 2. (a) Median of all computed basal melt-rate products covering the 2004-2024 period, (b-d) median of basal melt-rates on the 2004-2010, 2011-2017, 2018-2024 period respectively. (e) Ice shelf thickness mosaic for the year 2016. Inset shows Landat-8 images illustrated the detachment of PIG southern tributary from the main ice shelf trunk. Discussed features are highlighted by a white frame on (a) and (d).

Basal melt spatio-temporal variability. We further investigate the detailed spatio-temporal evolution of melt rates at inter-annual resolution to identify the timing, persistence, and migration of major melt anomalies. We first analyzed the evolution along several profiles (see locations in Fig. S11, Fig. 3b, Fig. S12 & S13), which pass across different portions of the grounding line hot spots, on the main ice shelf trunk. Along the northernmost profile (AA'), due to grounding line retreat in the late 2000s, a ~4-km portion of the ice base is prone to melt rates higher than 200 m yr^{-1} , but this portion diminishes after 2010. Around 2014 and 2018, the rates are significantly lower than the previous years at the grounding line ($\sim 170 \text{ m yr}^{-1}$). After 2020 and the beginning of a new retreat episode, the distance over which melt rates exceed 200 m yr^{-1} expands to reach more than 10 km in 2024. Similarly, basal melt rates reaching 50 m yr^{-1} also seem to propagate downstream over time after 2007, almost reaching the ice front after 2020 (Fig. 3b), as also seen in Figure 2. Basal melt rate evolution of other profiles along the main trunk show similar trends at lower scales of melt near the grounding line (see Fig. S12 & S13).

Comparing time evolution of melting rates in a section 5-km away from the GL, time series from PAO23 and

LOW25 are broadly consistent with the trends observed in this study (Fig. S7b) over their respective periods. However, PAO23 captures the long-term trend but systematically underestimates melt rates by approximately 30%, whereas LOW25 agrees more closely with our mean estimates but exhibits substantially larger short-term variability.

The temporal trend is even clearer when looking at the distribution of basal melt with ice shelf draft and thus the depth of the ice-ocean interface, illustrated by basal melt rate contour lines on Fig. 3a. After a short decrease in basal melt at the beginning of the time series, levels for 50, 100, 150 and 200 m yr^{-1} basal melting rates are approximately located at 550, 725, 950 and 1050 m depth, respectively. For all ice shelf sectors with a draft depth greater than 700 m, basal melt decreases sharply after 2009, leaving only the thicker areas (thicknesses $>1200 \text{ m}$) prone to very high basal melt rates ($>200 \text{ m yr}^{-1}$). Already after 2012, ice shelf portions with a base located between 1000 and 1200 m depth are subjected to increasing melt rates. Meanwhile, the rest of the ice shelf (*i.e.* with a draft depth lower than 900 m) stabilizes at the same level of basal melt for approximately 5 years. After 2020, the rise in basal melting rate magnitude is observed

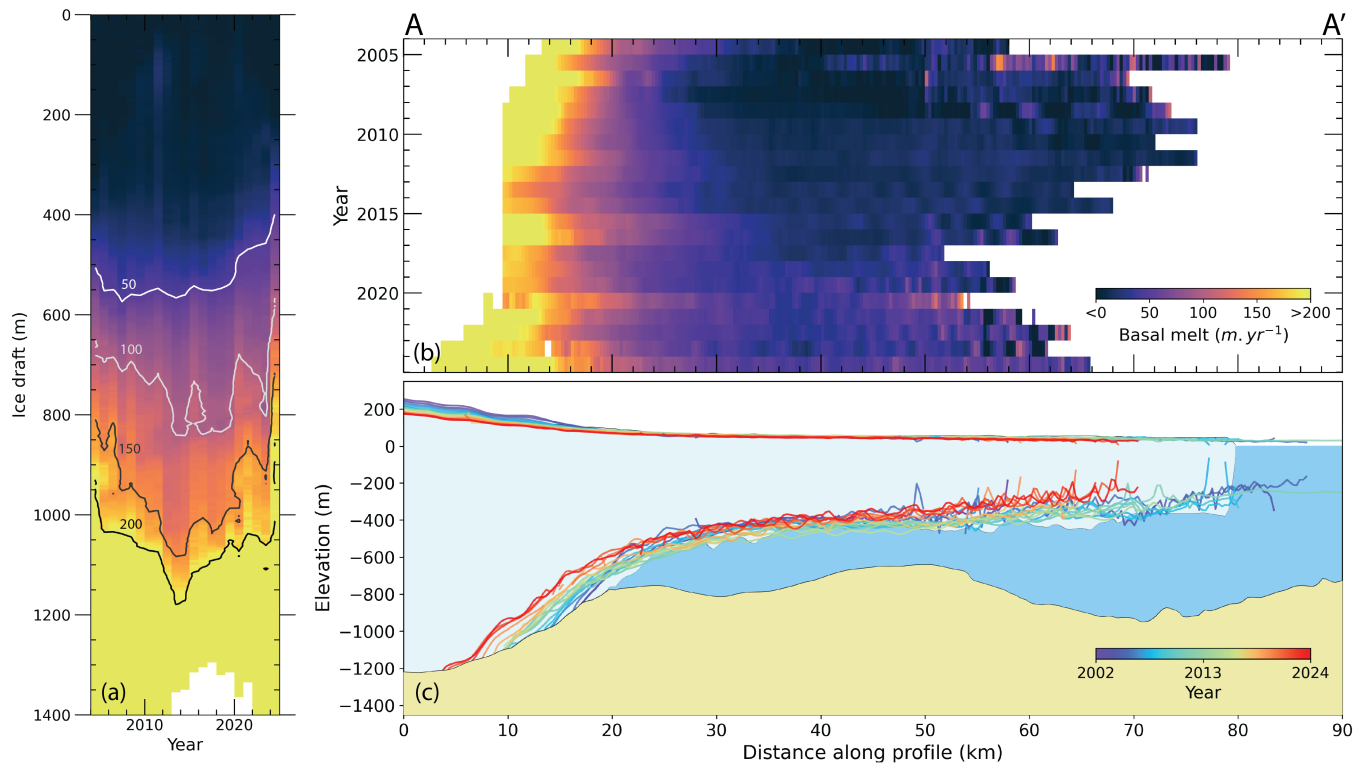


Fig. 3. (a) Hovmöller diagram of melt rate evolution as a function of binned ice shelf draft with contour lines plotted at 50, 100, 150 and 200 m yr⁻¹, (b) basal melt rate evolution averaged on a 2-km width along profile AA' displayed in Fig. 2, and (c) ice thicknesses evolution along the same profile (with background thicknesses and bedrock from BedMachine (2)).

but this time for all parts of the ice shelf with a draft thicker than 300 m, *i.e.* most of the floating ice (Fig. 3a). Changes in the distribution of basal melt with draft are mostly reflective of the evolution of the ice shelf's main trunk (Fig. S15) and are overall visible for all DEM pairs used for the basal melt computation (Fig. S16).

Cavity geometrical evolution. Analysis of ice-thickness profiles and temporal changes in ice-shelf draft reveals substantial geometric modifications of the PIG ice shelf cavity over the last two decades (Fig. 3c, Fig. S14). Grounding-line retreat across a predominantly retrograde bed has expanded the cavity inland, while sustained ice-shelf thinning has enlarged it vertically (Fig. 3c, Fig. S12 & S13). The newly formed cavity associated with recent grounding-line retreat (23) has an average water-column thickness of 89 m and contributes approximately 22 km³ of additional cavity volume. Between 2002–2010 and 2018–2024, the mean sub-ice-shelf water-column thickness increased by an average of 54 m across the entire ice shelf (due to thinning), corresponding to a total cavity-volume increase of about 315 km³, or 14% of the estimated cavity volume during 2002–2010. To investigate the spatial distribution of these changes, we defined a fixed deep-cavity sector using manually delineated polygons encompassing the principal deep regions of the cavity (Fig S14). Within this sector, the mean water-column thickness increased by approximately 107 m, nearly twice the increase observed over the remainder of the ice shelf (~48 m). Local increases in water-column thickness exceed 200 m, particularly in the vicinity of the grounding line and along deep, incised basal channels, highlighting the concentration of cavity growth within the deepest sectors

of the ice shelf, consistent with the distribution of the highest basal melt rates (Fig. 2, Fig S14).

Ocean changes. As CDW flows onto the continental shelf due to Ekman transport, it reaches the Amundsen Sea sector following deep troughs in the seabed bathymetry. The presence of modified Circumpolar Deep Water beneath PIG ice shelf is confirmed by oceanic records since the 1990s (39). To reach the ice shelf's grounding line, CDW must pass over a transverse ridge rising at 700 m depth below sea level (b.s.l.) (40, 41), which implies that it likely originates from around 400 m (the ice shelf thickness) to 700 m depth in the water column. The evolution of conductivity, temperature, and depth (CTD) data over the study period shows that between 2000 and 2009, mean ocean temperatures below 400 m depth increased from $0.98 \pm 0.04^\circ\text{C}$ to $1.14 \pm 0.02^\circ\text{C}$ (Fig. 4a–b). It is worth noting that in 2012, the temperature below 400 m was measured at a mean of approximately $(0.4 \pm 0.1)^\circ\text{C}$ in previous studies(17), which is 0.6°C colder than the temperature reported here in 2014 (Fig. 4b). This highlights an intermediate temperature peak between 2012 and 2014, before declining again through to 2019, when the temperature reached $(0.48 \pm 0.08)^\circ\text{C}$. A renewed warming was observed in 2020, with temperatures rising to $(0.8 \pm 0.02)^\circ\text{C}$ and further to $(1.05 \pm 0.03)^\circ\text{C}$ in 2024 and $(1.1 \pm 0.03)^\circ\text{C}$ in 2026.

Discussion

Using multi-sensor stereo-photogrammetric DEMs, we were able to reconstruct a two-decades long high-resolution time series of basal melting rates over PIG ice shelf,

revealing a recent widespread increase in basal melting rates below the ice shelf.

Uncertainties. Errors in basal melt rates are primarily driven by uncertainties on annual surface elevation DEMs. These are the highest for early 2000's mosaics as they rely exclusively on ASTER DEMs, where the image quality in this period led to lower SNR in the DEMs (NMAD: 3-4.5 m, Fig. S17). Errors in melt rates therefore reached about 30 m yr⁻¹ close to the grounding line, where the uncertainty is the highest. DEM uncertainties decrease for SPOT-derived mosaics (NMAD: 1.7-2.4 m) and further for REMA-based products (NMAD: 0.5-0.8 m), with resulting basal melt errors at the order 15-20 m yr⁻¹. Additionally, the availability of altimetry data, which can also be sparse for certain years (*e.g.*, 2010, 2012, see S1), directly affects vertical alignment and therefore DEM accuracy.

The grounding zone remains a particularly uncertain region, due to high basal melt rates and limitations in the hydrostatic equilibrium assumption. This can lead to artifacts such as apparent refreezing near the grounding line or ephemeral grounding points (38). Although our flexure-filtering approach does not provide a direct physical correction to the hydrostatic equilibrium assumption within the grounding zone, it introduces an efficient and observationally constrained framework that helps reduce uncertainties in regions where melt rates are most prone to biases (Fig. S18). Artifacts in satellite derived basal melt rates also arise from rapid changes in ice thickness due to damage or regular calving events. Ice velocity fields that are not perfectly matching the DEM time periods may also result in inaccurate advection estimates between epochs, producing artificially high basal melt rates in deep fractures. The use of DEM-to-DEM correlation velocities over 1-2 year intervals allows to partially reduce this effect. Areas such as the northern tributary are also especially uncertain: since basal melt rates are generally weak, the estimates approach the order of magnitude of errors in this region (~ 5 m yr⁻¹), and are close to the ice thickness changes from flux divergence. However, the distinct fork-shaped melting channel identified during the 2018-2024 period has also been previously detected by (25) (Fig. S10).

Comparison with variations in ocean conditions. The changes observed in basal melt rates are consistent with the evolution of ocean temperature throughout the water column below the ice shelf during the past two decades (Fig. 4). The ocean temperature at the intermediate depth (400-1500 m) peaked in 2007-2009 and again in 2024, when basal melt rates rose again to reach similar values as in 2005-2009 (Fig. 4c). Similarly, the decrease in basal melt rates starting in 2009 coincides with a period of decline in temperature after 2010. This is consistent with former studies indicating a cold anomaly between 2010 and 2012 (17) that actually lasted for several years (42) and that is attributed to changes in winds at the continental shelf's edge (17, 42). Likewise, the temperature minimum reported in 2012 by previous studies (17) coincides with the melt minimum mapped here (Fig 4). The subsequent temperature increase until 2014 is further reflected by an intermediate peak in melt rates at ice-ocean interfaces below 400 m over the same period (Fig 4). The recent increase in basal melt rates in regions where the ice

shelf has a draft of at least 400 m also coincides with a renewed warming at the corresponding ocean depths (400-1500 m, see Fig. 4b). Less documentation exists on this recent observed warming since 2019, however, oceanic measurements between 2020 and 2023 beneath Thwaites eastern ice shelf revealed a warming and a thickening of the modified CDW layer downstream of Pine Island Bay, which also corroborates our results (43). Delays in the onset of basal melt and ocean temperature increase over the 2013-2019 period may partly arise from geometric feed-backs associated with ice-shelf thinning: as the ice shelf melts, thicker and steeper slopes are exposed to warmer waters, which has the capacity to enhance basal melting even without changes in oceanic conditions (44). Observations of ice-shelf meltwater in the Amundsen Sea Embayment (ASE) further indicate that temporal variability in basal melting exhibits a nonlinear relationship with ocean temperature variability (18).

Sub- and supra-glacial lake drainage analysis. We also analyzed glacial hydrological activities, as freshwater release at depth has the potential to increase basal melt rates (45, 46). The analysis of subglacial lake drainage dataset from (47) and (48) derived from CryoSat-2 data revealed no active lakes drained in PIG catchment area throughout the study period. Nevertheless, smaller subglacial drainage events were detected from satellite radar interferometry (Fig S21), even if analysis in and near the grounding zone remain limited due to decorrelation in fast flowing regions (49). Most events are therefore located more than 100 km from the grounding zone. It is worth noting that these are smaller transient events, associated with centimetric-scale surface displacements and estimated water volumes <0.01 km³, whose potential to impact basal melting rates remain limited. We also used methods from (50) and (48) on Landsat-8 imagery to assess if surface lake drainage occurred near the grounding zone between 2013 and 2025. Only small recurrent lakes (<5 km², see Fig. S22) were detected on the northern tributary, in areas where thickness is typically <400 m, where freshwater inputs is less likely to interact with deep warm and saline CDW. Overall, these evidences suggest that sub- and supra-glacial hydrological activity must have played only a minor role in modulating basal melt rates between 2004 and 2024. Moreover, sudden subglacial lake drainage events are expected to produce a synchronous increase in basal melting on timescales of a few months (45). By contrast, the basal melt increase observed in this study occurs over longer timescales, spanning several years.

Ice shelf dynamical changes. The analysis of the spatio-temporal variability of the grounding line reveals a synchronicity with the evolution of basal melt rates and oceanic conditions. Following a retreat of several tens of kilometers between 1992 and 2011 along the entire central trunk of the glacier, InSAR observations reveal a marked slowdown in the grounding line retreat. The central sector of the grounding line stabilized over the 2011-2021 period, with retreat rates decreasing from 1.39 km yr⁻¹ to -0.05 km yr⁻¹ (Fig. 4c, Transect 2). The southern sector stabilized from 2015 onward, with slightly positive rates indicating a minor re-advance of the glacier between 2016 and 2021. In contrast, retreat in the northern sector

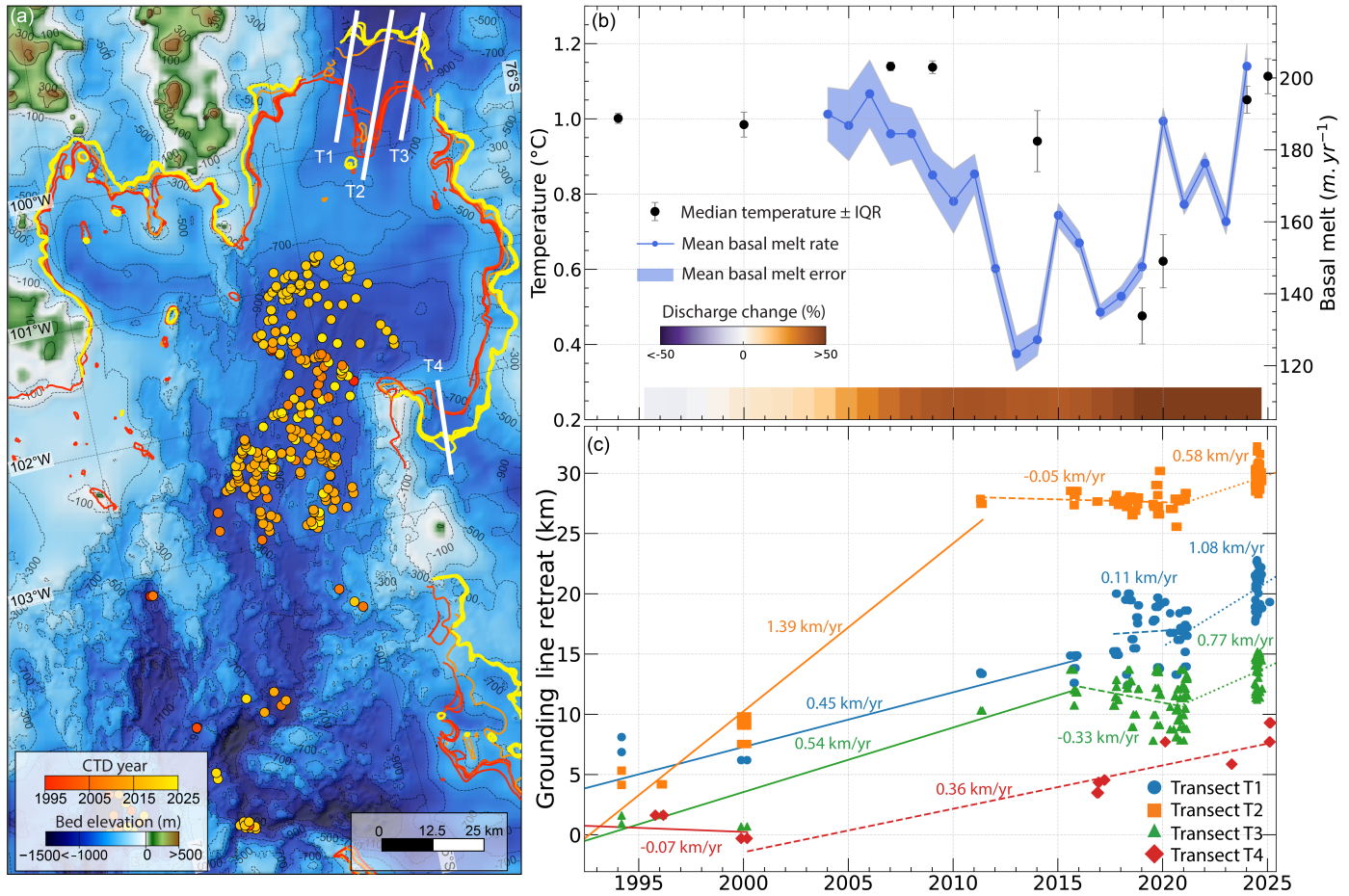


Fig. 4. Conductivity, temperature, depth measurements and ice dynamic changes at Pine Island Glacier ice shelf. (a) CTD location from 1990–2025 overlaid on IBCSO bathymetry with observed grounding lines displayed, (b) change in temperature averaged between 400 and 1500 m depth, as well as basal melt binned according to draft and averaged on the same depth interval. Ice discharge in PIG basin from (24) is also displayed, as the relative change compared to the 1996–2000 mean. (c) Grounding line changes from (23) in kilometers relative to 1992 along four transect (position in panel a). Linear fit indicates rates of retreat over sub-period.

remained more continuous and only began to slow down after 2017, with rates decreasing from 0.47 km yr^{-1} to 0.11 km yr^{-1} (Fig. 4c). These nearly synchronous pauses are consistent with the weakening of CDW intrusions and the corresponding reduction in basal melt rates during this period (Fig. 4b). After 2021, while basal melting had already resumed in response to increasing deep-ocean temperatures, the entire main grounding line of PIG retreated again. During this period, retreat rates increased to 0.58 km yr^{-1} , 1.08 km yr^{-1} , and 0.77 km yr^{-1} for the central, northern, and southern sectors, respectively. The retreat rate on the northernmost part of the grounding line (transect T1), is therefore now almost reaching the fastest retreat rate measured at PIG (1.39 km yr^{-1} in 1992–2011, Fig 4c).

Overall, the northern and southern flanks of the grounding line retreated more than its center (22), by approximately 5 km between 2021 and 2025, and correspond to the two previously identified hot spots of extremely high basal melt ($>200 \text{ m yr}^{-1}$, Fig. 2). This retreat occurred on a retrograde bedrock slope, which left the glacier grounded at 1200 m b.s.l. (2), exposing the ice shelf's base to more pronounced melt at the grounding line. Indeed, as the freezing point of sea water decreases with increasing pressure (depression of 0.75°C per kilometer of water depth (51)), the thermal forcing exerted by CDW

on the ice is stronger at depth (52). The strong coupling between ice shelf basal melting and glacier retreat has already been demonstrated in the ASE (20, 53, 54). We thus suggest that enhanced basal melt observed after around 2018 near the northern and southern flanks of the grounding line may have triggered its retreat in these locations. These melting-retreat processes are then further sustained in a positive feedback loop, as the grounding line position is inherently unstable on the retrograde bed (55).

A similar trend is observed at the grounding line of PIG southern tributary glacier, which also receded on average 2–4 km between the early 2000s and 2025 (22) (Fig 4a). The retreat was particularly pronounced in areas localized on a retrograde bed slope such as its westernmost point, where it reached 7 km (Fig 4c). This area coincides with the presence of highly channelized basal melt rates ($\sim 60 \pm 12 \text{ m yr}^{-1}$), which developed in recent years (Fig. 2e). Notably, the 10-km long channel corresponds to a bedrock trough, that is around 5 km-wide and plunges from 600 to 800 m depth b.s.l. going inland parallel to flow direction. This hollow is separated from the bedrock under the main ice shelf's trunk by a small ridge rising to around 600 m b.s.l. (visible in Fig. 4a). The recent thickening of the CDW layer (discussed in comparison with ocean conditions) may have allowed for the warm waters to

1019 overpass this ridge, in addition to the ice shelf thinning
1020 near the ice front at the junction between the main and
1021 the southern tributaries by enhanced melt. The intrusion
1022 of warmer CDW (see Results section) in this large trough
1023 leads to the formation of a basal channel that may trigger
1024 grounding line retreat. Neighboring spots of high basal
1025 melt very close to the grounding line also correspond to
1026 zones of observed grounding line retreat in recent years.

1027 The temporal evolution of basal melt is accompanied
1028 by changes in glacier flow speed. Indeed, after a strong
1029 increase in ice velocity since the 2000's, the latter remained
1030 around 3.9 km yr^{-1} at the grounding line position for the
1031 2012-2017 time period (see Fig. S3d). This is followed by
1032 a renewed acceleration of flow speed, reaching 4.3 km yr^{-1}
1033 in 2021-2022 at PIG main grounding line, which leads to
1034 an increased outflow of ice into the ocean. The evolution
1035 of ice velocity along the southern tributary also indicates
1036 an acceleration of flow speed both on the ice shelf and up
1037 to 10 km inland (Fig. S23). Around the 2022 grounding
1038 line position, the increase reached approximately 30 %
1039 between 2017 and 2023. This is consistent with recent
1040 ice discharge estimations at PIG grounding line from (56)
1041 that show a stabilization around 130 Gt yr^{-1} from 2011
1042 to 2017, followed by a sharp increase reaching $\sim 155 \text{ Gt}$
1043 yr^{-1} in 2024 (Fig. 4b). Overall, this combination of high
1044 basal melt, grounding line retreat along with ice flow
1045 acceleration could indicate that PIG is entering a new
1046 phase of instability. This is coherent with known feedback
1047 loops between ice shelf buttressing, retreat and increased
1048 glacier flow (55, 57, 58), and has also been observed for
1049 other glaciers in the ASE (13, 14).

1050 Besides, evidence of damage on the southern part of
1051 the shelf are also visible in basal melt rates from 2018-
1052 2024, with the signature of basal crevasses ($\sim 40 \text{ m yr}^{-1}$)
1053 along the eastern edge of the basal channel previously
1054 mentioned. We also notice the presence of high melt rates
1055 ($\sim 80\text{-}100 \text{ m yr}^{-1}$) located near the southernmost edge
1056 of the main trunk (feature 6 on Fig 2). Indeed, optical
1057 imagery of this specific area reveals a collapse of the ice
1058 shelf over the 2018–2026 period, leaving this part of the
1059 main trunk shear zone replaced by ice melange or open sea
1060 (see Fig. S9). This collapse is accompanied by increased
1061 damage, along with the appearance of more pronounced
1062 basal and surface crevasses (Fig. S9). Such damage
1063 events have been documented to structurally weaken the
1064 ice shelf and precondition their eventual disintegration
1065 through self-reinforcing feedbacks (15). These feedbacks
1066 can be initiated by ocean-driven basal melting that
1067 thins shear margins, and mechanically weakens the ice
1068 shelf, making it prone to fracturing (59). Our results
1069 suggest that sustained channelized basal melting along
1070 the Pine Island Glacier southern tributary (feature 6 in
1071 Fig. 2) may have initiated a sequence of changes involving
1072 grounding-line retreat and ice-flow acceleration. Together,
1073 these processes likely enhanced thinning and strain rates
1074 within this already vulnerable sector of the ice shelf,
1075 promoting damage propagation and ultimately leading
1076 to the complete disconnection from the main ice shelf
1077 trunk (Fig. 2, Fig. S9). The disintegration of this shear-
1078 margin sector leaves PIG southern tributary effectively
1079 unbuttressed, creating conditions favorable for further
1080 acceleration and increased ice discharge to the ocean. The
1081 emergence of such rapid changes in a sector that was not
1082 previously considered the most unstable part of the glacier

highlights the evolving nature of PIG's dynamic response
to ocean-driven change.

Conclusions

Through a combination of multi sensor stereo-imagery and laser altimetry, this study provides a 20-year long time series of high-resolution basal melt rates and geometrical changes below the PIG ice shelf. Our results reveal a marked intensification of basal melt following a relatively stable period of anomaly in the inflow of CDW in the cavity in the mid-2010s. We show that after 2018-2020, all ice-ocean interfaces beneath 400 m are affected by intense melting of several tens of meters per year. The low level of hydrological activity in the region, combined with the close correspondence between basal melt rates and contemporaneous oceanographic records, indicates that variations in cavity water masses are the dominant driver of the observed changes. Temperatures at depth and melting are now reaching peak levels similar to the condition observed in the 2000s. The evolving pattern of ice-ocean interactions is accompanied by a reactivation of grounding-line retreat along both the main trunk of PIG and its southern tributary. Combined with ongoing ice-shelf thinning, these processes have driven to an expansion of the sub-ice-shelf cavity at depth over the last two decades. Channelized basal melting, coupled with grounding-line retreat and ice-flow acceleration along the southern tributary, is consistent with a marked increase in structural damage within this sector. The resulting progressive weakening ultimately led to its complete detachment from the main trunk of the ice shelf, leaving the southern tributary effectively unbuttressed. Overall, these findings may suggest the onset of a renewed phase of instability, pointing toward a potential future phase of increased mass loss for PIG.

Materials and Methods

Digital Elevation Models. We utilized DEMs from the REMA database (28) generated from imagery acquired by WorldView-1 to -3 and GeoEye-1 satellites. Stereoscopic images were collected over the polar regions mainly between 2009 and 2024, with native panchromatic resolution better than 50 cm. The time-stamped DEMs are produced and distributed at a 2-m resolution by the Polar Geospatial Center (PGC) until 2024. We also use acquisitions from the SPOT-5 High-Resolution Stereoscopic (HRS) instrument, which is characterized by rectangular pixels of 5 by 10 meters. These observations, distributed through the Spot World Heritage (SWH) program, cover the 2007–2010 period over the PIG region. The DEMs were generated using the Ames Stereo Pipeline (ASP; (60)) using the parameters from (61) and the semi global matching algorithm (62), to produce DEMs at a 20-m spatial resolution. In order to extend the time series up to the present, stereoscopic acquisitions from the Pléiades satellites were scheduled in early 2025. The images were acquired at a spatial resolution of 50 cm and cover nearly the entire ice shelf as well as part of the grounded ice near the grounding zone. DEMs were then produced with ASP at a spatial resolution of 2 m using the block-matching correlator (63). Finally, we complemented this time series with acquisitions from the ASTER instrument aboard NASA's TERRA satellite. Although its 15 m spatial resolution is lower than that of the previously mentioned sensors, ASTER offers the unique advantage of having collected continuous stereoscopic imagery since the early 2000s. ASTER DEMs were generated at 30 m resolution using the MMASTER processing chain. The latter is based on the MicMac photogrammetric software (30) and allows for efficient jitter-correction which was previously tested on Greenlandic ice shelves (7).

Laser Altimetry. To align the DEMs both horizontally and vertically, we used satellite and airborne laser altimetry data. We compiled all available data from NASA's Airborne Topographic Mapper (ATM) collected between

2002 and 2018 over PIG, particularly during the NASA Operation IceBridge (OIB) mission. Over the course of OIB, four different laser systems were used, with pulse repetition frequencies ranging from 3 to 10 kHz and laser footprints between 0.6 m and 1.2 m above the ground surface (64). We also used airborne data from the wide-swath laser altimeter LVIS (Land, Vegetation, and Ice Sensor) instrument, also deployed during OIB. Analyses of multiple crossover points and comparisons with GNSS data reported a vertical accuracy better than 0.06 m (65) and 0.07 m (66, 67) for ATM and LVIS respectively. Finally, we used the complete dataset from NASA's satellite laser altimetry missions: ICESat/GLAS (2003–2009) and ICESat-2 (2018–2024), with a reported vertical accuracy of 0.02–0.03 m (68, 69). All elevation values were corrected for the geoid using the EIGEN-6C4 geoid model to obtain heights above mean sea level. Floating sections of the ice shelves were then corrected for the inverse barometer effect using the ERA5 reanalysis from ECMWF (70), and for mean ocean topography using the CNES mean dynamic topography model *CNES – CLS – 22* (71). The corrections associated with these components are on the order of 0.07 m for atmospheric pressure and 1.14 m for mean dynamic topography, averaged over the study period. Finally, elevations were also corrected for ocean tides using the pyTMD toolbox, which applies the CATS2008 tide model (72) at each measurement point via a massively parallelized workflow. This correction is approximately 0.03 m over the entire study period. Since tides do not affect ice shelves uniformly, we applied a smoothing factor near the transition to grounded ice, to modulate this effect gradually as in (29). The distribution of available altimetry points on PIG ice shelf over time is displayed in Figure 1b. To ensure consistency in the reference datasets used for both vertical and horizontal DEM alignment, we chose to rely exclusively on laser altimetry measurements and therefore excluded CryoSat-2 radar altimetry data from this study. A more detailed comparison between laser altimetry and CryoSat-2 datasets is provided in the Supplement. However, this choice implies that, in years where sparse laser altimetry coverage (e.g., 2010), the DEM alignment is likely sub-optimal.

Co-registration procedure and construction of annual summer mosaics. We used a two-stage co-registration and mosaicking workflow to ensure both DEM accuracy and consistent annual coverage over the ice shelf. First, all corrected altimetry data acquired during the extended austral summer (September to April) were combined to generate an annual reference DEM. This temporal window was extended up to two summers during the early ASTER era (2000–2006) because of sparse data availability. All DEM strips acquired within the same time period were then filtered according to their consistency with the REMA mosaic (73) (50 m threshold, 100 m for ASTER products), ASTER correlation masks, and sufficient spatial overlap with altimetry data following criteria adapted from (74). Selected DEMs were corrected, vertically aligned, and tilt-corrected to the altimetry reference DEM using xDEM (75). The strip with the largest overlap initialized the annual pre-mosaic, after which remaining DEMs were progressively co-registered horizontally using the method of (76) and mosaicked in order of decreasing overlap. Light feathering was applied along boundaries to reduce edge artifacts (28), and the final pre-mosaic was vertically adjusted to the altimetry reference DEM.

In a second stage, all remaining DEM tiles from the considered summer were progressively added to the pre-mosaic following a similar procedure. When insufficient overlap with altimetry data existed, DEMs were first vertically aligned and tilt-corrected relative to the REMA mosaic. Tiles with poor alignment statistics (e.g., standard deviation > 5 m) were discarded and stored separately as leftover scenes. Remaining gaps in the final mosaics were filled using the altimetry reference DEM. If the resulting coverage was insufficient, overlap thresholds were relaxed to incorporate additional scenes. For years with particularly sparse coverage near the grounding line (mainly 2020–2023), multi-sensor mosaics combining REMA and ASTER DEMs were generated despite the lower accuracy of ASTER products. Finally, uncertainty maps were produced using the reported altimetry accuracies and comparisons with stereo-derived DEM datasets.

Grounding line changes and ice shelf boundaries. Boundaries of glacier floating parts evolve dynamically from year to year. For the ice shelf fronts, we used annual front maps derived from satellite imagery by (77) and updated until 2025 using summer Landsat-8/9 imagery. For the grounding lines, we relied on maps provided by the MEaSUREs project, which are obtained using a Quadruple Differential SAR Interferometry technique (QDInSAR, (23, 78)). We also used recent 1-day interferograms from the Sentinel-1C commissioning phase (22, 23), which allows us to update the position of the grounding line to 2025. Throughout the entire study period, we also used the dense fringe pattern at the grounding line, to provide a proxy on the width of the flexure zone, which is later used for basal melting rates post-processing. For years without available observations,

grounding line, flexure zones and ice front position were linearly interpolated toward their next available location. We analyzed grounding-line dynamics along four transects: three located across the main glacier trunk and one across the southern tributary, corresponding to the areas that experienced the largest changes during the study period (Fig. 4a & c). For each transect, we calculated the distance between the intersection point of each grounding line and the intersection point of the 1992 grounding line along the transect, providing a measure of grounding-line migration relative to its 1992 position.

Surface flow velocity processing. Flow velocity data were used both to calculate particle trajectories along the ice shelf and to compute the divergence of the flux. We first computed the displacements between each pair of DEMs in order to minimize high-frequency artifacts related to incorrect advection of crevasses. We used a modified version of the ampcor algorithm developed at the Jet Propulsion Laboratory (79). The cross-correlation calculation was applied directly to the DEM mosaics at 50 m resolution. We used a 64×64 pixel correlation window, with a grid spacing of 5×5 pixels, and a 100×100 pixel search window for 1-year pairs, and 200 pixels when the temporal spacing between pairs is larger. The displacement maps were then filtered using a median filter (3×3 pixels) and a directional filter at 45° (see below) to remove the main outliers. Since the spatial coverage of the image correlation is almost never complete over the ice shelf, we then merged this result with an external source of ice flow velocity (see below). For a given year, we first rescaled the external velocity using the overlapping areas with our correlation-based estimates, specifically in the fast-flowing parts of the glacier (>2 km yr⁻¹). The two maps were then merged using feathering at the boundaries between them. When the DEMs were spaced by three years or more, the surface features of the ice shelves were no longer temporally consistent due to the excessive flow of PIG. In this case, we used the annual velocity mosaics assembled as part of the MEaSUREs project, covering the study period from 2002 to 2023. These velocity maps (500 m resolution) were produced by combining radar interferometry (InSAR) techniques with speckle and feature tracking algorithms (33). In order to have a complete coverage over the ice shelf, we applied a different type of post-processing to annual velocities according to their raw ice shelf coverage (see Fig. S3). Firstly, if it was deemed sufficient (i.e. coverage > 70%), v_x and v_y velocity components were median filtered. We initially use a 4.5 km window, which was incrementally reduced to a 1 km minimum if the coverage drops below 70%. We computed the average flow direction and removed outliers directions diverging by more than 45° within a 10.5 km radius average (80). When coverage was above 70%, missing data on ice were interpolated using an image inpainting algorithm based on (81). Finally, velocities were smoothed using a Gaussian kernel with a standard deviation of 2 km. For coverage between 5–70%, gaps in velocity maps were first linearly interpolated in time pixel by pixel, using the nearest available years, then the same procedure was applied. For coverage < 5%, the full area was temporally interpolated before processing. In that way, we insured a comprehensive surface flow velocity mosaic for every year of our study period (see Fig. S24).

Surface mass balance and firn air content. To assess the impact of firn processes on basal melt rate estimates, we used three physically based firn models: GEMB v1.0, GSFC-FDM v1.2.1 and IMAU-FDM v1.2A—to simulate firn air content (FAC) across the ice shelf. All three models represent firn column evolution, but differ in their physical complexity, spatial resolution, and atmospheric forcing (34, 37, 82). We downscaled FAC estimates and interpolated missing regions near the ice shelf front using cubic interpolation and smoothed using a 5.5 km window median filter. While all models provide FAC estimates, only GEMB provides open-access estimates of SMB. They are expressed as surface height tendencies (in meters), which can be directly incorporated into the mass conservation equation. For this reason, GEMB SMB was used as the surface boundary condition in our basal melt rate estimates. However, SMB variability over the PIG ice shelf is small (Fig. S4). Mean SMB during the study period is 0.2 m yr⁻¹, with a mean interannual trend of only -0.01 m yr⁻². This contribution is negligible relative to observed elevation changes and ice-flux divergence (<1%), and therefore has little influence on the inferred basal melt rates. FAC variations from year to year are less than 1% (see Fig. S25) throughout the study period, further indicating that firn densification has a minor contribution to long-term ice shelf thickness change in this region. Therefore we neglect FAC temporal evolution term in the basal melt calculation as in (29). The impact of firn models on basal melting rates is therefore evaluated through its impact on the calculation of ice flux divergence, as surface height changes are corrected from the contribution of firn compaction and porosity. Over the 2002–2024 period, mean FAC over the ice shelf is equivalent to 13.3 m, 17.8 m and 21.3 m for GEMB, IMAU-FDM and GSFC outputs, respectively, resulting in a cross-model standard deviation of 3.3 m.

Basal melt rates. The ice shelf basal melt rates were calculated by solving the mass conservation equation over an ice column in a Lagrangian framework:

$$\dot{b} = \dot{a} - \left(\frac{DH}{Dt} + H(\nabla \cdot \mathbf{u}) \right) \quad [1]$$

with $\dot{b}$ the basal melt rate (defined here as positive for melt), $\dot{a}$ the surface mass balance (in meters of ice equivalent), H the ice thickness, $\frac{DH}{Dt}$ the Lagrangian thickness change and $\mathbf{u}$ the ice flow velocity. Assuming hydrostatic equilibrium of floating ice we can invert ice thickness H from surface height h :

$$H \approx \left(\frac{\rho_w}{\rho_w - \rho_i} \right) (h - d) \quad [2]$$

with d the firn air content and assuming constant density for ice (917 kg m^{-3}) and ocean water (1.028 kg m^{-3}). We assume that changes in d are negligible over the study period as in (29) (Fig. S25).

To compute the ice shelf basal melt rate between time stamps t_i and t_j in a Lagrangian framework, we tracked pixel trajectories along flow paths from DEM_i to DEM_j using a 15-day time step as in (29). Using yearly DEMs from 2002–2024, we generated 87 different melt-rate products with temporal baselines ranging from 1 to 6 years. Longer intervals ($Dt = t_j - t_i > 3$ years) showed a better signal-to-noise ratio but a restrained spatial extent compared to short time intervals. Before processing, DEM points below 10 m elevation and below firn air content were removed to avoid including open ocean, sea ice, and unrealistic negative thickness estimates (38). Surface elevation changes, ice-flux divergence, and SMB variations were all sampled along Lagrangian trajectories as in (29). Since we were using dynamically evolving ice front and grounding lines, basal melt rates were computed over common ice-shelf area shared between DEM_i and DEM_j , and assigned to the initial pixel locations following the initial-pixel approach (29). We post-processed the melt rates in the ice-shelf flexure zone, where the assumption of hydrostatic equilibrium is not fully satisfied. Unrealistic refreezing pattern were removed and replaced by median-filtered rates (2 km window), with a light feathering applied to ensure smoother transitions. The final multi-year melt-rate mosaics was produced by stacking all products from 2002–2024 to maximize spatial coverage (Fig. 2). Each melt map was assigned to the central date between t_i and t_j , and annual mosaics were subsequently produced by taking the median of all maps whose central dates fall within a given year. Because individual melt-rate estimates integrate elevation change over variable time intervals ranging from 1 to 6 years, the resulting products should be interpreted as a temporally smoothed representation of basal melt rate variability rather than as independent annual estimates. Separate analyses using short (1–3 years), long (4–6 years), and full (1–6 years) DEM pairs showed consistent spatial and temporal patterns (Fig. 2, 3, S16), supporting the robustness of the results.

Uncertainty quantification. Basal melt rate uncertainties were quantified through quadratic error propagation, incorporating independent uncertainties from SMB and FAC model outputs, ice velocity fields, and DEM mosaics. We assumed an arbitrary and conservative errors of 20 % on the SMB and

the FAC products, which is justified by their comparatively minor contribution to the basal melt budget and by the spreading between FAC models. For ice flow velocity, a constant uncertainty of 5 m yr^{-1} was assumed on the yearly mosaics for both horizontal components (33). Uncertainty in surface elevation for each annual DEM mosaic was estimated using the normalized median absolute deviation (NMAD) relative to the corresponding reference DEM. For pixels directly derived from the reference DEM, a fixed vertical uncertainty of 0.1 m, corresponding to laser altimetry accuracy, was applied. These individual uncertainty sources were then propagated to generate spatially distributed uncertainty maps for each basal melt estimate. To derive area-averaged uncertainties in melt rates over entire ice shelf or sub-regions, spatial error correlations were accounted for by estimating the effective number of independent samples (N_{eff}). The uncertainty in the mean basal melt rate (σ_{melt}) was then computed as:

$$\sigma_{melt} = \frac{\sigma_{melt}}{\sqrt{N_{eff}}} \quad [3]$$

The effective sample size N_{eff} was determined by modeling the spatial variogram of the basal melt rate data using built-in functions from the xDEM framework (83).

CTD analysis. To compare the evolution of melt rates derived from satellite observations with oceanic conditions, we reconstructed the oceanic forcing by compiling CTD measurements from the Hadley Centre (<https://bodc.ac.uk>, last access: April 2026) in the Amundsen Sea, with a particular focus on the deep trough extending toward the Pine Island Glacier Ice Shelf (Fig. 4a). The Hadley Centre compiles CTD data dating back to the 1960s, but for our analysis, we concentrated on the period from the late 1990s to nowadays. Within the defined area of interest, we manually filtered out outliers and redundant CTD casts by visually inspecting each profile (Fig. S26). We then extracted the mean temperature values of the warm water layer located below 400 m depth (see Fig. 4b).

Code and data availability. ASTER imagery and REMA DEMs are freely available at earthdata.nasa.gov, and at <https://www.pgc.umn.edu/data/rema/>. All laser altimetry data used in this study are available at <https://nsidc.org>. SPOT imagery can be accessed at <https://regards.cnes.fr> through CNES's Spot World Heritage program. Pléiades Digital Elevation Model from 2025 and basal melting rates derived in this study will be made available upon acceptance of this manuscript on Zenodo. All codes to replicate the analysis in this study will be made available upon acceptance of the manuscript on a public github repository.

ACKNOWLEDGMENTS. This research has been supported by the European Research Council (ERC) under the Horizon Framework research and innovation program of the European Union (grant no. 101164392; "IceDaM" project) and by the Fondation Albédo sous l'Egide du CNRS through the ANTARMELT project. J.A acknowledges support from the Villum Fonden (grant no. 29456).

1. B Fox-Kemper, et al., Ocean, cryosphere and sea level change in *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*, eds. V Masson-Delmotte, et al. (Cambridge University Press, Cambridge, UK and New York, NY, USA), pp. 1211–1361 (2021).
2. M Moriglighem, et al., Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the antarctic ice sheet. *Nat. Geosci.* **13**, 132–137 (2020).
3. M Oppenheimer, et al., Sea level rise and implications for low-lying islands, coasts and communities in *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*, eds. HO Pörtner, et al. (Cambridge University Press, Cambridge, UK and New York, NY, USA), pp. 321–445 (2019).
4. TA Scambos, J Bohlander, CA Shuman, P Skvarca, Glacier acceleration and thinning after ice shelf collapse in the larsen b embayment, antarctica. *Geophys. Res. Lett.* **31** (2004).
5. E Rignot, et al., Accelerated ice discharge from the antarctic peninsula following the collapse of larsen b ice shelf. *Geophys. research letters* **31** (2004).
6. J Mouginot, et al., Fast retreat of zachariæ isstrøm, northeast greenland. *Science* **350**, 1357–1361 (2015).
7. R Millan, et al., Rapid disintegration and weakening of ice shelves in north greenland. *Nat. Commun.* **14**, 6914 (2023).
8. E Rignot, et al., Four decades of antarctic ice sheet mass balance from 1979–2017. *Proc. Natl. Acad. Sci.* **116**, 1095–1103 (2019).
9. FS Paolo, et al., Widespread slowdown in thinning rates of west antarctic ice shelves. *The Cryosphere* **17**, 3409–3433 (2023).
10. E Rignot, S Jacobs, J Mouginot, B Scheuchl, Ice-shelf melting around antarctica. *Science* **341**, 266–270 (2013).
11. S Adusumilli, HA Fricker, B Medley, L Padman, MR Siegfried, Interannual variations in meltwater input to the southern ocean from antarctic ice shelves. *Nat. geoscience* **13**, 616–620 (2020).
12. K Lowery, et al., Spatio-temporal melt and basal channel evolution on pine island glacier ice shelf from cryosat-2. *The Cryosphere* **19**, 4893–4911 (2025).
13. E Rignot, J Mouginot, M Moriglighem, H Seroussi, B Scheuchl, Widespread, rapid grounding line retreat of pine island, thwaites, smith, and kohler glaciers, west antarctica, from 1992 to 2011. *Geophys. research letters* **41**, 3502–3509 (2014).
14. J Mouginot, E Rignot, B Scheuchl, Sustained increase in ice discharge from the amundsen sea embayment, west antarctica, from 1973 to 2013. *Geophys. Res. Lett.* **41**, 1576–1584 (2014).
15. S Lhermitte, et al., Damage accelerates ice shelf instability and mass loss in amundsen sea embayment. *Proc. Natl. Acad. Sci.* **117**, 24735–24741 (2020).
16. M Izeboud, L Stef, Fracturing of antarctic ice shelves depends on future climate warming rate. *Nat. Clim. Chang.* **15**, 1277–1278 (2025).
17. P Dutrieux, et al., Strong sensitivity of pine island ice-shelf melting to climatic variability. *Science* **343**, 174–178 (2014).
18. A Jenkins, et al., West antarctic ice sheet retreat in the amundsen sea driven by decadal oceanic variability (2018).
19. P St-Laurent, PL Yager, RM Sherrell, SE Stammerjohn, MS Dinniman, Pathways and supply of dissolved iron in the amundsen sea (antarctica). *J. Geophys. Res. Ocean.* **122**, 7135–7162 (2017).
20. B Reed, JM Green, A Jenkins, GH Gudmundsson, Recent irreversible retreat phase of pine island glacier. *Nat. Clim. Chang.* **14**, 75–81 (2024).
21. SHR Rosier, et al., The tipping points and early warning indicators for pine island glacier, west antarctica. *The Cryosphere* **15**, 1501–1516 (2021).

1403	22. JK Andersen, et al., Brief communication: Updated grounding line mapping in the amundsen sea embayment, antarctica, from one day repeat sentinel-1 sar data. <i>The Cryosphere</i> 20 , 1589–1598 (2026).	62. H Hirschmuller, Stereo processing by semiglobal matching and mutual information. <i>IEEE Transactions on Pattern Analysis Mach. Intell.</i> 30 , 328–341 (2008).	1467
1404		63. MJ Willis, BG Herried, MG Bevis, RE Bell, Recharge of a subglacial lake by surface meltwater in northeast greenland. <i>Nature</i> 518 , 223–227 (2015).	1468
1405	23. E Rignot, et al., Thirty years of glacier grounding line retreat in antarctica. <i>Proc. Natl. Acad. Sci.</i> 123 , e2524380123 (2026).	64. JA MacGregor, et al., The scientific legacy of nasa's operation icebridge. (2021).	1469
1406	24. BJ Davison, et al., Annual mass budget of antarctic ice shelves from 1997 to 2021. <i>Sci. Adv.</i> 9 , eadi0186 (2023).	65. CF Martin, et al., Airborne topographic mapper calibration procedures and accuracy assessment, Technical report (2012).	1470
1407		66. M Hofton, J Blair, S Luthcke, D Rabine, Assessing the performance of 20–25 m footprint waveform lidar data collected in icesat data corridors in greenland. <i>Geophys. Res. Lett.</i> 35 (2008).	1471
1408	25. A Zinck, S Lhermitte, M Wearing, B Wouters, Channelized melt beneath antarctic ice shelves previously underestimated. <i>Nat. Clim. Chang.</i> 16 , 350–353 (2026).	67. KM Brunt, et al., Assessment of nasa airborne laser altimetry data using ground-based gps data near summit station, greenland. <i>The Cryosphere</i> 11 , 681–692 (2017).	1472
1409	26. Q Zhou, et al., Channelized topography amplifies melt-sensitivity of cold Antarctic ice shelves. <i>Nat. Commun.</i> 17 , 3790 (2026).	68. JB Abshire, et al., Geoscience laser altimeter system (glas) on the icesat mission: On-orbit measurement performance. <i>Geophys. Res. Lett.</i> 32 (2005).	1473
1410		69. T Markus, et al., The ice, cloud, and land elevation satellite-2 (icesat-2): Science requirements, concept, and implementation. <i>Remote. Sens. Environ.</i> 190 , 260–273 (2017).	1474
1411	27. J Korona, E Berthier, M Bernard, F Rémy, E Thouvenot, Spirit. spot 5 stereoscopic survey of polar ice: reference images and topographies during the fourth international polar year (2007–2009). <i>ISPRS journal photogrammetry remote sensing</i> 64 , 204–212 (2009).	70. H Hersbach, et al., The era5 global reanalysis. <i>Q. Journal royal meteorological society</i> 146 , 1999–2049 (2020).	1475
1412	28. IM Howat, C Porter, BE Smith, MJ Noh, P Morin, The reference elevation model of antarctica. <i>The Cryosphere</i> 13 , 665–674 (2019).	71. S Jousset, et al., New global mean dynamic topography cnes-cls-22 combining drifters, hydrography profiles and high frequency radar data. <i>Earth Syst. Sci. Data Discuss.</i> 2025 , 1–32 (2025).	1476
1413	29. DE Shean, IR Joughin, P Dutrieux, BE Smith, E Berthier, Ice shelf basal melt rates from a high-resolution digital elevation model (dem) record for pine island glacier, antarctica. <i>The Cryosphere</i> 13 , 2633–2656 (2019).	72. L Padman, HA Fricker, R Coleman, S Howard, L Erofeeva, A new tide model for the antarctic ice shelves and seas. <i>Annals Glaciol.</i> 34 , 247–254 (2002).	1477
1414	30. L Girod, C Nuth, A Käab, R McNabb, O Galland, Mmaster: Improved aster dems for elevation change monitoring. <i>Remote. Sens.</i> 9 (2017).	73. I Howat, et al., The Reference Elevation Model of Antarctica - Mosaics, Version 2 (2022) Last accessed: 2026-03-04.	1478
1415	31. M Bernat, et al., Mass changes of the antarctic peninsula ice sheet and peripheral glaciers, 2007–2021. <i>EGUosphere</i> 2026 , 1–44 (2026).	74. ASP Zinck, B Wouters, E Lambert, S Lhermitte, Unveiling spatial variability within the dotson melt channel through high-resolution basal melt rates from the reference elevation model of antarctica. <i>The Cryosphere</i> 17 , 3785–3801 (2023).	1479
1416	32. ER Council, Icedam (2025) Grant agreement ID: 101164392.	75. xDEM contributors, xdem (v0.1.0) (2024).	1480
1417	33. J Mougintot, E Rignot, B Scheuchl, R Millan, Comprehensive annual ice sheet velocity mapping using landsat-8, sentinel-1, and radarsat-2 data. <i>Remote. Sens.</i> 9 , 364 (2017).	76. C Nuth, A Käab, Co-registration and bias corrections of satellite elevation data sets for quantifying glacier thickness change. <i>The Cryosphere</i> 5 , 271–290 (2011).	1481
1418	34. AS Gardner, NJ Schlegel, E Larour, Glacier energy and mass balance (gemb): a model of firn processes for cryosphere research. <i>Geosci. Model. Dev.</i> 16 , 2277–2302 (2023).	77. CA Greene, AS Gardner, NJ Schlegel, AD Fraser, Antarctic calving loss rivals ice-shelf thinning. <i>Nature</i> 609 , 948–953 (2022).	1482
1419	35. Firn on ice sheets. <i>Nat. Rev. Earth & Environ.</i> 5 , 79–99 (2024).	78. E Rignot, J Mougintot, B Scheuchl, Antarctic grounding line mapping from differential satellite radar interferometry. <i>Geophys. Res. Lett.</i> 38 (2011).	1483
1420	36. SBM Veldhuijsen, WJ van de Berg, M Brils, P Kuipers Munneke, MR van den Broeke, Characteristics of the 1979–2020 antarctic firn layer simulated with imau-fdm v1.2a. <i>The Cryosphere</i> 17 , 1675–1696 (2023).	79. R Millan, et al., Mapping surface flow velocity of glaciers at regional scale using a multiple sensors approach. <i>Remote. Sens.</i> 11 , 2498 (2019).	1484
1421	37. B Medley, TA Neumann, HJ Zwally, BE Smith, CM Stevens, Simulations of firn processes over the greenland and antarctic ice sheets: 1980–2021. <i>The Cryosphere</i> 16 , 3971–4011 (2022).	80. A Rabatel, E Ducasse, R Millan, J Mougintot, Satellite-derived annual glacier surface flow velocity products for the european alps, 2015–2021. <i>Data</i> 8 (2023).	1485
1422	38. AM Chartrand, IM Howat, A comparison of contemporaneous airborne altimetry and ice-thickness measurements of antarctic ice shelves. <i>J. Glaciol.</i> 69 , 1663–1676 (2023).	81. M Bertalmio, AL Bertozzi, G Sapiro, Navier-stokes, fluid dynamics, and image and video inpainting. <i>Proc. 2001 IEEE Comput. Soc. Conf. on Comput. Vis. Pattern Recognition. CVPR 2001</i> 1 , I–I (2001).	1486
1423	39. SS Jacobs, HH Hellmer, A Jenkins, Antarctic ice sheet melting in the southeast pacific. <i>Geophys. Res. Lett.</i> 23 , 957–960 (1996).	82. JM van Wessem, et al., Modelling the climate and surface mass balance of polar ice sheets using racmo2 – part 2: Antarctica (1979–2016). <i>The Cryosphere</i> 12 , 1479–1498 (2018).	1487
1424	40. A Jenkins, et al., Observations beneath pine island glacier in west antarctica and implications for its retreat. <i>Nat. Geosci.</i> 3 , 468–472 (2010).	83. R Hugonnet, et al., Uncertainty analysis of digital elevation models by spatial inference from stable terrain. <i>IEEE J. Sel. Top. Appl. Earth Obs. Remote. Sens.</i> 15 , 6456–6472 (2022).	1488
1425	41. R Millan, E Rignot, V Bernier, M Morlighem, P Dutrieux, Bathymetry of the amundsen sea embayment sector of west antarctica from operation icebridge gravity and other data. <i>Geophys. Res. Lett.</i> 44 , 1360–1368 (2017).		1489
1426	42. BG Webber, et al., Mechanisms driving variability in the ocean forcing of pine island glacier. <i>Nat. communications</i> 8 , 14507 (2017).		1490
1427	43. CT Wild, et al., Thwaites eastern ice shelf cavity observations reveal multiyear sea ice dynamics and deepwater warming in pine island bay, west antarctica. <i>Ocean. Sci.</i> 21 , 2605–2629 (2025).		1491
1428	44. PR Holland, SL Bevan, AJ Luckman, Strong ocean melting feedback during the recent retreat of thwaites glacier. <i>Geophys. Res. Lett.</i> 50 , e2023GL103088 (2023).		1492
1429	45. N Gourmelen, et al., The influence of subglacial lake discharge on thwaites glacier ice-shelf melting and grounding-line retreat. <i>Nat. Commun.</i> 16 , 2272 (2025).		1493
1430	46. A Jenkins, Convection-driven melting near the grounding lines of ice shelves. <i>J. Phys. Oceanogr.</i> 41 , 2279–2294 (2011).		1494
1431	47. SF Wilson, et al., Detection of 85 new active subglacial lakes in antarctica from a decade of cryosat-2 data. <i>Nat. Commun.</i> 16 , 8311 (2025).		1495
1432	48. D Corr, A Leeson, M McMillan, C Zhang, T Barnes, An inventory of supraglacial lakes and channels across the west antarctic ice sheet. <i>Earth Syst. Sci. Data</i> 14 , 209–228 (2022).		1496
1433	49. N Maier, JK Andersen, J Mougintot, F Gimbert, O Gagliardini, Wintertime supraglacial lake drainage cascade triggers large-scale ice flow response in greenland. <i>Geophys. Res. Lett.</i> 50 , e2022GL102251 (2023).		1497
1434	50. M Moussavi, et al., Antarctic supraglacial lake detection using landsat 8 and sentinel-2 imagery: towards continental generation of lake volumes. <i>Remote. Sens.</i> 12 , 134 (2020).		1498
1435	51. NP Fofonoff, RC Millard, Algorithms for computation of fundamental properties of seawater. <i>UNESCO Tech. Pap. Mar. Sci.</i> (1983).		1499
1436	52. E Rignot, et al., Widespread seawater intrusions beneath the grounded ice of thwaites glacier, west antarctica. <i>Proc. Natl. Acad. Sci.</i> 121 , e2404766121 (2024).		1500
1437	53. L Favier, et al., Retreat of pine island glacier controlled by marine ice-sheet instability. <i>Nat. Clim. Chang.</i> 4 , 117–121 (2014).		1501
1438	54. I Joughin, BE Smith, B Medley, Marine ice sheet collapse potentially under way for the thwaites glacier basin, west antarctica. <i>Science</i> 344 , 735–738 (2014).		1502
1439	55. J Weertman, Stability of the junction of an ice sheet and an ice shelf. <i>J. Glaciol.</i> 13 , 3–11 (1974).		1503
1440	56. BJ Davison, AE Hogg, T Slater, R Rigby, N Hansen, Antarctic ice sheet grounding line discharge from 1996–2024. <i>Earth Syst. Sci. Data</i> 17 , 3259–3281 (2025).		1504
1441	57. TK Dupont, RB Alley, Assessment of the importance of ice-shelf buttressing to ice-sheet flow. <i>Geophys. Res. Lett.</i> 32 (2005).		1505
1442	58. RH Thomas, CR Bentley, A model for holocene retreat of the west antarctic ice sheet. <i>Quat. Res.</i> 10 , 150–170 (1978).		1506
1443	59. KE Alley, TA Scambos, MR Siegfried, HA Fricker, Impacts of warm water on antarctic ice shelf stability through basal channel formation. <i>Nat. Geosci.</i> 9 , 290–293 (2016).		1507
1444	60. RA Beyer, O Alexandrov, S McMichael, The ames stereo pipeline: Nasa's open source software for deriving and processing terrain data. <i>Earth Space Sci.</i> 5 , 537–548 (2018).		1508
1445	61. C Deschamps-Berger, et al., Snow depth mapping from stereo satellite imagery in mountainous terrain: evaluation using airborne lidar data. <i>The Cryosphere Discuss.</i> 2020 , 1–28 (2020).		1509
1446			1510
1447			1511
1448			1512
1449			1513
1450			1514
1451			1515
1452			1516
1453			1517
1454			1518
1455			1519
1456			1520
1457			1521
1458			1522
1459			1523
1460			1524
1461			1525
1462			1526
1463			1527
1464			1528
1465			1529
1466			1530



Supporting Information for

Ocean-driven reactivation of basal melt and dynamic instability beneath Pine Island Glacier ice shelf since 2018

Lucille Gimenes^a, Romain Millan^a, Jean Baptise-Barré^a, Amandine Clement^a, Jonas K. Andersen^b, Maud Bernat^d, Kaian Shahateet^{a,c}, and Antoine Rabatel^a

^aUniv. Grenoble-Alpes, CNRS, IRD, INRAE, Grenoble-INP, Institut des Géosciences de l'Environnement, Grenoble, France;

^bDepartment of Geosciences and Natural Resource Management, University of Copenhagen, Copenhagen K, Denmark;

^cAeronautics Institute of Technology, São José dos Campos, Brazil;

^dUniversité de Toulouse, LEGOS (CNES/CNRS/IRD/UT3), Toulouse, France

Lucille Gimenes

E-mail: lucille.gimenes@univ-grenoble-alpes.fr

Romain Millan

E-mail: romain.millan@univ-grenoble-alpes.fr

This PDF file includes:

Supporting text

Figs. S1 to S26

Tables S1 to S2

SI References

Supporting Information Text

Comparison of basal melt rate product

We compare our melt rates spatial distribution with five published datasets. Three are based on radar altimetry at resolutions of 250 m (1) (LOW25), 1 km (2) (DAV23) and 2 km (3) (PAO23). We also include products based on stereo imagery and altimetry at 250 m (4) (SHE19) and 50 m (5) (ZIN26). All datasets assume hydrostatic equilibrium and use mass conservation (Lagrangian, except for PAO23), but only PAO23 and LOW25 provide time series. The basal melt patterns from our study are broadly consistent at large scale with basal melt rates derived from PAO23 and DAV23 global altimetry product, but reveal an enhanced level of detail thanks to the high spatial resolution (50 m), and significant differences in magnitude (Fig. S7). More specifically in the main grounding zone region, PAO23 and DAV23 show a smoother signal and underestimate the basal melt magnitude up to 90% compared to our study (Fig. S7j & m). LOW25 mosaics derived from CryoSat-2 swath processing is however much closer to our estimate in this region, but is still lower by 30% near the two hot spots (features 1 & 2) previously identified (Fig. S7k). In regions of low basal melt rate magnitude, such as PIG northern tributary ice shelf (Fig. S7c), the use of stereo-photogrammetric data allows to resolve basal channels at the scale of tens of meters. This is not the case in radar altimetry products that are either too smooth (PAO23, DAV23), or with a low signal-to-noise ratio (SNR) (LOW26) (Fig. S7e). Although these have the advantage to provide year-round measurements, the high-frequency noise is further exacerbated when examining yearly mosaics (Fig. S8f), where the SNR from stereo-photogrammetric DEMs seems to outperform the one from CryoSat-2.

Overall, our estimates are consistent with previous products based on stereo-photogrammetric data (4, 5), both in terms of magnitude and spatial patterns (Fig. S7c, f & h). The main differences are visible in the basal melt map of (5), particularly within the grounding zone region, where “tree-ring”-like artifacts are apparent, likely related to advection patterns near the grounding line (Fig. S7l). In regions of low basal melt (*e.g.*, the northern tributary ice shelf), (5) also show anomalous refreezing patterns that spatially coincide with REMA DEM strip boundaries, hence suggesting biases related to mosaicking and/or alignment.

We also compared the evolution of the average basal melt near the main grounding line with time series from other datasets (Fig. S7b). Quarterly basal melt rates from PAO23 follow a trend similar to that of our time series, with a good agreement in the early 2000s. However, PAO23 rates are approximately 30% lower than ours and with greater temporal variability until 2017. From 2012 to 2022, the LOW25 monthly time series remains on average close to the melting rates computed in this study but display a much stronger high frequency signal with, for example, mean basal melt rates varying between 95 and 125 m yr⁻¹ within 2019-2020 (Fig. S7b).

Discrepancies between DEM-based basal melt maps and those derived from altimetry at the Antarctica scale may stem from several factors. In the case of PAO23, this could be related to the use of an Eulerian framework, with coarse grid resolution (<2 km) and the application of strong smoothing techniques to reduce noise in raw data and time series. The remaining high-frequency spatial and temporal noise in regional CryoSat-2 derived datasets (1) suggests that the use of stereo-photogrammetry might be better suited for high-resolution monitoring of channelized basal melting rates. Additional uncertainties in these datasets also arise from interferometric phase unwrapping errors (6), waveform re-tracking biases (7), and radar penetration effects that vary with firn properties and location (Fig S19- S20). While stereo-photogrammetric estimates agree on average, discrepancies in signal-to-noise ratio can be explained by differences in the temporal periods used for the mosaics, different sensors (*e.g.*, ASTER which is generally noisier), and various co-registration and mosaicking approaches.

Comparison of ICESat-2 and Cryosat-2 data

We compared the CryoTEMPO-EOLIS (L3) products from CryoSat-2 with ICESat-2 ATL06 data over grounded ice (Fig. S19, S20). This comparison indicates a mean difference of -1 ± 10 m (mean $\pm$ standard deviation) for Point-of-Closest-Approach (POCA) data and -3.2 ± 5.5 m for interferometric swath data. An analysis of these biases by elevation band (Fig. S19c, S20d) shows that the standard deviation is highest at low elevations and near the grounding zone, where it can exceed 14 m for POCA data and 7 m for swath data. It then decreases progressively toward the interior of the ice sheet. Conversely, both median and mean biases tend to increase inland. While these biases remain small for POCA data, a systematic bias greater than 2 m persists in the swath products. Based on these results, we chose not to use CryoSat-2 radar altimetry data in this study, in order to ensure consistency in the calibration datasets by relying exclusively on laser altimetry measurements. This choice, however, implies that for years with sparse laser data coverage (*e.g.*, 2010), the DEM alignment is likely suboptimal and may rely primarily on one or two airborne survey lines.

Table S1. Comparison of integrated basal melt over PIG ice shelf for the three different used FAC estimates.

FAC model	Melt rate (Gt.yr ⁻¹)
GEMB	136
GSFC	134
IMAU-FDM	135

Table S2. Comparison of integrated basal melt over PIG ice shelf, computed on the area of intersection between the valid data zones of four datasets (4247 km²). Calculation was only done for datasets that we were able to compute over the same time period (2010-2017)

Dataset	Melt rate (Gt.yr ⁻¹)	Error (Gt.yr ⁻¹)
This study	62	± 21
PAO23	62	± 8
LOW25	76	-
ZIN26	97	-

References

1. K Lowery, et al., Spatio-temporal melt and basal channel evolution on pine island glacier ice shelf from cryosat-2. *The Cryosphere* **19**, 4893–4911 (2025).
2. BJ Davison, et al., Annual mass budget of antarctic ice shelves from 1997 to 2021. *Sci. Adv.* **9**, eadi0186 (2023).
3. FS Paolo, et al., Widespread slowdown in thinning rates of west antarctic ice shelves. *The Cryosphere* **17**, 3409–3433 (2023).
4. DE Shean, IR Joughin, P Dutrieux, BE Smith, E Berthier, Ice shelf basal melt rates from a high-resolution digital elevation model (dem) record for pine island glacier, antarctica. *The Cryosphere* **13**, 2633–2656 (2019).
5. A Zinck, S Lhermitte, M Wearing, B Wouters, Channelized melt beneath antarctic ice shelves previously underestimated. *Nat. Clim. Chang.* **16**, 350—353 (2026).
6. N Gourmelen, et al., Cryosat-2 swath interferometric altimetry for mapping ice elevation and elevation change. *Adv. Space Res.* **62**, 1226–1242 (2018) The CryoSat Satellite Altimetry Mission: Eight Years of Scientific Exploitation.
7. A Morris, G Moholdt, L Gray, TV Schuler, T Eiken, Cryosat-2 interferometric mode calibration and validation: A case study from the austfonna ice cap, svalbard. *Remote. Sens. Environ.* **269**, 112805 (2022).
8. JK Andersen, et al., Brief communication: Updated grounding line mapping in the amundsen sea embayment, antarctica, from one day repeat sentinel-1 sar data. *The Cryosphere* **20**, 1589–1598 (2026).
9. M Morlighem, et al., Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the antarctic ice sheet. *Nat. Geosci.* **13**, 132–137 (2020).
10. M Moussavi, et al., Antarctic supraglacial lake detection using landsat 8 and sentinel-2 imagery: towards continental generation of lake volumes. *Remote. Sens.* **12**, 134 (2020).

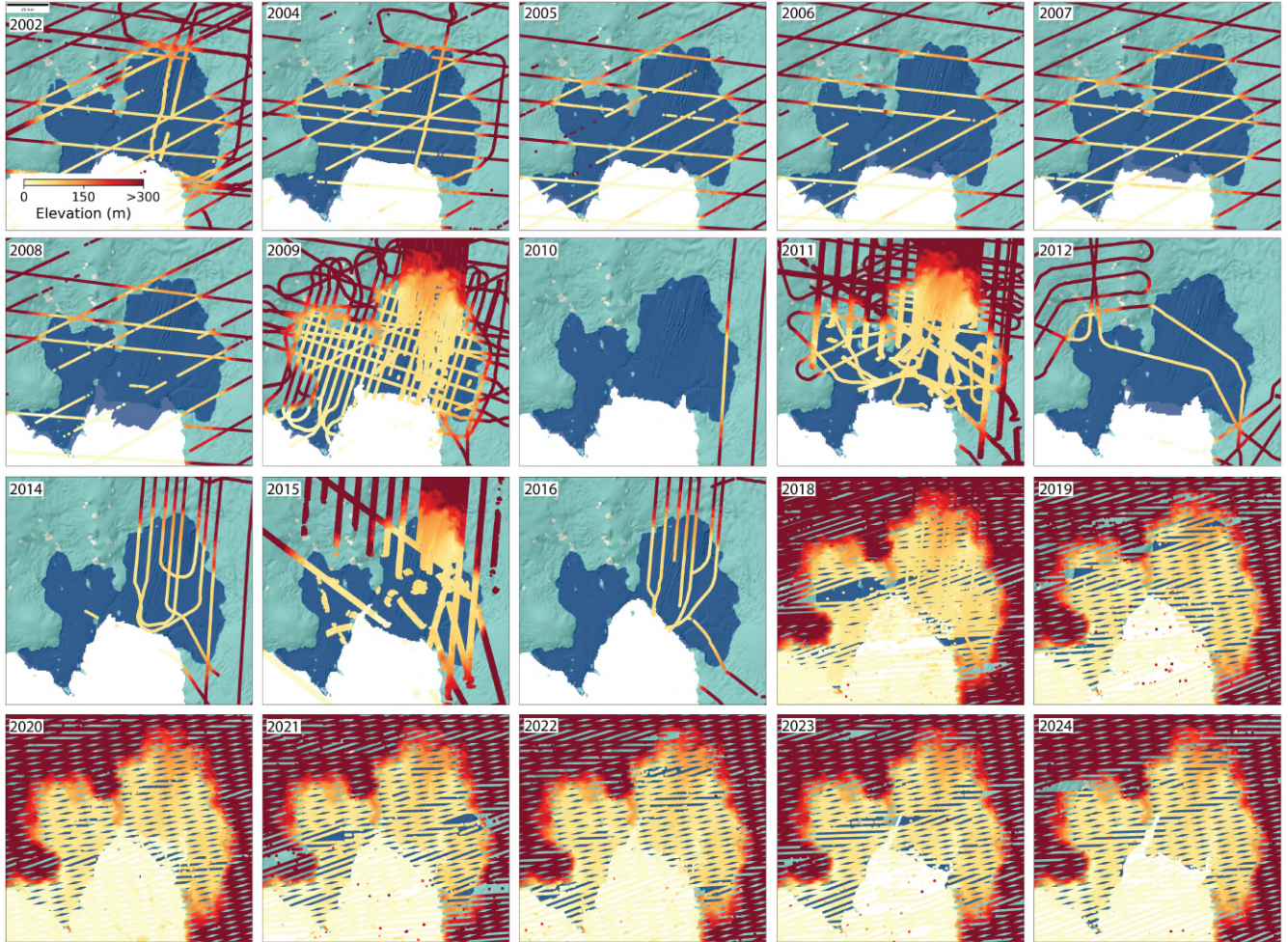


Fig. S1. All reference elevation DEM generated over PIG for the 2002-2024 period.

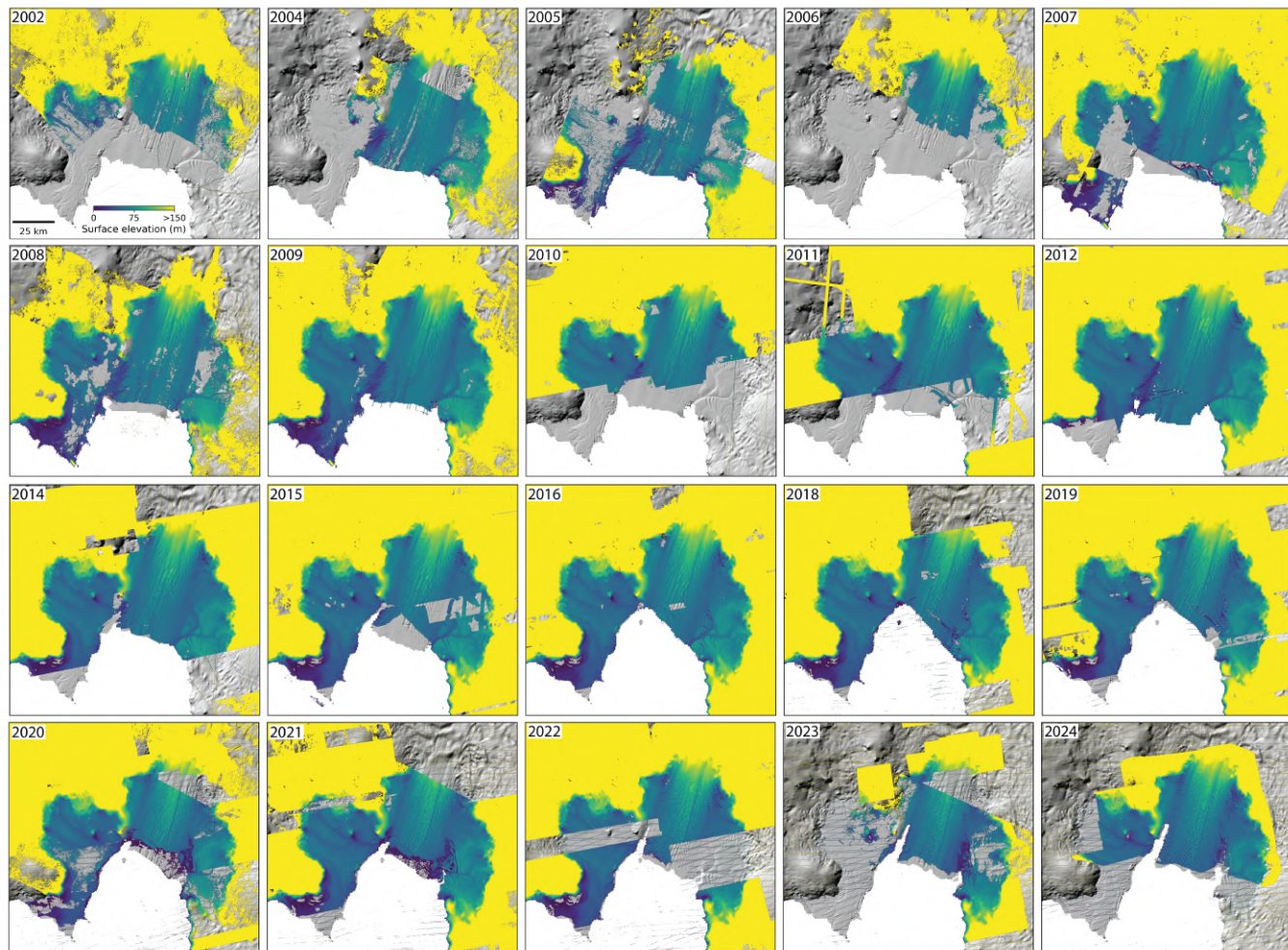


Fig. S2. All surface elevation mosaics generated over PIG for the 2002-2024 period.

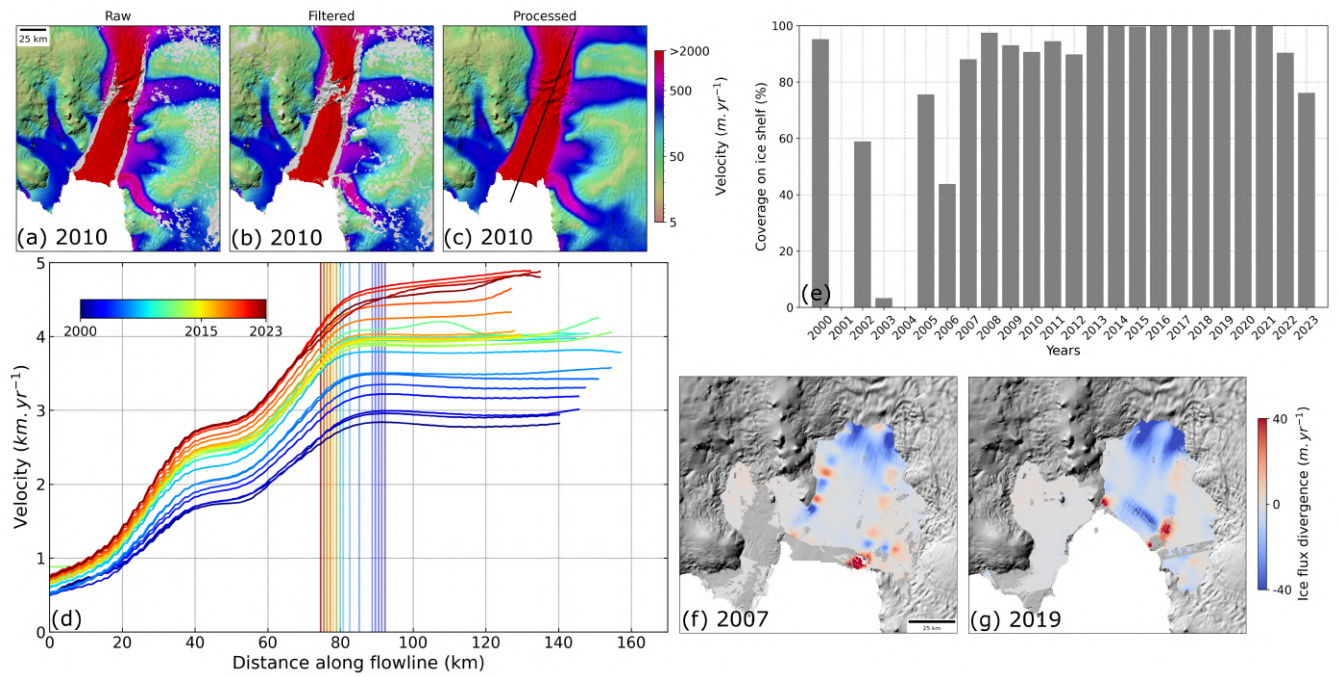


Fig. S3. Example of processing steps on the velocity magnitude for the year 2010 (a-c). Average velocity profile along flowline displayed in (c), color-coded by years, with vertical lines representing to grounding line positions (d). Temporal variability of raw velocity coverage on ice shelf (e). Examples of computed ice flux divergence for the years 2007 and 2009 (f-g).

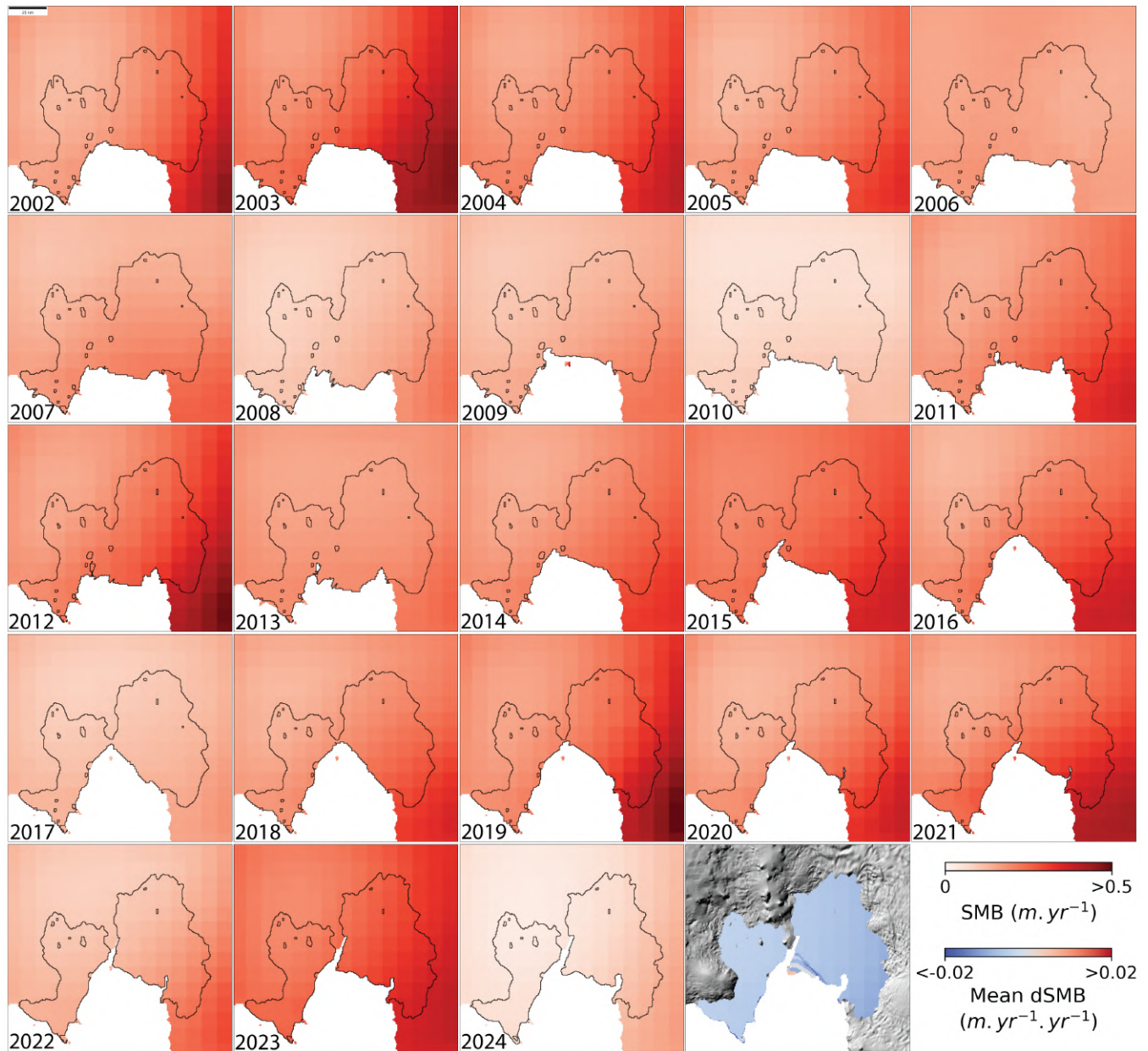


Fig. S4. All GEMB SMB maps for the 2002-2024 period, with mean annual change.

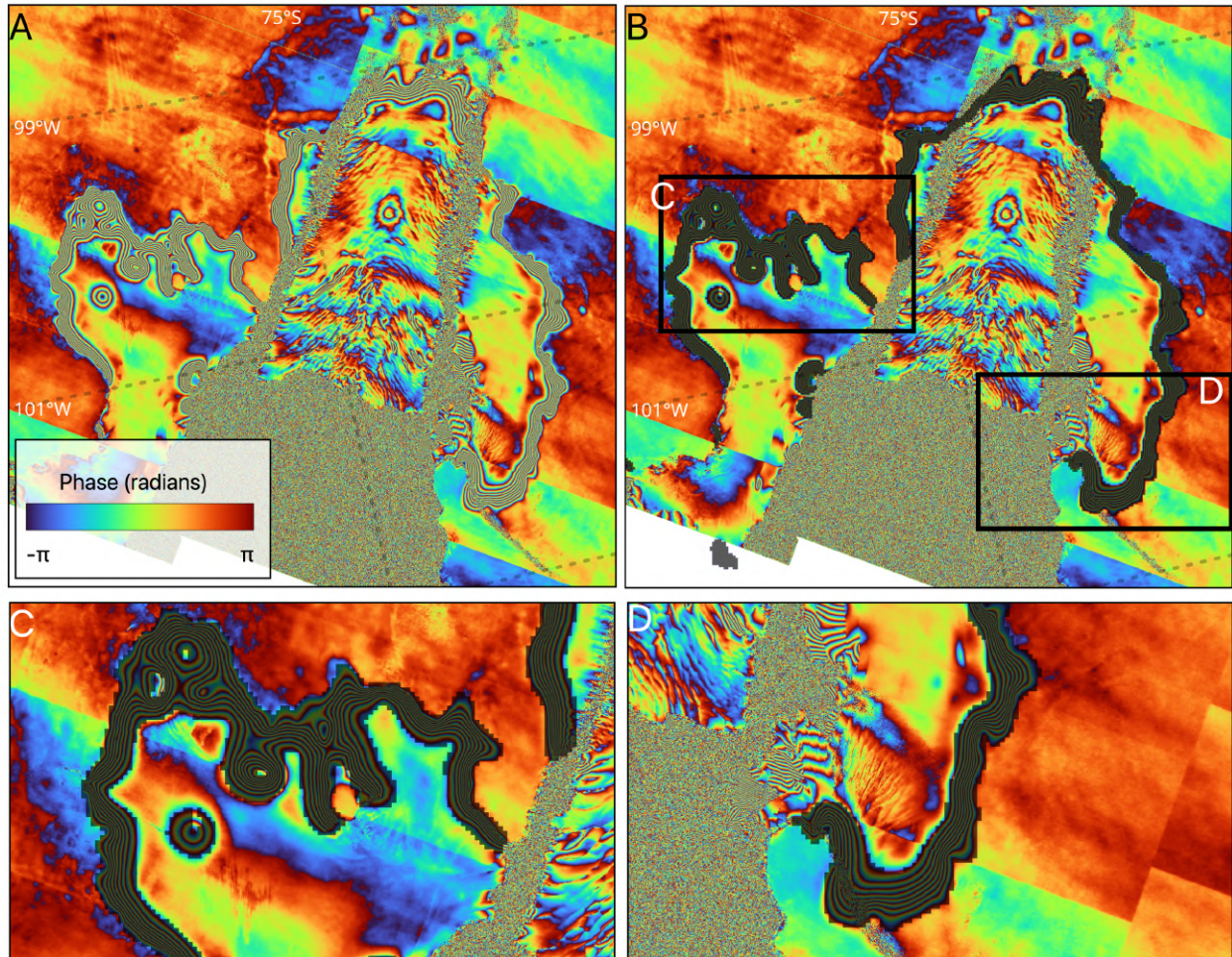


Fig. S5. Example of flexure zone mapping using QDInSAR. **A)** Double difference interferogram from 1-day Sentinel-1 imagery in 2025 (8), **B)** Flexure zone manually mapped in dark. **C** and **D** are zoom-in over different sections of the grounding line.

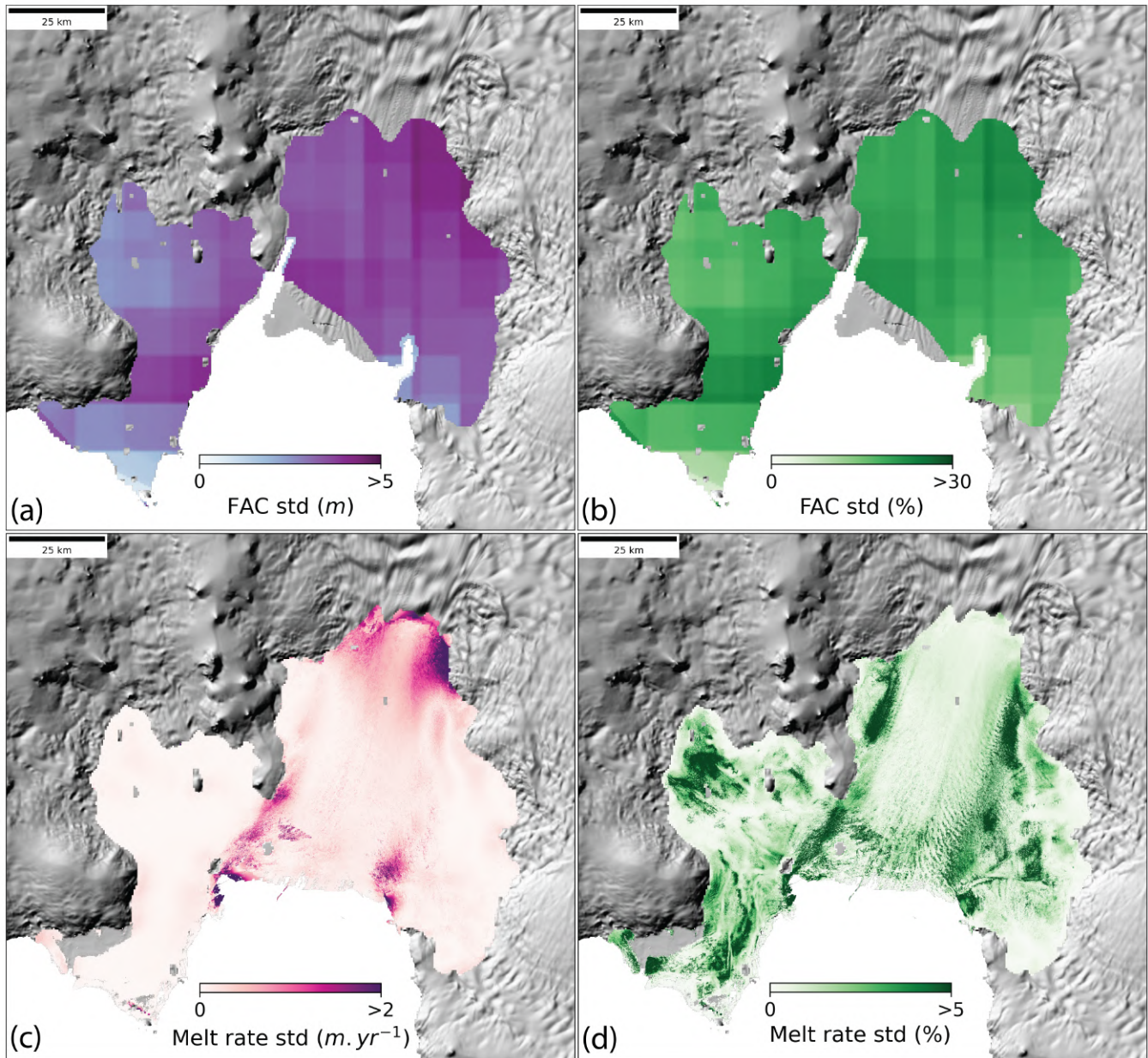


Fig. S6. Standard deviation of the three FAC products averaged over the study period, in absolute value (a) and in percentage (b). Standard deviation of the median melt rate mosaics using the three different FAC products in absolute value (c) and in percentage (d).

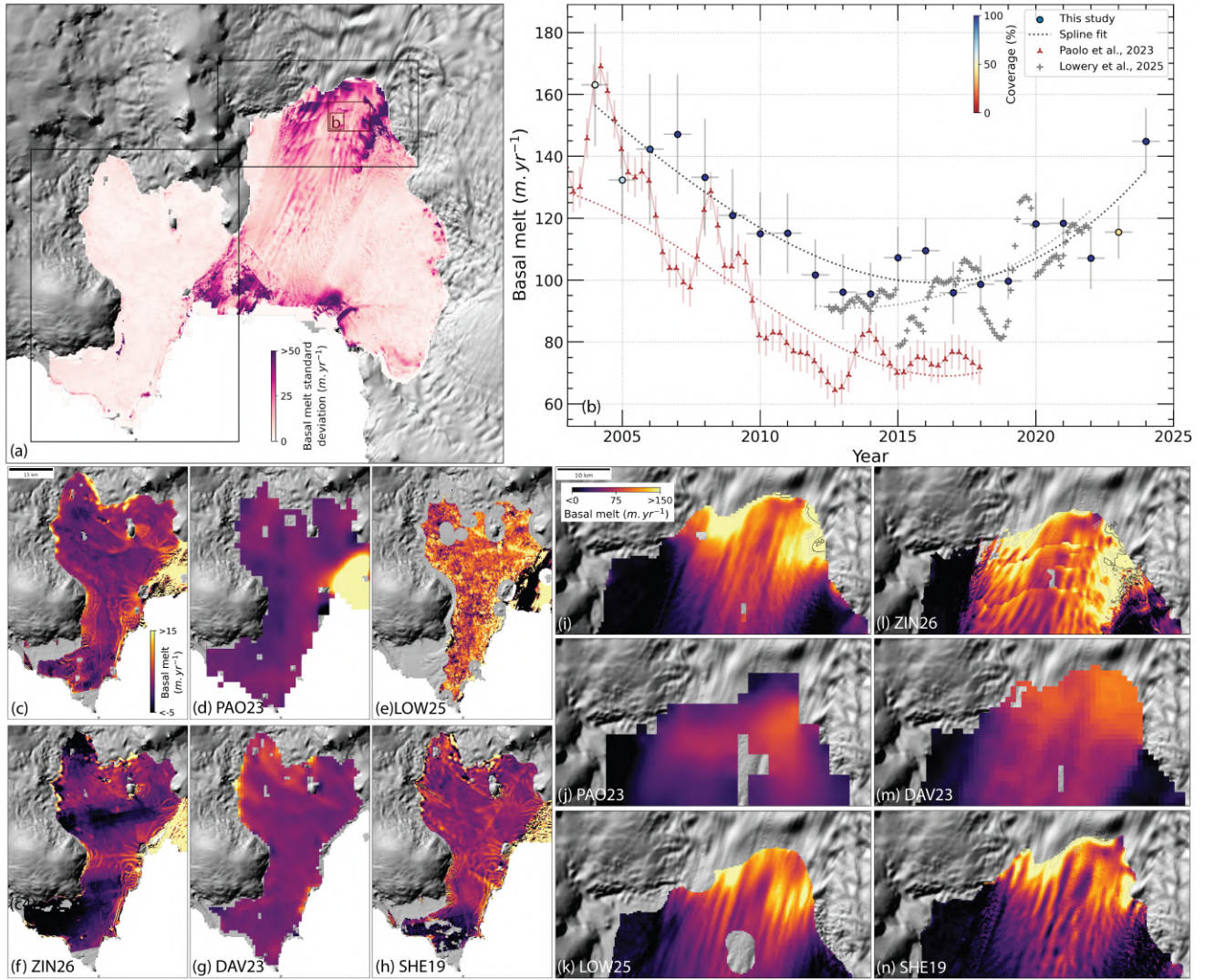


Fig. S7. (a) Standard deviation between basal melt rates mosaics from 6 different datasets. (b) Time series of melt rates averaged inside the burgundy box for this study, PAO23 and LOW25 with errors represented as vertical and horizontal bars. Circles are color-coded according to the percentage of spatial coverage of the melt maps inside the box. Inset maps of the northern tributary for this study (c), PAO23 (d), LOW25 (e), ZIN26 (f), DAV23 (g) and SHE19 (h). Inset maps in the grounding line region are also represented in (i-n). For comparison purposes, median mosaics from this study, PAO23 and LOW25 shown in this figure are computed over the 2010 - 2017 time period. In the same effort, ZIN26 mosaic is advected back to the 2015 grounding line position using ice flow velocities as in (5), and we used the initial pixel mosaic from (4).

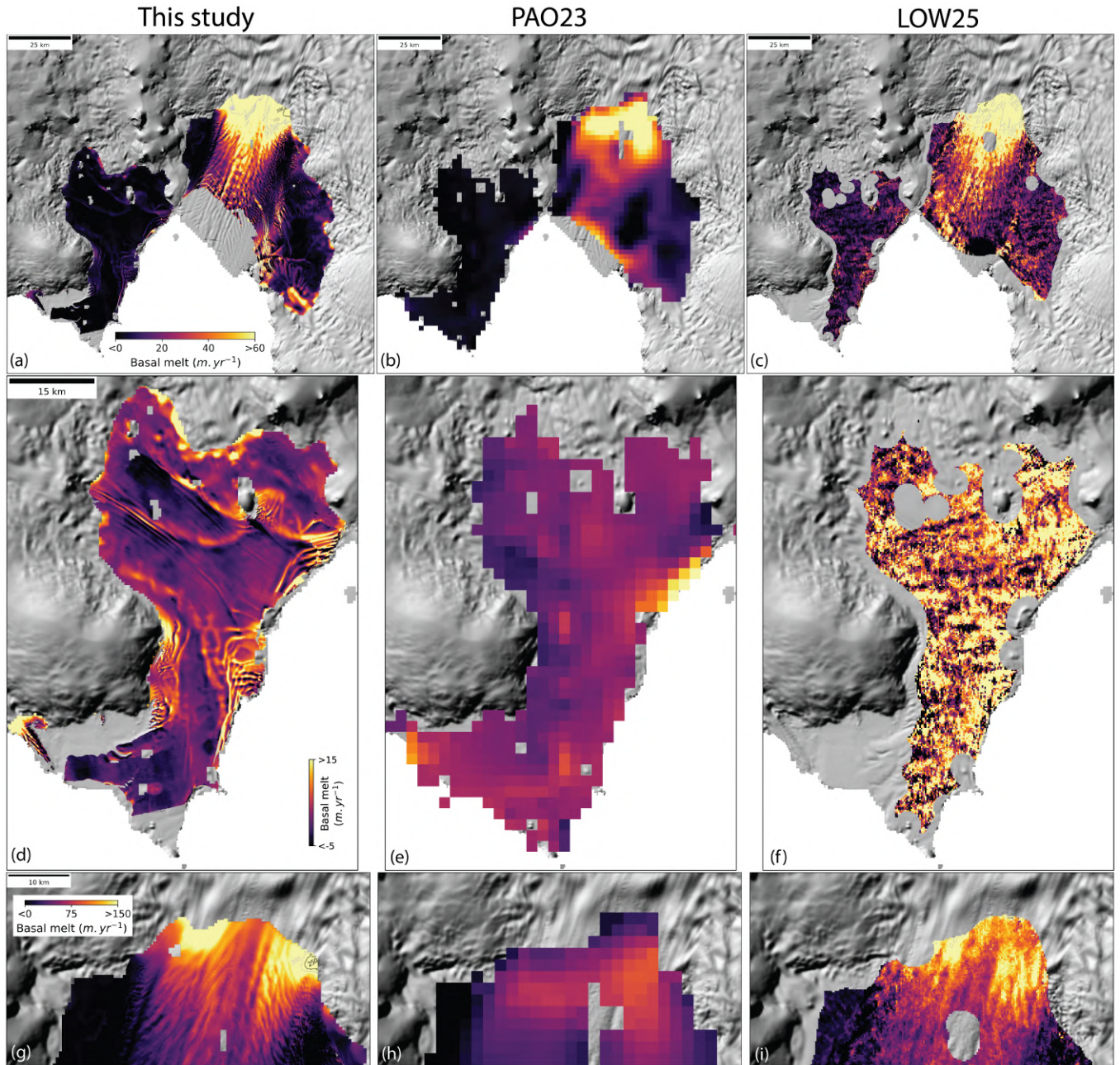


Fig. S8. Comparison of melt rate in 2017 at the scale of the ice shelf (a-b-c), the northern tributary (d-e-f) and the grounding line (g-h-i) between our study, PAO23 and LOW25.

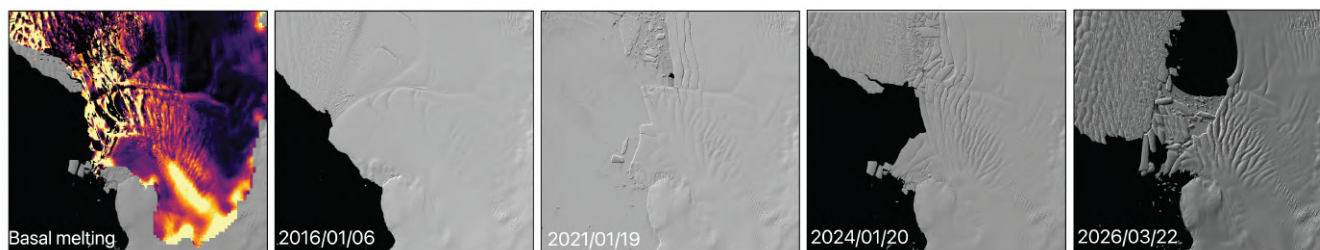


Fig. S9. Zoom on the southern tributary: basal melt from 2018-2024 and the evolution of damage from 2016 to 2026 with imagery from USGS's Landsat-8/9.

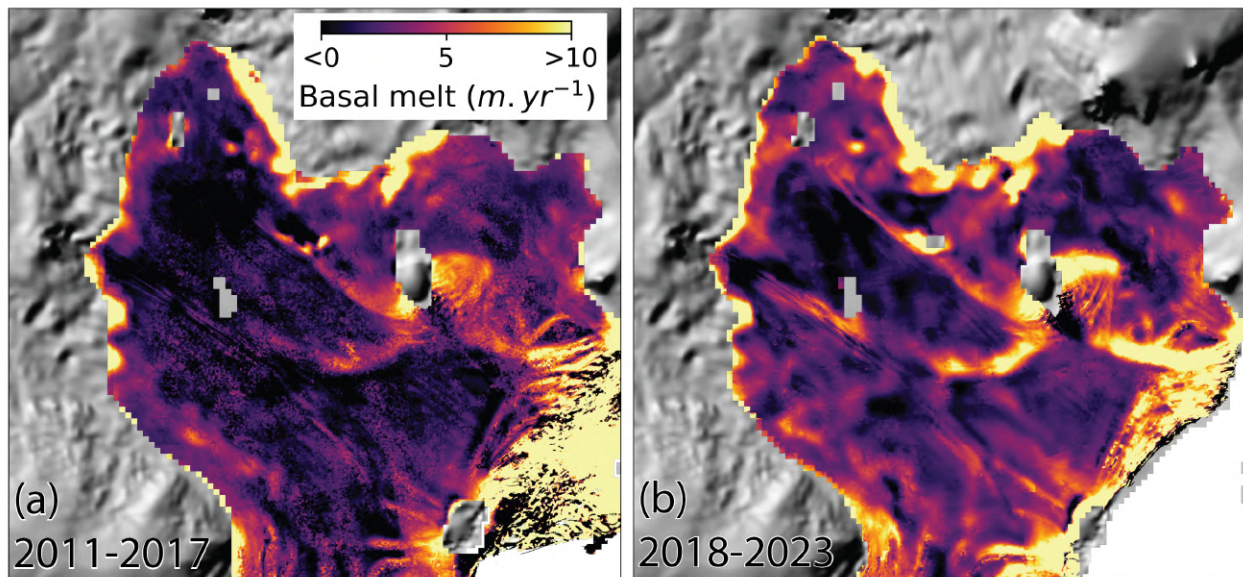


Fig. S10. Comparison of melt rate between the 2011-2017 (a) and 2018-2024 (b) sub-periods, zoomed on the ice shelf northern tributary.

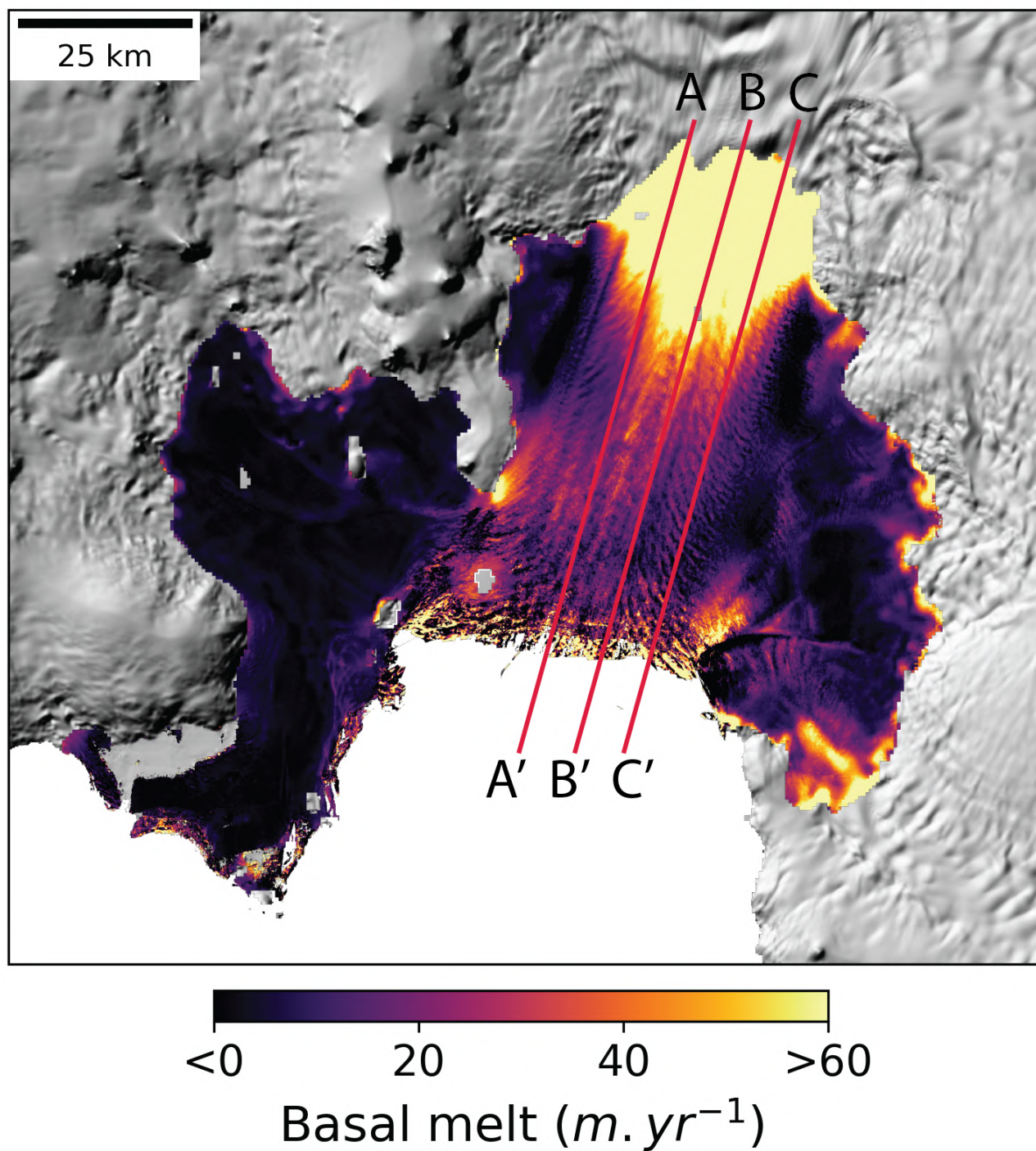


Fig. S11. Locations of various profiles along PIG ice shelf main trunk, overlaid on the 2004-2024 basal melt map.

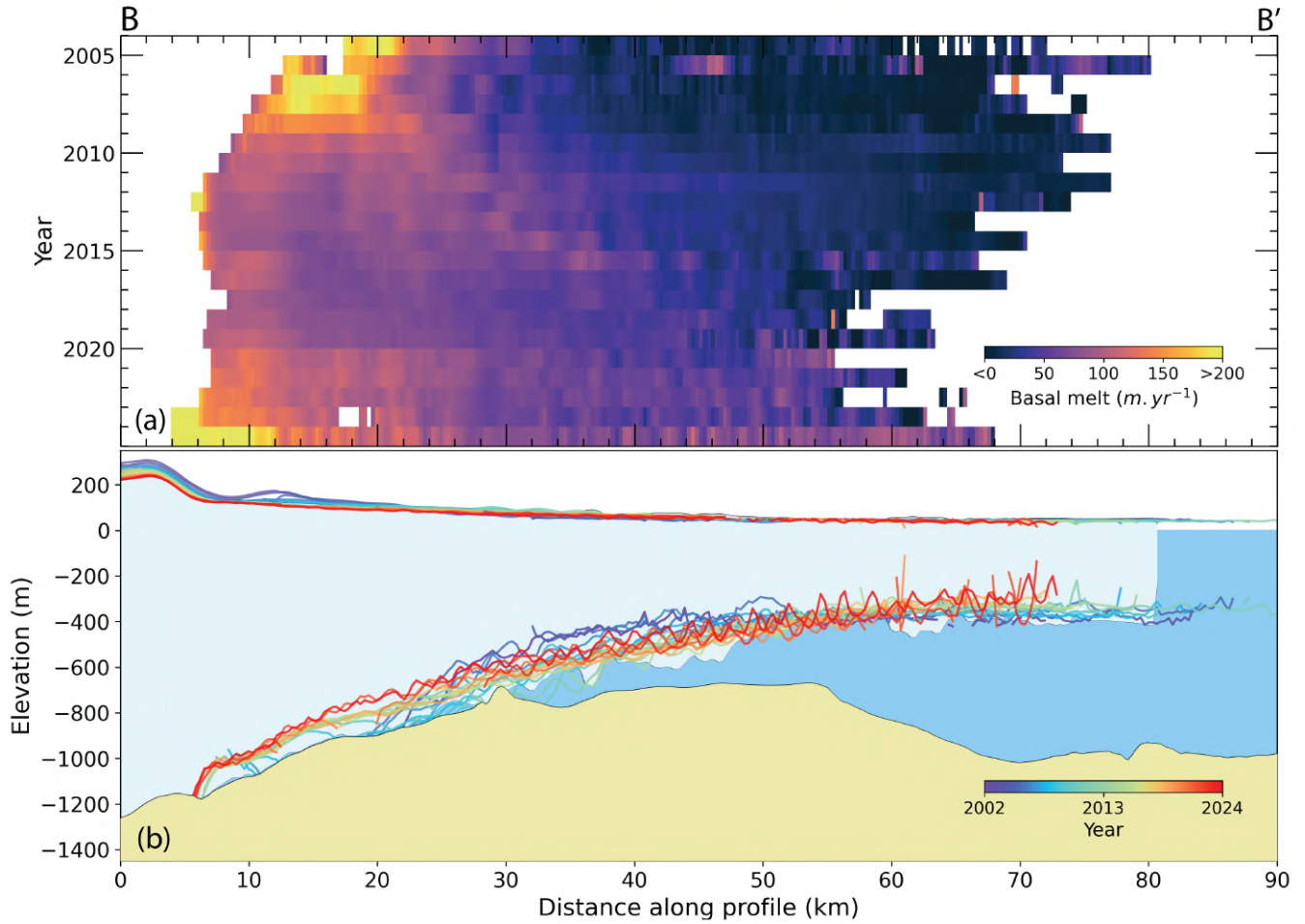


Fig. S12. (a) Melt rate evolution averaged on a 2 km width along profile BB' displayed in Fig. S11, and (b) ice thicknesses evolution along the same profile (with background thicknesses and bedrock from BedMachine (9)).

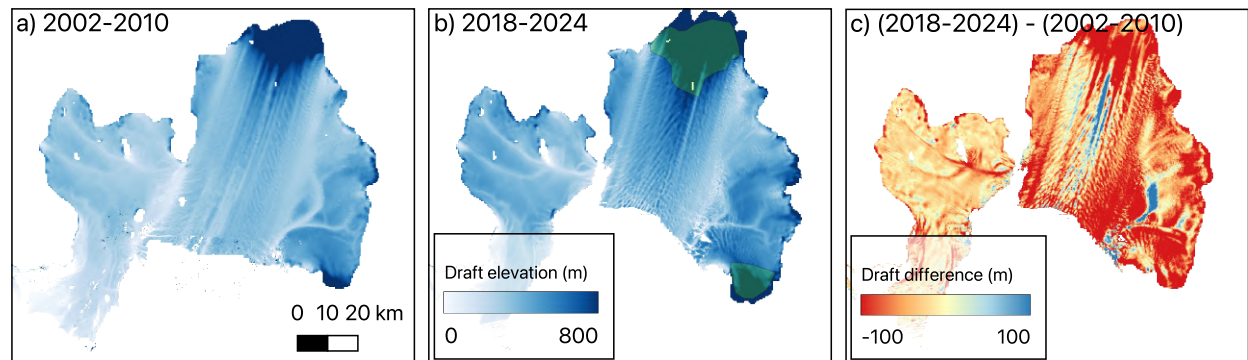


Fig. S14. Average ice shelf draft calculated over the period 2002-2010 (**a**), and 2018-2024 (**b**). Deep areas in (**b**) are highlighted in green. Panel (**c**) is the difference in draft elevation, illustrating vertical changes in water column thickness between the two periods over common areas.

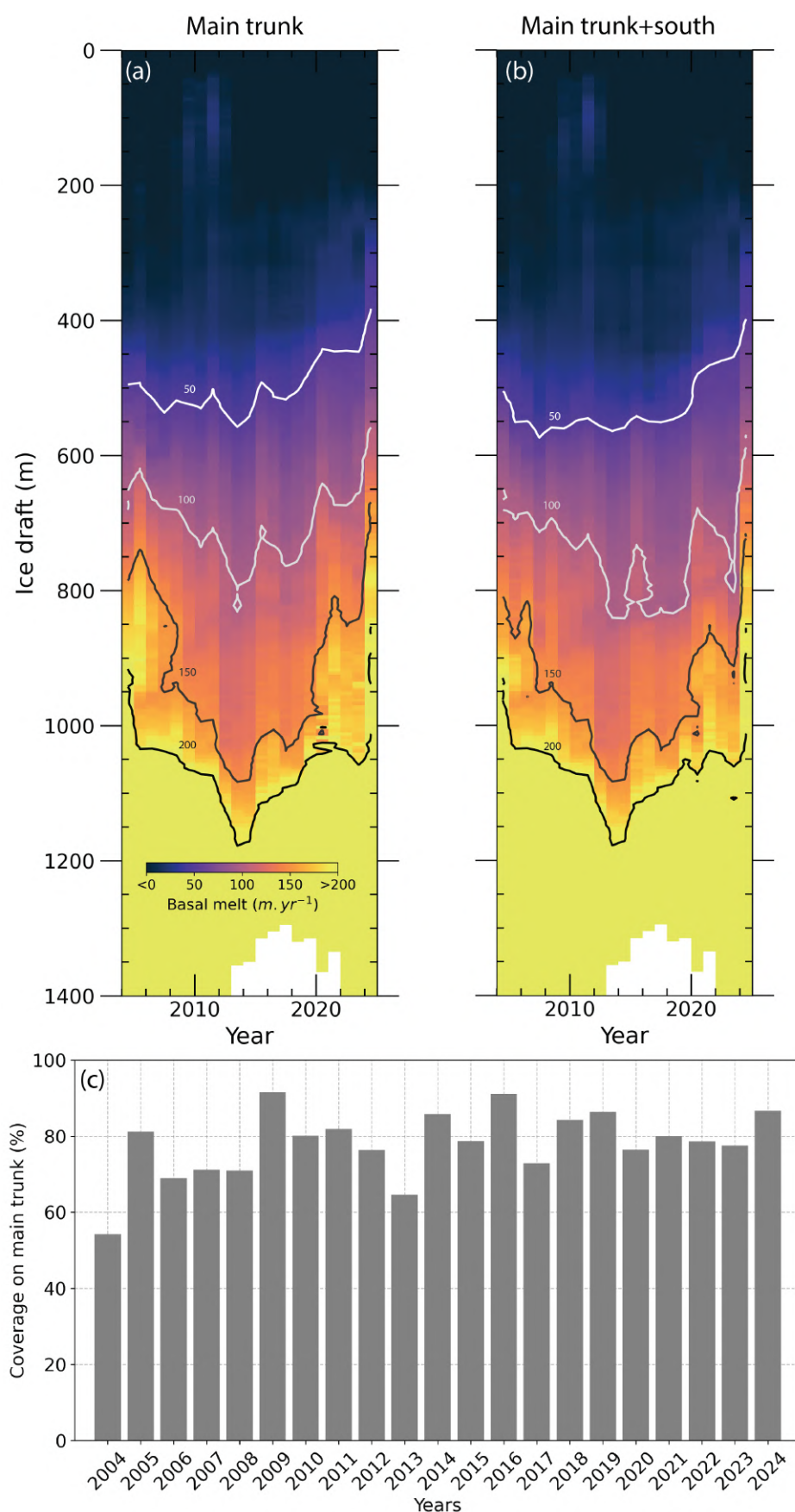


Fig. S15. (a) Hovmöller diagram of melt rate evolution as a function of binned ice shelf draft with contour lines plotted at 50, 100, 150 and 200 $m.yr^{-1}$ only on PIG ice shelf main trunk. (b) is the same as (a) but also including the south tributary of the ice shelf. Data coverage on the main trunk is also plotted as a function of time (c)

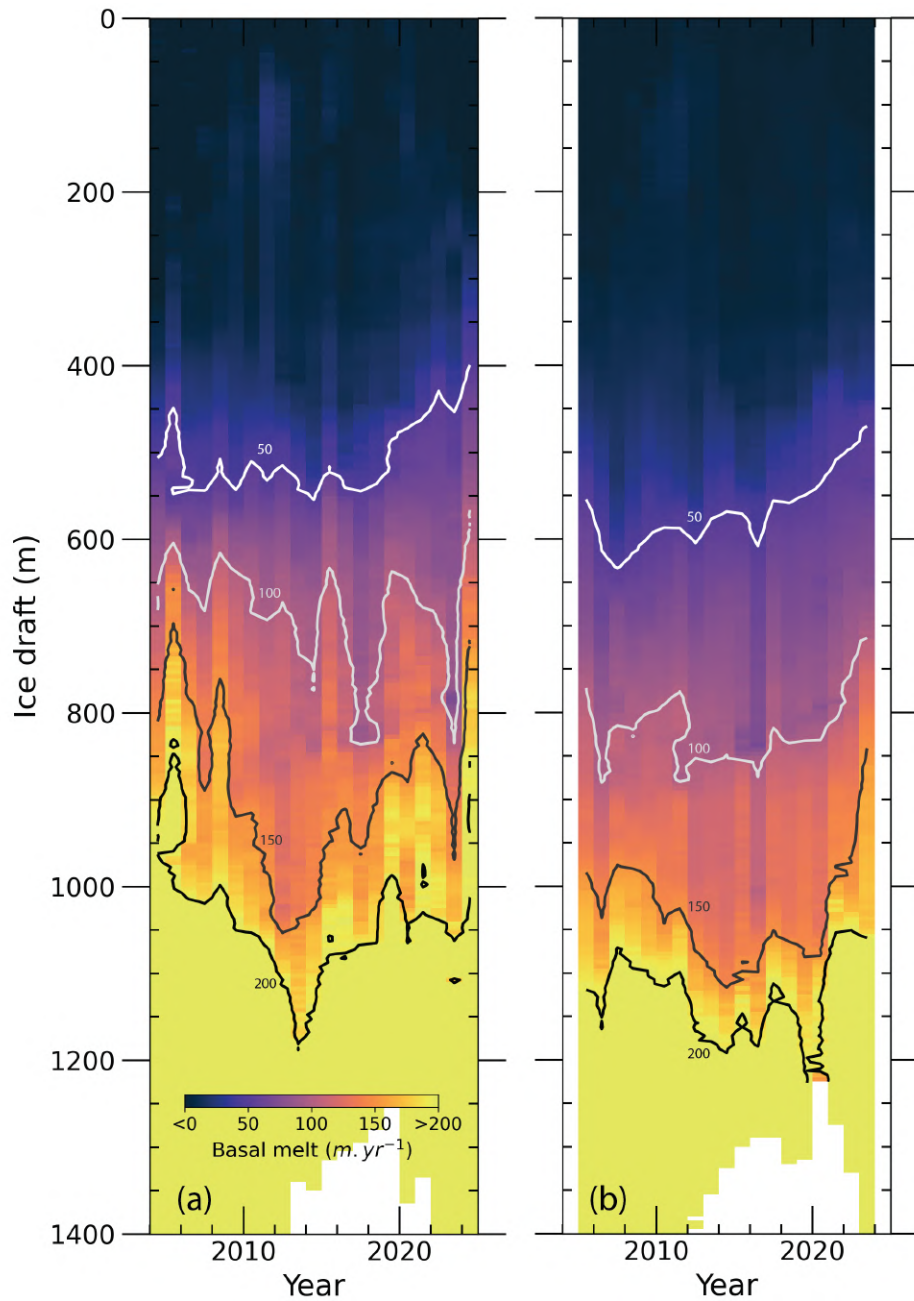


Fig. S16. Hovmöller diagram of melt rates evolution as a function of binned ice shelf draft with contour lines plotted at 50, 100, 150 and 200 $m.yr^{-1}$ computed from (a) short (1-3 years) and (b) long (4-6 years) temporal cycles.

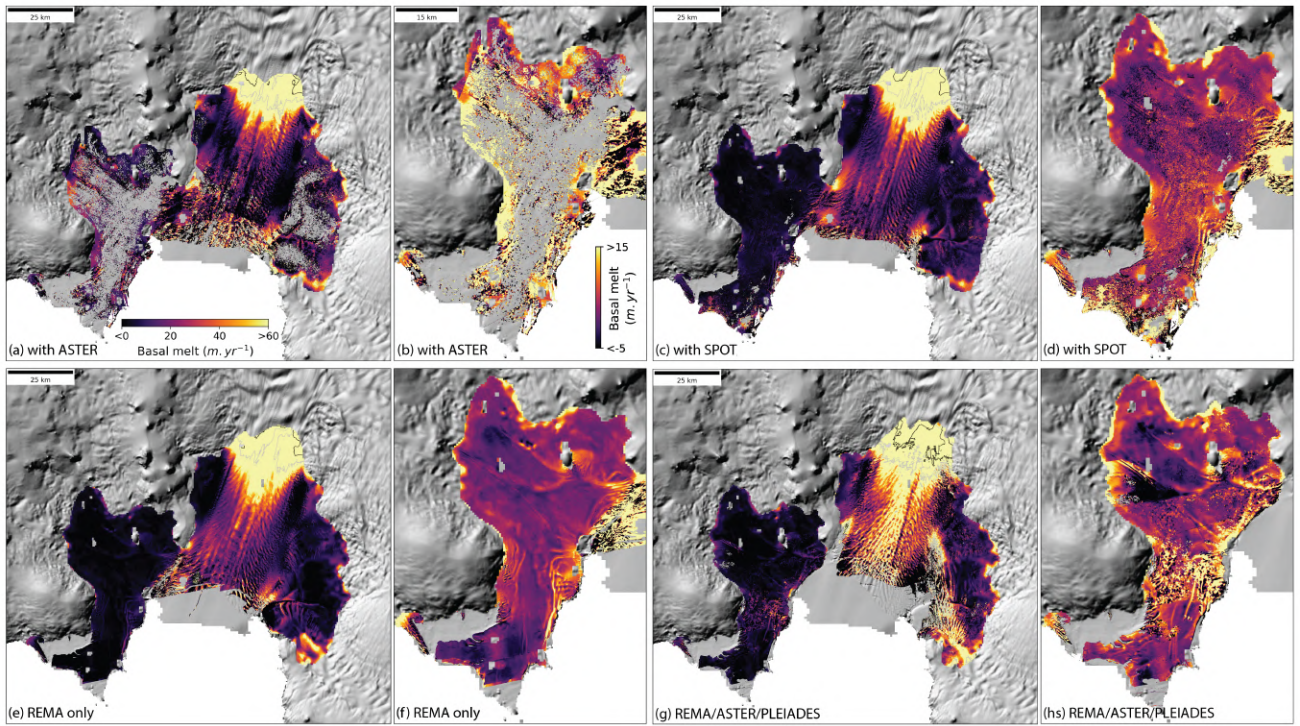


Fig. S17. Comparison of melt rates resolution at the scale of the ice shelf and of the northern tributary computed with different DEM sources: with ASTER DEMs (**a-b**), with SPOT DEMs (**c-d**), only using REMA DEMs (**e-f**) or REMA DEMs filled with ASTER data (**g-h**). Grey and black scale contours at 100 and 200 m.yr^{-1} .

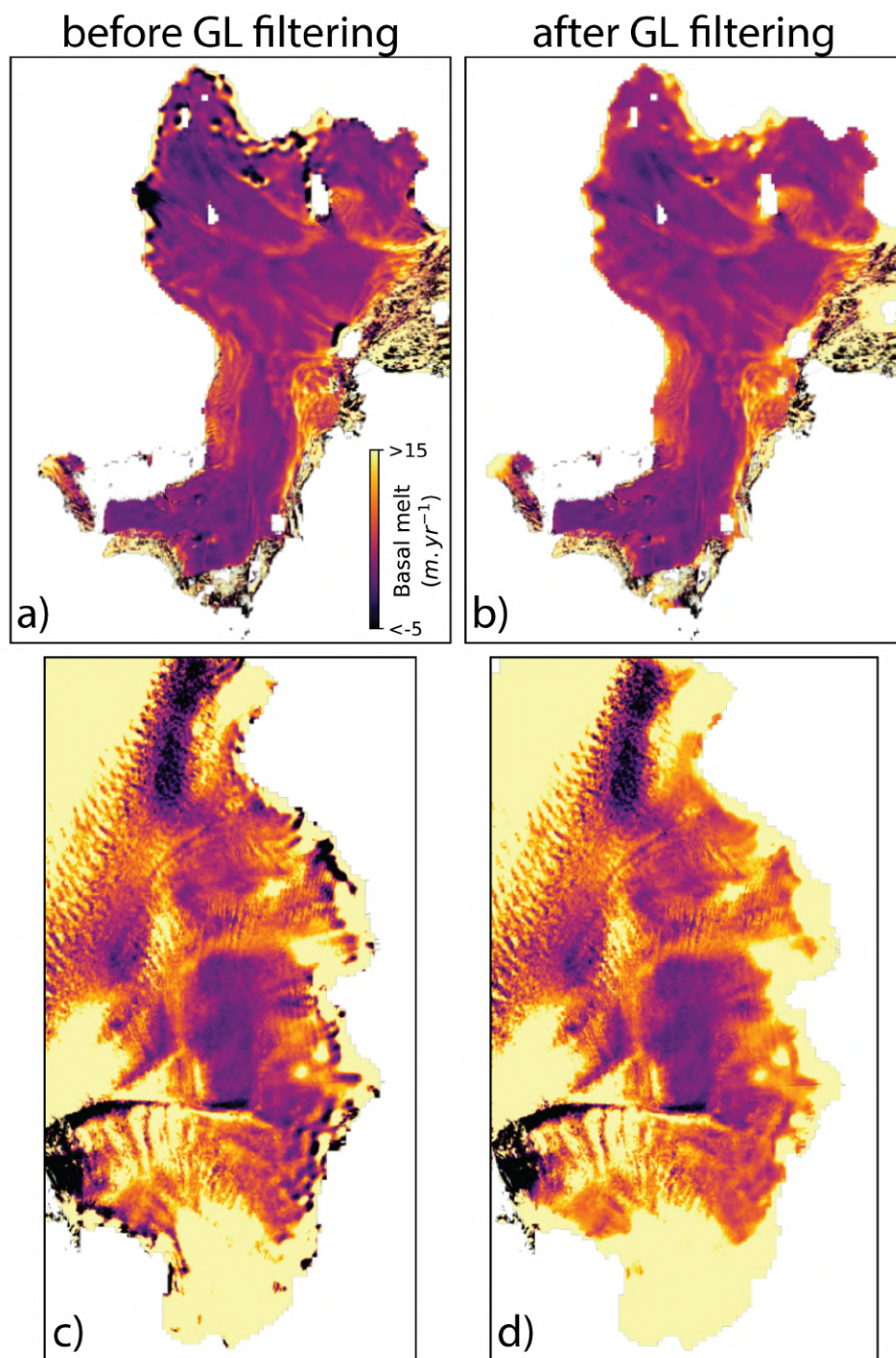


Fig. S18. Comparison of basal melt median mosaic (2004-2024) before and after applying the GL filter on the northern (a-b) and southern (c-d) tributaries.

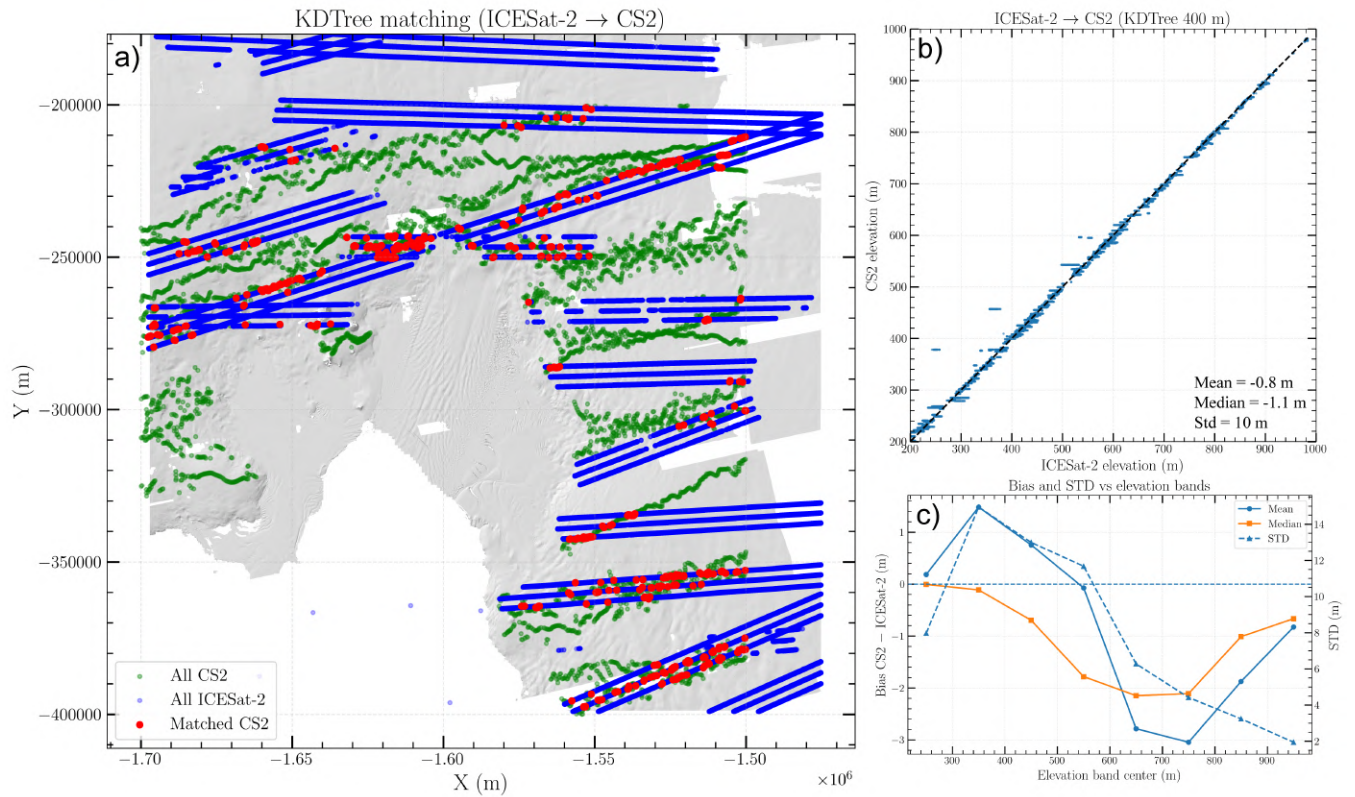


Fig. S19. Comparison of Cryosat-2 (CS2, CryoTEMPO-EOLIS Thematic Point Product) Point-of-closest-approach (POCA) data to ICESat-2 ATL06 measurements both acquired in February 2025. **a)** Location of measurements on Pine island glacier from CS2 (green point), ICESat-2 (blue point), and CS2 data within 400 meters of ICESat-2 (red). **b)** Scatter plot of CS2 elevation against ICESat-2 elevation on grounded ice (above 200 m elevation). Inset text shows the statistics of the difference between CS2 and ICESat-2. **c)** Evolution of statistics of the difference by elevation bands.

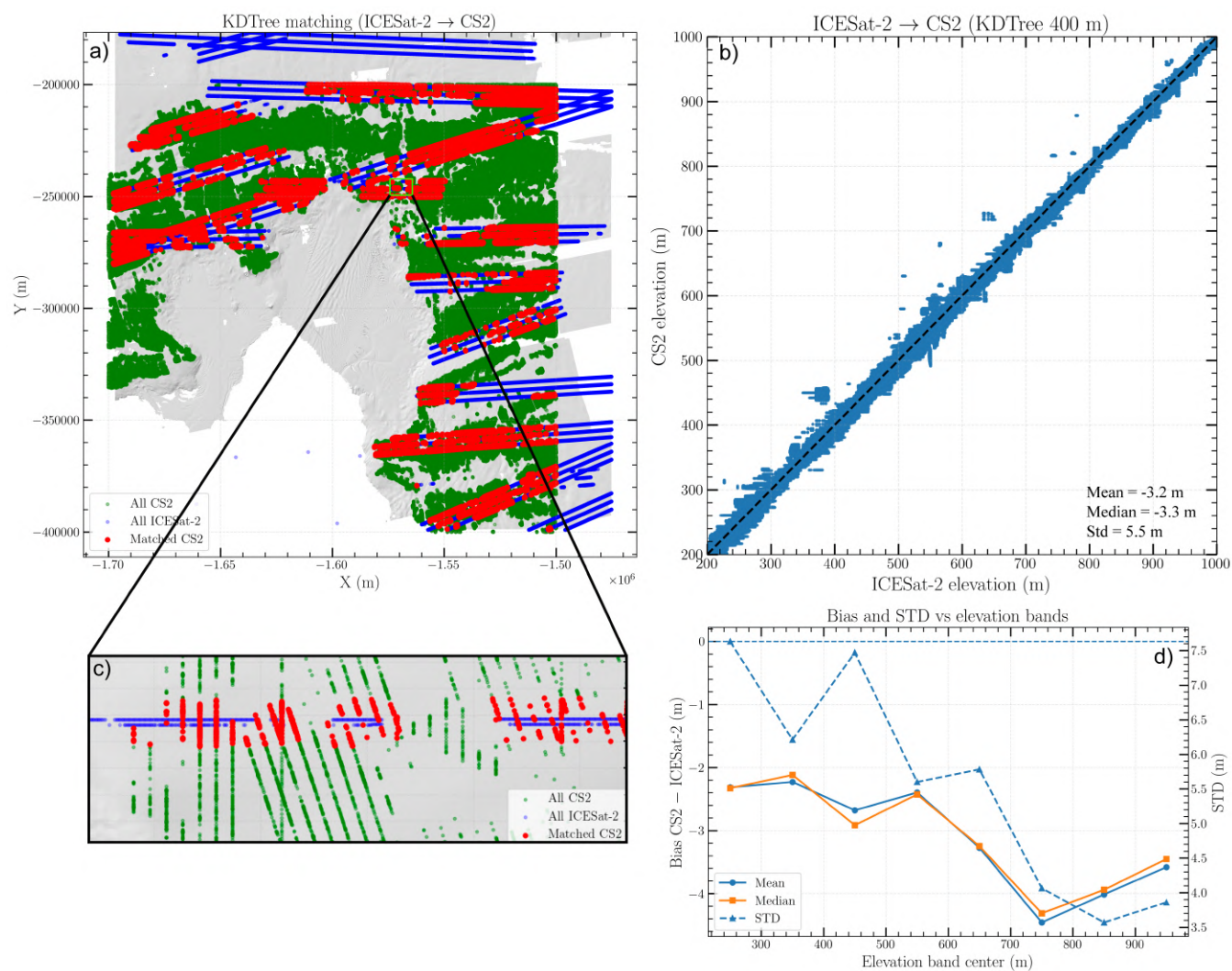


Fig. S20. Comparison of Cryosat-2 (CS2, CryoTEMPO-EOLIS Thematic Point Product) swath data to ICESat-2 ATL06 measurements both acquired in February 2025. **a)** Location of measurements on Pine Island glacier from CS2 (green point), ICESat-2 (blue point), and CS2 data within 400 meters of ICESat-2 (red). Panel **c)** is a zoom on a narrow section on grounded ice. **b)** Scatter plot of CS2 elevation against ICESat-2 elevation on grounded ice (above 200 m elevation). Inset text shows the statistics of the difference between CS2 and ICESat-2. **d)** Evolution of statistics of the difference by elevation bands.

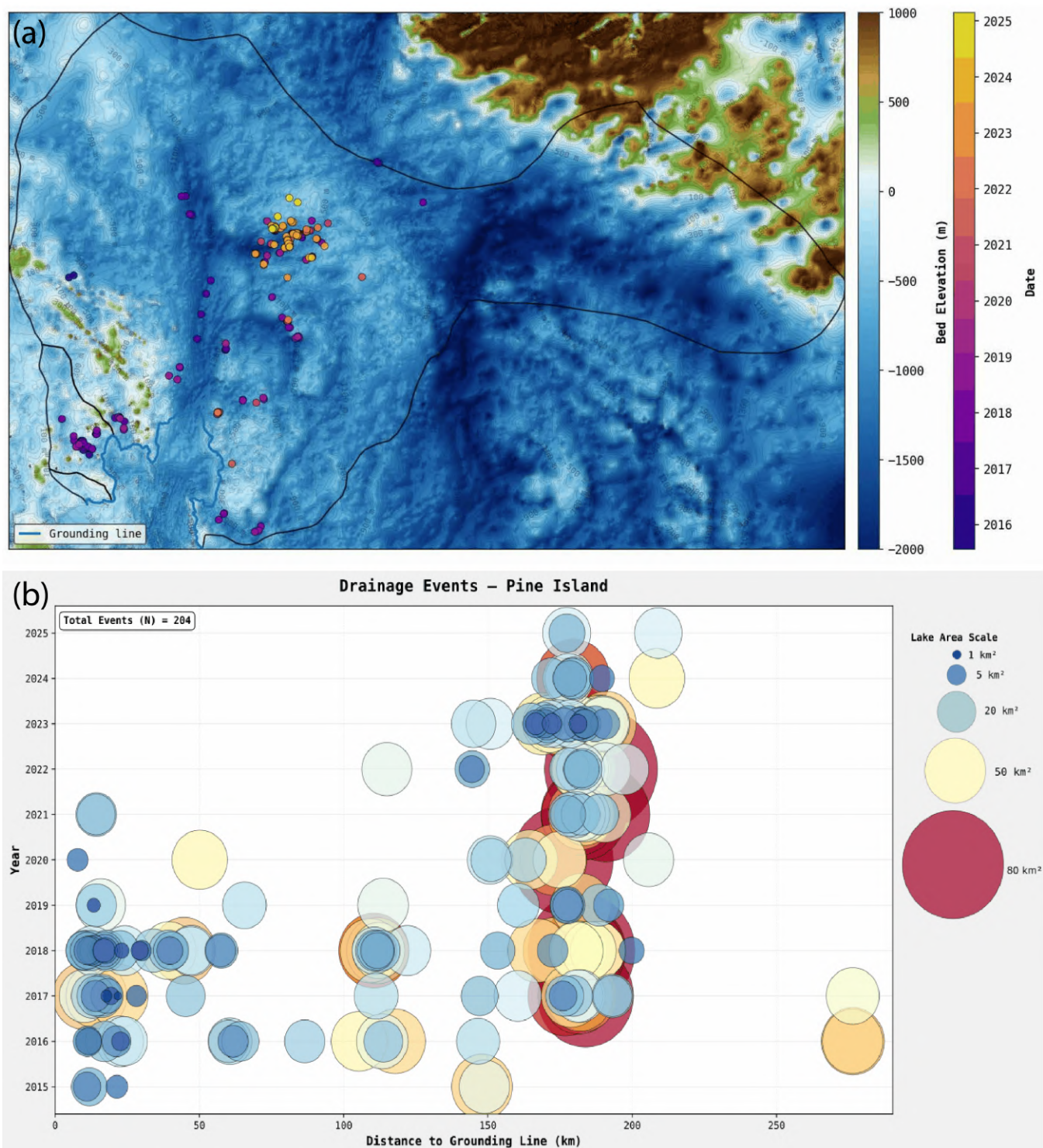


Fig. S21. (a) Localization of subglacial drainage events detected in Pine Island drainage basin using SAR interferometry, overlaid on bedrock topography and (b) drainage events date and area as a function of the distance to the grounding line

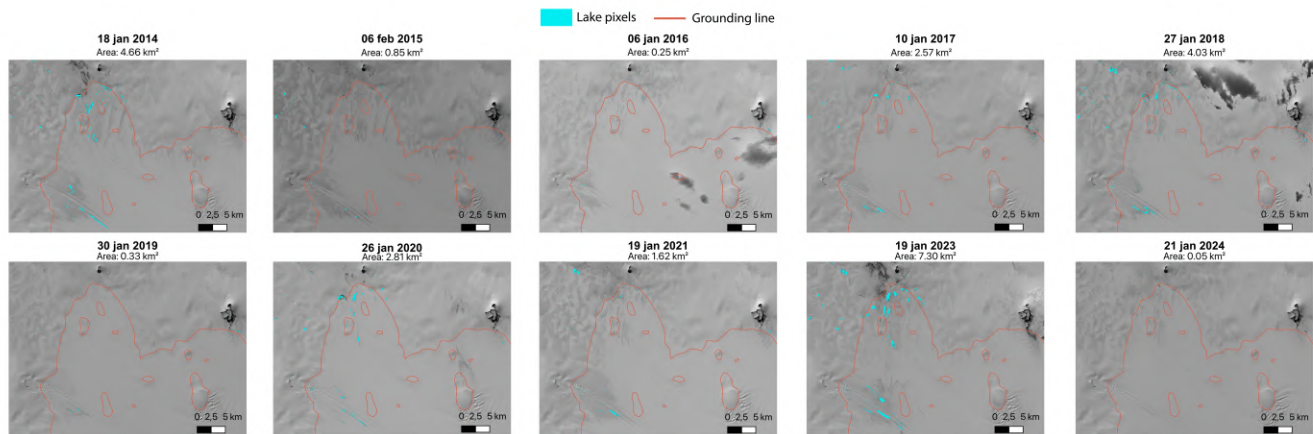


Fig. S22. Multi-annual detection of supra-glacial lakes on PIG ice shelf between 2014 and 2024 using classification of Landsat-8/9 optical imagery (10).

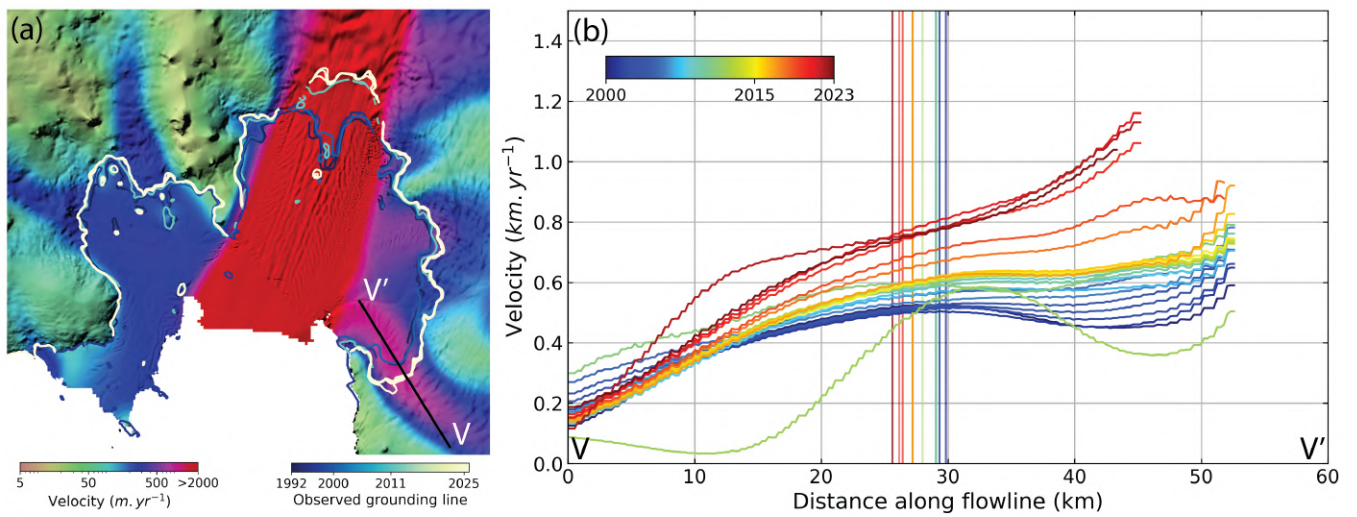


Fig. S23. Average ice velocity during 2000-2024 overlaid on the 200m REMA mosaic, with observed grounding lines (a). Average velocity profile along flowline VV', color-coded by years, with vertical lines representing the grounding line positions (b).

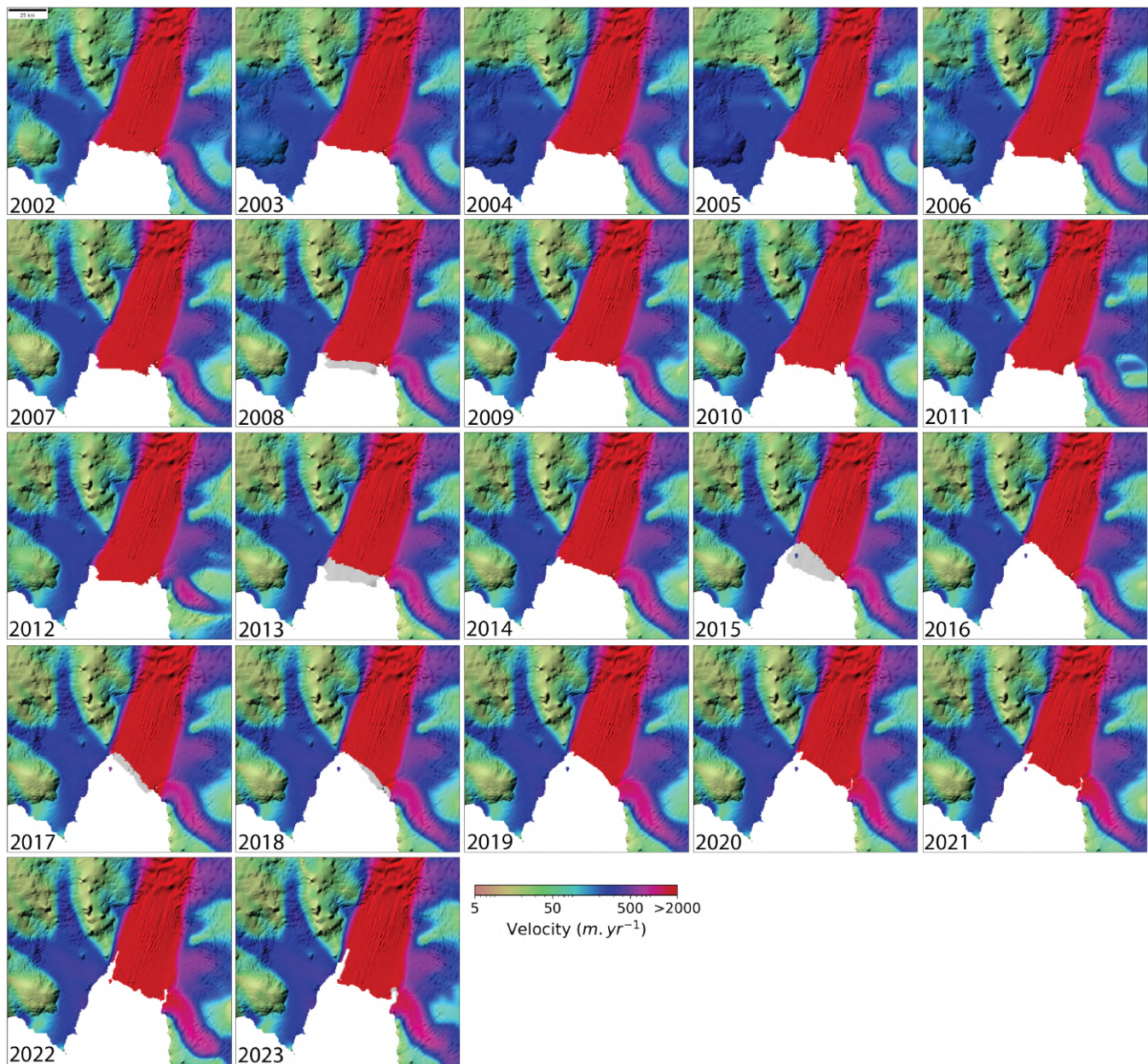
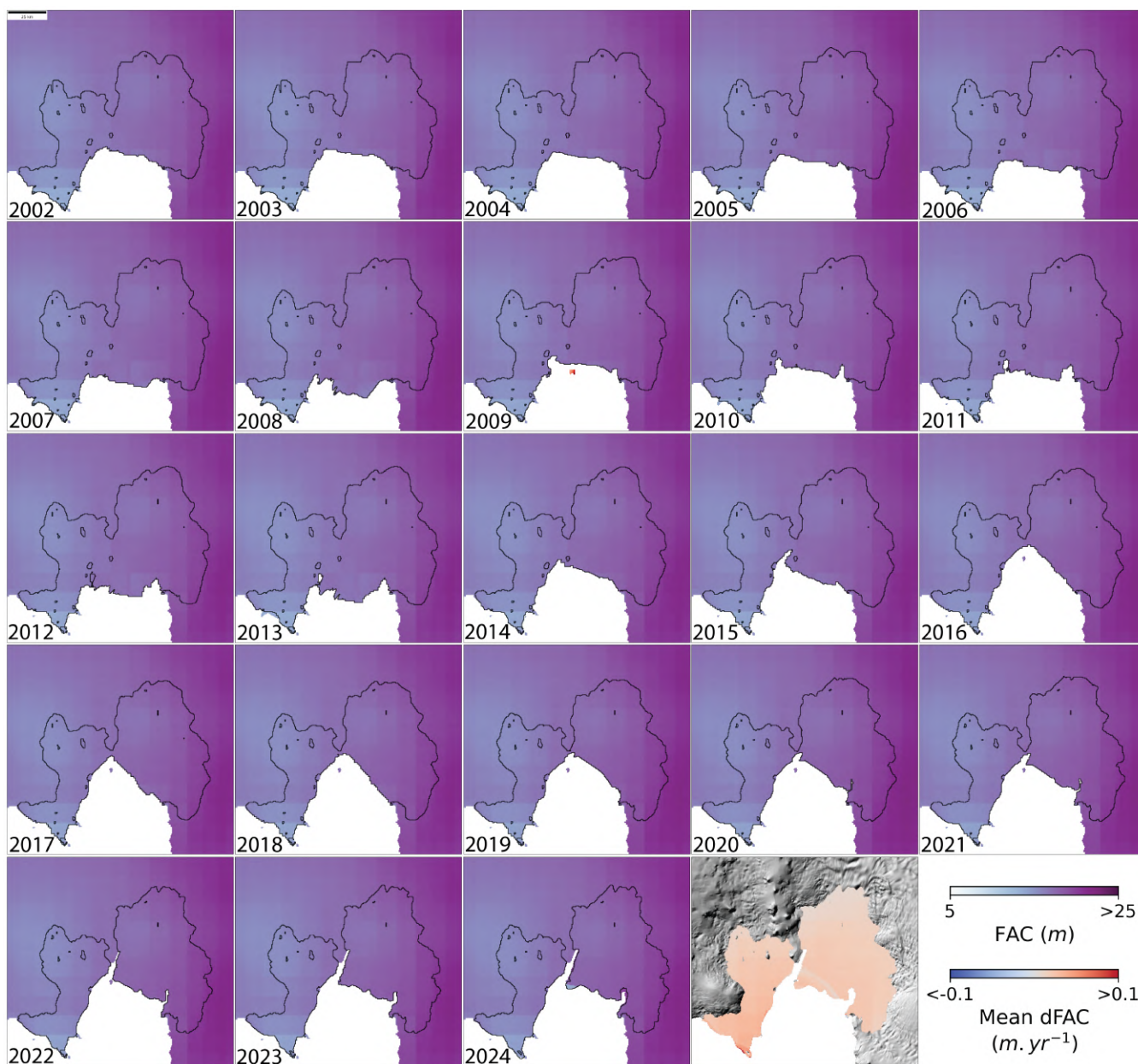


Fig. S24. All processed ice velocities ($\sqrt{v_x^2 + v_y^2}$) used for the 2002-2024 period.



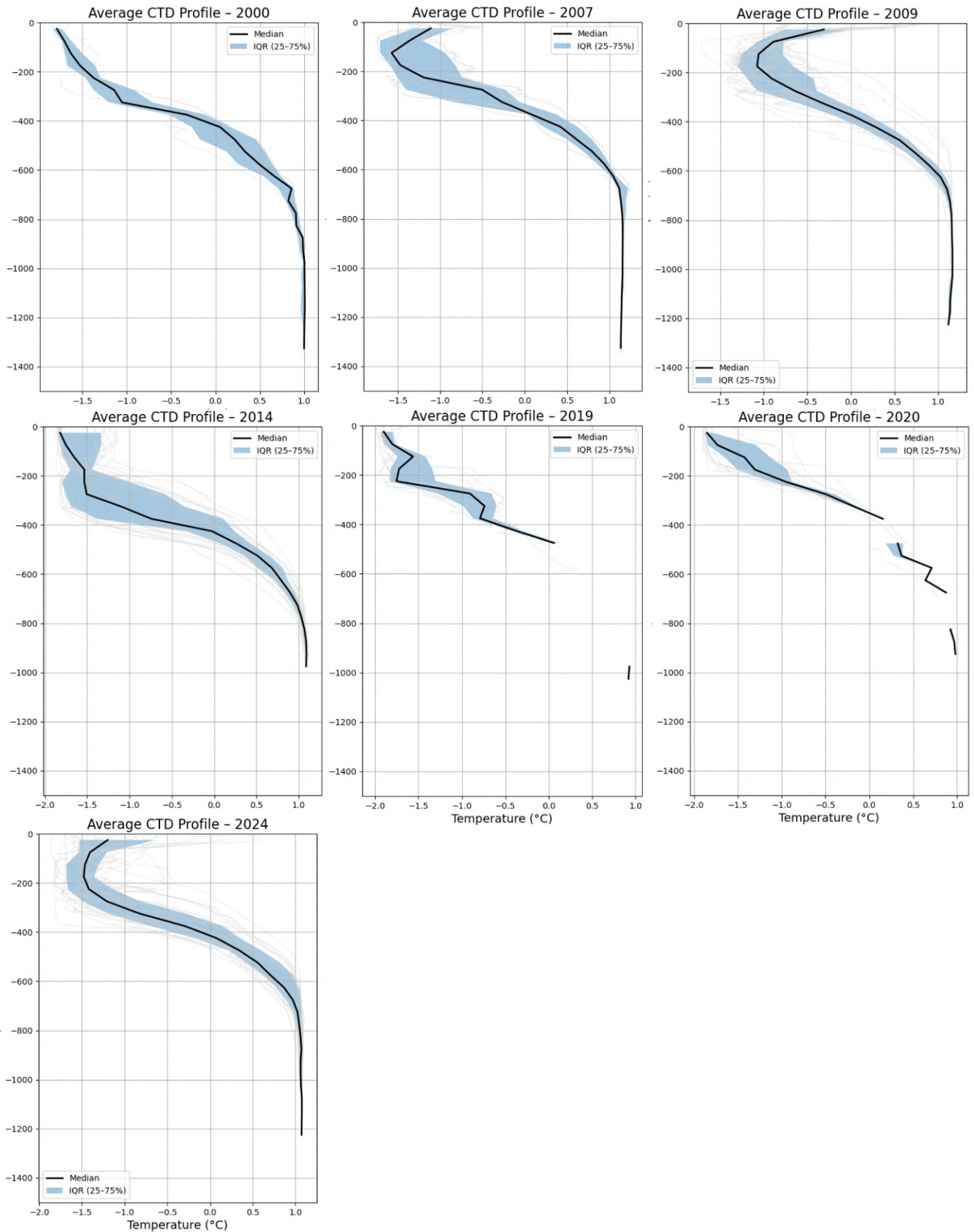


Fig. S26. Conductivity–Temperature–Depth (CTD) measurements in PIG Bay for each year. The light gray lines show the evolution of temperature with depth for each individual cast. The dark black line represents the median temperature profile, computed in 10 m depth bins, with the shaded area indicating the interquartile range (25–75%).