

Sea-level contribution from Thwaites Glacier doubles in an observation-constrained ice-sheet model

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An approach to reducing uncertainty in sea-level projections is to calibrate ice-sheet models against observations^{9,15–18}, which now span more than four decades¹. This strategy is based on the premise that recent trends strongly influence near-future glacier evolution¹⁹ and that the reliability of projections should be measured by how well they match the modern observation record. High-resolution satellite observations from the US–Finnish Synthetic Aperture Radar (SAR) constellation ICEYE have recently revealed kilometer-scale fluctuations in grounding-line position along the main trunk of Thwaites Glacier at tidal frequencies²⁰. These observations (Fig. 1b) indicate that seawater penetrates several kilometers beneath grounded ice, episodically lifting it during tidal cycles and forming a dynamic “grounding zone” characterized by active flushing of pressurized, warm Circumpolar Deep Water (CDW). Past pioneering modeling studies suggested that kilometer-scale seawater intrusions at the grounding zone could substantially enhance ice mass loss^{21–25}. Yet, many ice-sheet models treat the grounding line as a discrete boundary between grounded and floating ice and do not represent a dynamic, actively melting grounding zone. In addition, commonly used ocean-induced melt parameterizations are based on simplified plume models in which the melt rates, in many implementations, decrease toward zero at the grounding line by construction²⁶. In contrast, recent satellite observations^{3,27} and high-resolution ocean modeling²⁸ indicate that ocean-driven melt is not only substantial within the grounding zone, it may even be a dominant factor in glacier dynamics²⁴.

Here, we quantify the contribution of melt induced by seawater intrusions beneath grounded ice to the historical retreat of Thwaites Glacier from 1996 to 2023 using the Ice-sheet and Sea-level System Model (ISSM)²⁹. We implement actively melting grounding zones within ISSM (Fig. 1c) building on past numerical frameworks^{22,25}. Glacier sensitivity over the historical period is evaluated across a range of grounding-zone lengths and submarine melt rates using a Bayesian calibration approach (Methods). To evaluate the model, we compare the results with three key independent observational records: (i) surface velocity³⁰, (ii) ice thickness³¹, and (iii) position of the grounding line³². Together, these independent constraints provide a stringent test of the model fidelity, enabling a complete characterization of changes in glacier state. Following calibration of the hindcast ensemble, we extend the simulations to 2100 to estimate the contribution of Thwaites Glacier to global sea-level rise under a controlled, constant climate forcing baseline (i.e. a lower-bound scenario) that isolates the role of grounding-zone physics from future ocean-warming uncertainty.

Reconstruction of historical changes in glacier state

The most extensive retreat of the Thwaites Glacier has occurred along its western and fastest-flowing sector, the Thwaites Main Trunk (TMT), which has migrated inland along a deep bathymetric channel at a rate of approximately 800 m yr^{−1}, or 25 km since 1996 (Fig. 1b, Supplementary Fig. 1). Currently, the glacier remains pinned on the Mouginit Ridge, a bathymetric high located roughly 950 m below sea level²⁰. High-resolution ICEYE SAR observations have shown recurrent, albeit temporary, ungrounding extending up to 10 km inland of Mouginit Ridge (Fig. 1b, Supplementary Fig. 1), indicating recurrent kilometer-scale seawater intrusions beneath grounded ice. In contrast, the eastern sector of the glacier, the Thwaites Eastern Ice Shelf (TEIS), has undergone a slower and more spatially heterogeneous retreat at a rate of 400–700 m yr^{−1} since 1996, corresponding to 10–20 km^{3,32}.

Using ISSM initialized to a state representative of the mid-1990s (Fig. 1a, Supplementary Fig. 2–3, Methods), we simulate the evolution of the Thwaites Glacier until 2023 (Fig. 2). Ice flow is governed by a regularized Coulomb friction law, which best reproduces the acceleration of Pine Island Glacier^{12,33}, located 250 km north of Thwaites. In this configuration, we explicitly represent an actively melting grounding zone (Fig. 1C, Methods) that allows ocean-driven basal melt to occur inland of the flotation

line calculated by the model, consistent with satellite observations^{20,28,34}. Model results indicate that the rate of grounding-line retreat is strongly controlled by both the basal melt rate beneath floating ice, consistent with previous studies¹⁰, and the length of seawater intrusions into the grounding zone (Fig. 2). Overall, higher melt rates and longer intrusions produce a greater retreat, as shown in past work^{22,25}.

However, the model systematically overestimates the retreat of TEIS by 15 km, a bias also reported in previous modeling studies^{8,10,35} using the state-of-the-art bed topography² (Extended Data Fig. 1). When initializing the model to reconstruct the glacier state in 1996, we find that seafloor elevations under floating ice are too deep to be consistent with the grounding-line location calculated via flotation. Within the uncertainty bounds of bed topography², we apply a localized +100 m correction in regions constrained by gravity inversion to satisfy the flotation consistency with the observed 1996 grounding line (Extended Data Fig. 1). With this correction in place, the model reproduces a rate of grounding-line retreat, glacier thinning, and ice-flow acceleration even when the grounding line is treated as a discrete boundary between grounded and floating ice, as in traditional configurations. In this configuration, the model requires unrealistically high submarine melt rates^{27,28} of 150 m yr⁻¹ to align with observations between 1996 and 2004. However, these results still underestimates discharge and thinning observed in the mid 2000s, the glacier acceleration (Fig. 3) as well as the grounding-line retreat (Fig. 2).

Explicitly resolving kilometer-scale grounding zones enables agreement with all observational constraints over the 25-year record. Using a grounding zone of 1 km in length, the model captures the concurrent retreat of both TEIS and TMT (Fig. 2D2) over 25 years, but with a relatively high submarine melt rate of 150 m yr⁻¹. For a 3-km grounding zone, the ensemble simulations fully reproduce the concurrent retreat of TEIS and TMT at a submarine melt rate of 120 m yr⁻¹ (Fig. 2C3), and capture the associated ice discharge at 100 m yr⁻¹ (Fig. 3A). Moreover, a melt rate of 100-120 m yr⁻¹, which is consistent with satellite observations (Extended Data Fig. 2) and ocean modeling²⁸, yields modeled thinning that closely matches the 32 m of grounded-ice thinning observed over the simulated 25 years (Fig. 3B).

Overall, a Bayesian analysis (Methods) reveals a well-constrained region of joint parameter space for grounding-zone length and submarine melt rate that successfully and simultaneously reproduces the observed grounding-line retreat, ice thinning, and glacier acceleration (Fig. 4A, Extended Data Fig. 3, Supplementary Tables 1-4). The joint posterior distribution exhibits a tilted structure, indicating a compensatory relationship between the two parameters, as previously noted²⁵. The posterior distribution identifies parameter combinations that minimize misfit with historical satellite trends, with central values of grounding-zone length between 2 and 5 km and melt rate of 110 m yr⁻¹ (Fig. 4A, Extended Data Fig. 3). This structure reflects parameter trade-offs rather than a single optimal solution, underscoring the importance of jointly calibrating multiple variables to reduce parameter uncertainty¹⁷. Notably, simulations without a grounding zone are four times less likely to reproduce the historical trend than simulations with a grounding zone of a few kilometers.

Simulating the future of Thwaites Glacier

Under a lower-bound forcing scenario derived from climate conditions averaged during the historical period 1996–2023³⁶, we extend the simulations to the end of the century using the same physical configuration that successfully reproduces the observational record. We find that the retreat of the TMT is only temporarily stabilized by the Mouginit Ridge (Fig. 5A, Extended Data Fig. 4). Once unpinned, the glacier continuously retreats across predominantly retrograde seafloor until reaching the next major topographic high, South Ridge (Fig. 1B). The Mouginit Ridge therefore exerts a stabilizing influence on the present-day position of Thwaites Glacier, with the likelihood of detachment from it controlled by the

length of the grounding zone (Fig. 5A,B). Simulations that omit a grounding zone or employ a narrow 1-km length, regardless of the imposed melt rate, do not retreat beyond Mouginot Ridge by 2100 (Fig. 5A, Extended Data Fig. 4), consistently with their fourfold lower likelihood of reproducing the historical retreat (Fig. 4A).

In contrast, simulations with a 2-3 km grounding zone predict a 45% probability of unpinning from the Mouginot Ridge by the end of the century, with simulations employing 2-km to retreat beyond the ridge on average in 2088 and in 2065 for a 3-km length (Fig. 5). On the other hand, simulations consistent with recent (2020s) satellite-inferred grounding-zone extent (4–6 km) show a higher probability - approximately 60% - of crossing the ridge by 2100 (Fig. 5A). In particular, at 5-6 km, unpinning occurs as early as 2038, allowing the grounding line to reach South Ridge well before 2100. Across our model ensemble, the extent of intrusion is therefore the dominant predictor of ridge crossing.

Sea-level projections

All simulations produce sustained sea-level contributions throughout the century (Fig. 4B, Extended Data Table 1) without indication of slowdown or stabilization feedback, in agreement with previous studies^{7,8}. Notably, the envelope of our projections lies in the upper half of the uncertainty range of the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6) under the same forcing scenario³⁷. The ISMIP6 projections exhibit a large uncertainty, with a mean of 15.2 ± 21.8 mm of sea level equivalent, and some simulations even project a late-century slowdown in ice discharge³⁸. Although the large spread in the ISMIP6 envelope partly reflects the use of a single modeling framework rather than a multi-model ensemble³⁷, our results further establish calibration with historical data as a reliable path to narrowing projection uncertainty in ensemble efforts^{9,18,19}.

In these projections, we assume that there is no interannual variability in the thermal forcing of the ocean by the end of the century. Oceanographic observations collected in the Amundsen Sea Embayment since the 1990s indicate that water temperatures at 700 m depth have remained stable in the range of 1.0 to 1.2°C^{4,5,39}, while the temperature of Circumpolar Deep Water at the edge of the continental shelf does not likely exceed 1.5°C⁴. As a sensitivity test, future warming could increase ocean temperatures by 0.5 to 0.7°C compared to our model, which - assuming a linear relationship between thermal forcing and melt rates⁴⁰ - would increase basal melting by approximately 10 m yr⁻¹. We calculate that such an increase in melt would raise projections by approximately 15% (Extended Data Fig. 5) hence posit that our projections are less sensitive to moderate changes in thermal forcing than to grounding-zone extent.

The timing of unpinning from Mouginot Ridge exerts a first-order control on the contribution of Thwaites Glacier to sea level (Fig. 4,5). Adopting a melt rate of 110 m yr⁻¹ - which (i) minimizes misfit with observations (Fig. 4A, S6A), (ii) is consistent with satellite-based estimates (Extended Data Fig. 2), and (iii) aligns with ocean numerical models²⁸ - we project that simulations omitting a grounding zone, or prescribing a narrow, 1-km grounding zone, yield a sea-level contribution of 8.3 ± 3.0 mm by 2100, which is lower than the ISMIP6 mean ensemble and inconsistent with the calibrated hindcast constraints. Increasing the length of the grounding-zone to 3 ± 1 km, i.e., the maximum of the Bayesian posterior distribution (Fig. 4A), we estimate a lower-bound sea-level contribution resulting in a contribution of 19.4 ± 5.2 mm by 2100, i.e., an increase of 130% relative to the mean of the ISMIP6 ensemble. A grounding zone length of 5 ± 1 km, consistent with recent (2020s) high-resolution observations, with frequent documented intrusions that extend even beyond Mouginot Ridge (Fig. 1B), produces projections of 30.1 ± 5.7 mm by 2100, that is twice the mean of the ISMIP6 ensemble under constant climate forcing scenario.

Discussion

We present a calibrated model that reproduces independently acquired historical trends including velocity and thickness change as well as glacier migration, establishing a strong benchmark for projection skill. This performance improvement arises from explicitly representing actively melting grounding zones, rather than treating the grounding line as a discrete grounded–floating boundary. This framework also highlights sensitivity to bed topography uncertainties that hinder accurate reconstruction of historical grounding-line positions². The results underscore that, within the observationally supported range of melt rates, grounding-zone extent exerts a stronger control on retreat and unpinning than moderate variations in thermal forcing.

A remaining limitation of our simulations is the assumption of a fixed grounding-zone length, whereas the extent of the grounding zone likely varies with local bed and surface slopes, height above flotation, tidal amplitude and subglacial water pressure^{20,23,34,41}. In our framework, the grounding zone length should be interpreted as an effective, time-averaged intrusion scale that reproduces decadal evolution. Satellite observations indicate that kilometer-scale seawater intrusions have persisted over multiple years and expanded gradually over decadal timescales, supporting the use of a time-averaged grounding-zone length as a first-order representation of contemporary conditions. Consistent with this interpretation, our Bayesian calibration constrains an effective grounding-zone scale that integrates these processes over the historical period. Developing a prognostic parameterization that links grounding-zone extent to evolving geometry and subglacial hydrology remains an important direction for future work²³, but our results suggest that the mean kilometer-scale extent, rather than short-term variability, exerts the dominant control on multi-decadal retreat and ridge unpinning.

Our treatment of submarine melting relies on a simplified parameterization that prescribes spatially uniform submarine melt rates, rather than resolving ocean circulation and heat transport within evolving ice-shelf cavities. Although this approach does not capture the full complexity of the time-variable dynamics of the ocean, the melt rates adopted here are consistent with satellite-derived estimates²⁷ and high-resolution ocean modeling studies^{28,42}. By calibrating these effective melt rates against observed changes in glacier state, we constrain the integrated impact of ocean forcing over multi-decadal timescales. Fully coupled ice–ocean simulations would allow a detailed representation of cavity circulation and feedback¹⁰, but such models remain limited by sparse observational constraints within the Thwaites Ice Shelf and are computationally expensive. Comprehensive in situ measurements of temperature, salinity and circulation beneath the ice shelf⁴³ remain therefore important to reduce structural uncertainty and develop more physically complete melt parameterizations.

More importantly, our results show that kilometer-scale, actively melting grounding zones fundamentally alter both hindcast and forecast modeling skills for Thwaites glacier. By explicitly resolving kilometer-scale, actively-melting grounding zones and calibrating the model against multiple independent observational constraints, we reproduce the modern observational record and obtain sea-level contributions that are substantially larger than previously reported. Therefore, actively melting grounding zones should be systematically incorporated in regional- and continental-scale ice-sheet models, including within international intercomparison frameworks. At the same time, expanded in situ observations near grounding zones are urgently needed^{13,24} to better constrain heat delivery to sub–ice-shelf cavities and melt dynamics. Improving the representation of these processes is critical to narrowing the uncertainty in Antarctica’s contribution to future sea-level rise¹¹.

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212 Conceptualization: M.P. and E.R. Methodology: M.P., H.S. and J.C. Investigation and Formal Analysis:
213 M.P., E.R., H.S., J.C., T.P. Visualization: M.P. Software: M.P., H.S. Supervision: E.R., E.L. Funding
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224 **Competing Interests Statement**

225 The authors declare no competing interests.

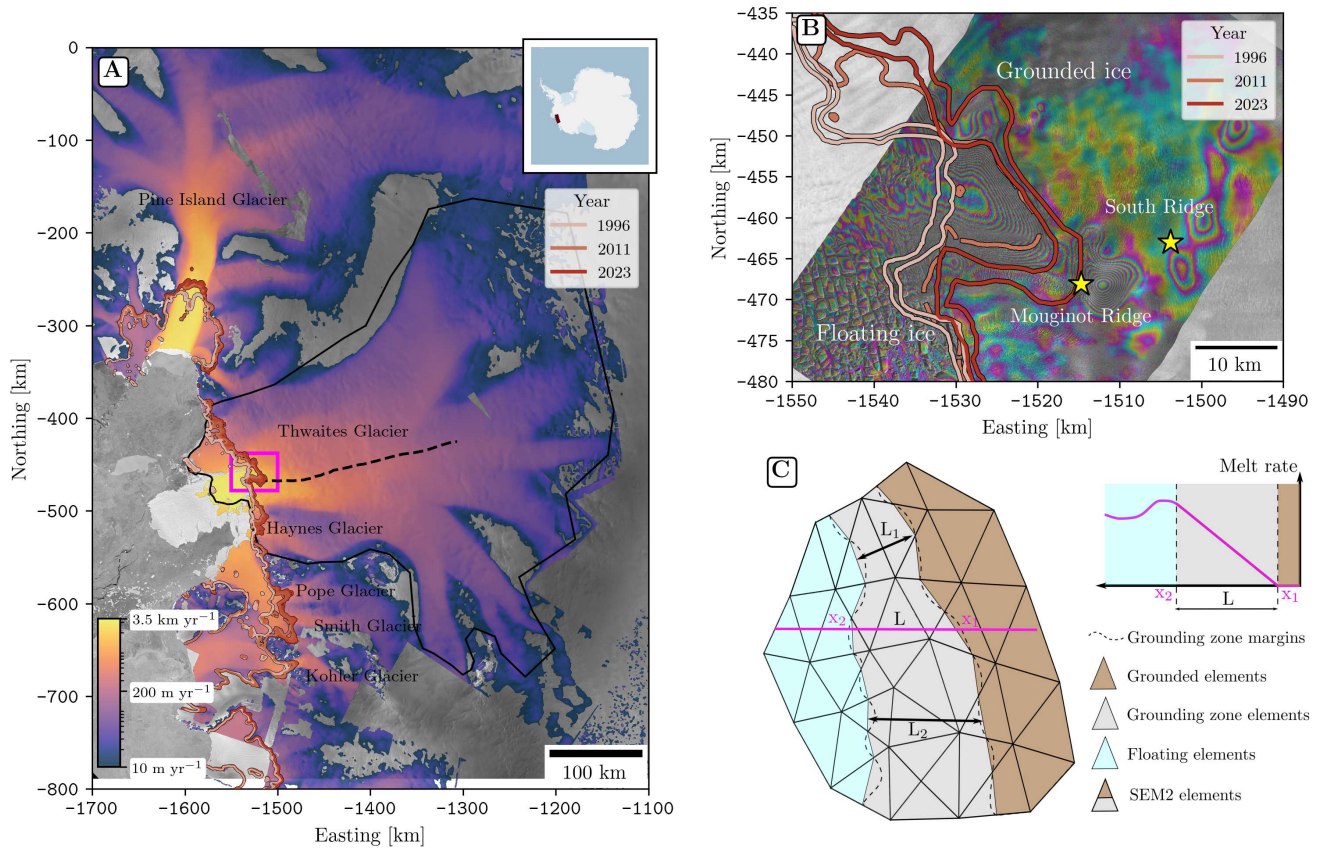


Fig. 1. Thwaites Glacier underwent rapid changes in the past few decades. (A) Ice sheet velocity map derived from ERS-1/2 and RADARSAT-1 Synthetic Aperture Radar (SAR) data between 1996 and 2001³⁰, overlaying the RADARSAT-derived SAR mosaic. Red lines are grounding line locations from 1996 to 2023³². Dashed line is a flow-line along the fastest flowing sector of Thwaites. Magenta box is the area of observed fastest retreat. (B) Differential, interferometric SAR fringes from ICEYE acquired in 2023 in the magenta box of panel (A). Red lines are the observed grounding zone extent from 1996 to 2023. (C) Grounding zone parameterization in ISSM, where L is grounding-zone length, blue and dark brown elements are floating and grounded ice elements. Light brown elements are actively melting grounding zone elements. SEM2 are elements with a sub-element-melt parameterization⁴⁴.

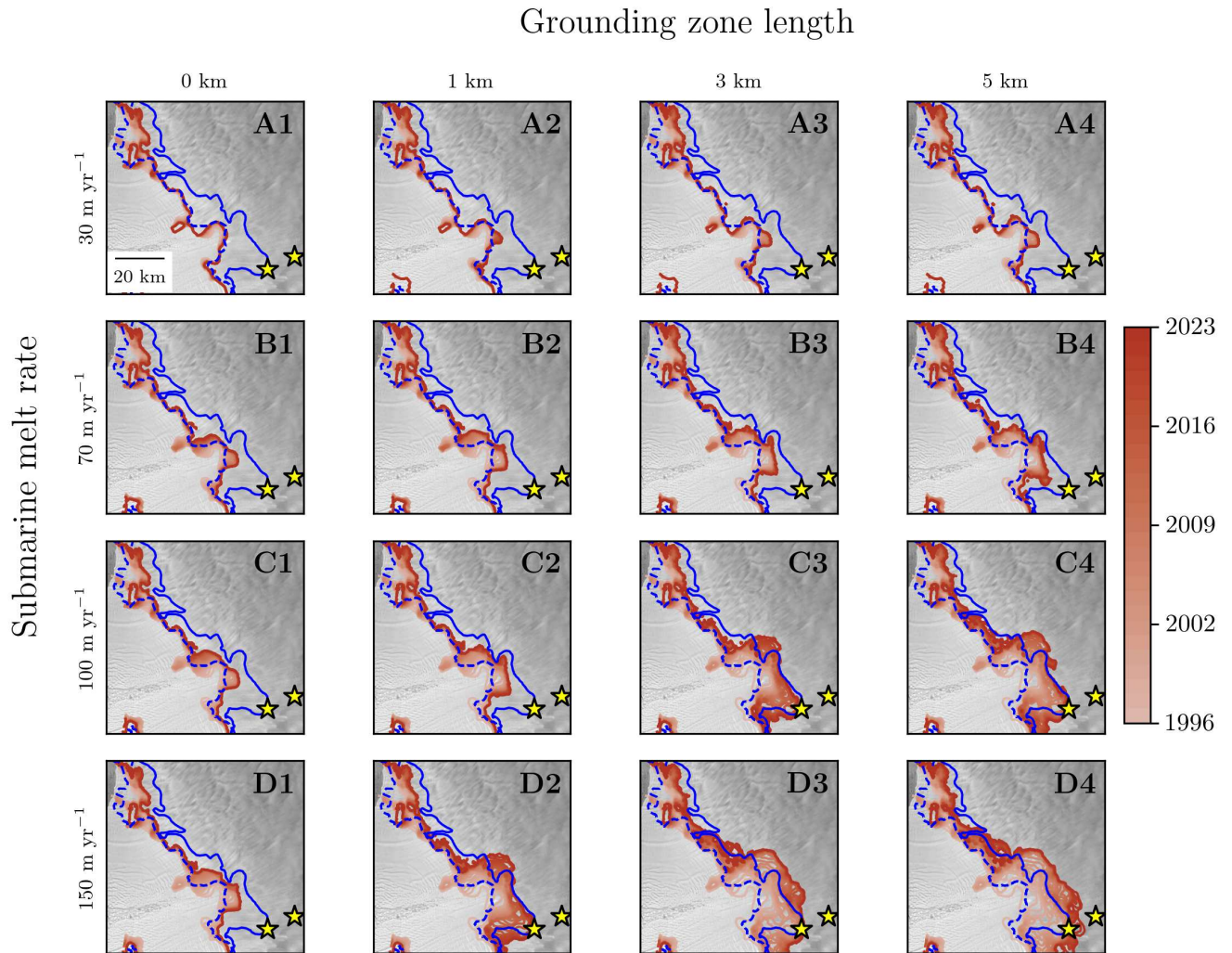


Fig. 2. Reconstruction of Thwaites' historical retreat. Color-coded lines are the modeled grounding line location from 1996 to 2023. The dashed and solid blue lines are observed locations in 1996 and 2023, respectively²⁰. Yellow stars are the location of Mouginot Ridge and South Ridge. Background is the RADARSAT-derived SAR mosaic.

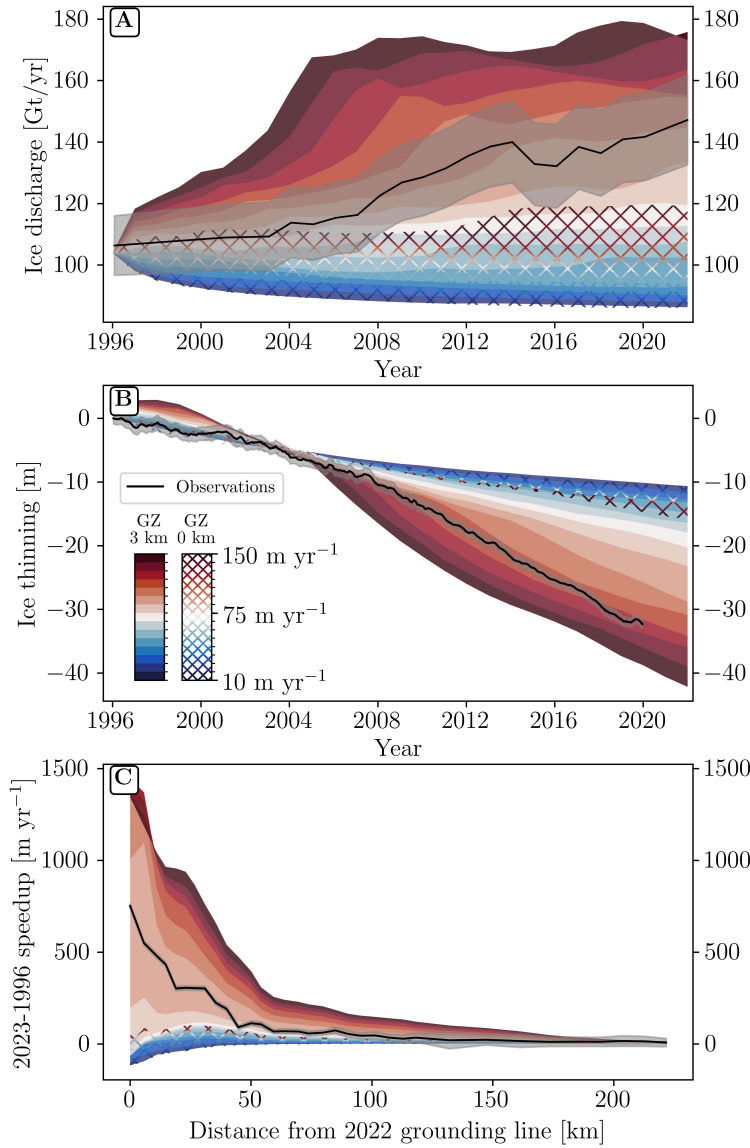


Fig. 3. Reconstruction of past ice discharge, glacier thinning and ice flow speed-up. Solid colored areas are model results with a melt rate range between 10 and 150 m yr⁻¹ (color-coded) and a grounding zone length of 3 km (solid area) and no grounding zone (hatched area). (A) Ice discharge at the grounding line. (B) Glacier thinning. (C) Glacier speed-up along the fastest moving sector of Thwaites (dashed black line in Fig. 1A). Black lines with grey shades are observed time series and associated error bar from¹ in (A),³¹ in (B), and³⁰ in (C).

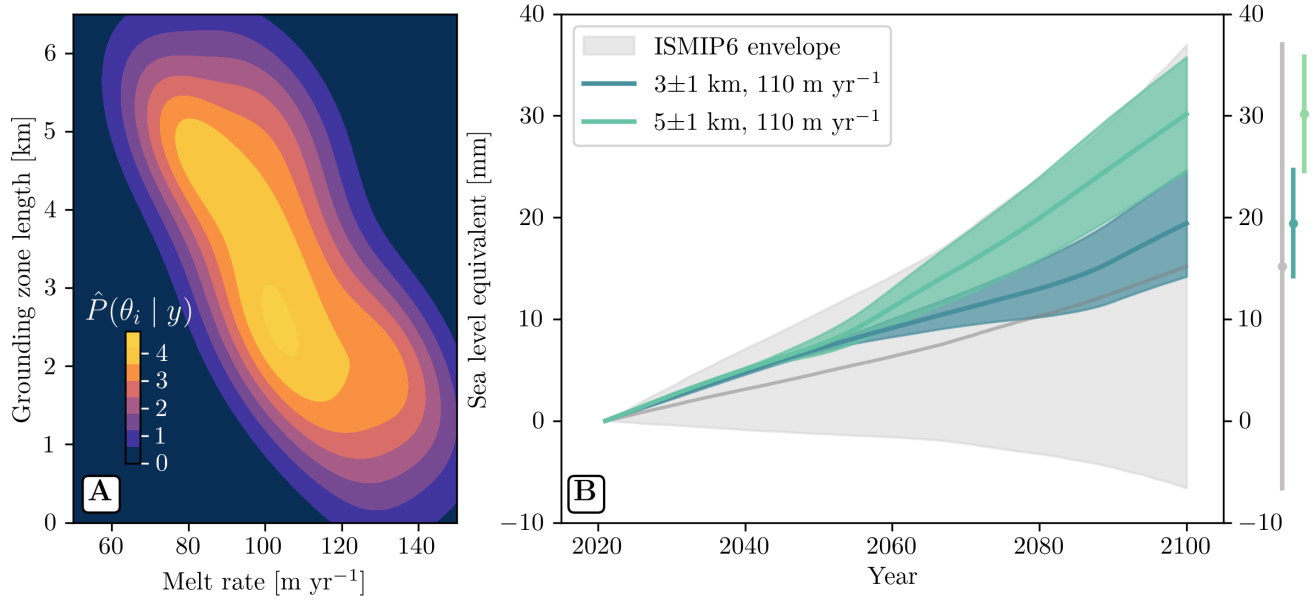


Fig. 4. Calibrated sea-level projections of Thwaites Glacier. (A) Normalized posterior distribution in a Bayesian framework, where $\hat{P}(\theta_i | y)$ is the probability of the parameter θ_i (melt rate and grounding zone length) conditioned on the observed historical trend y . (B) Ensemble of sea-level equivalent from model simulations and the ISMIP6 ensemble under constant forcing scenario³⁸. Colored lines are simulations for different grounding-zone length and melt rate of 110 m yr^{-1} .

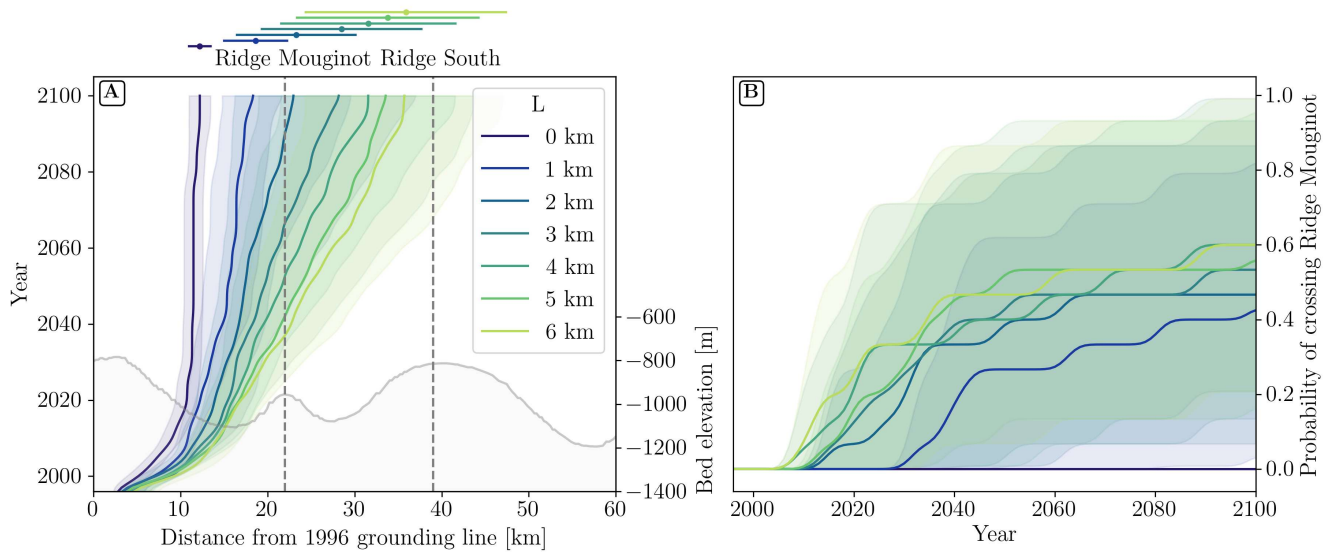


Fig. 5. Future retreat beyond Mouginot Ridge. (A) Modeled grounding-line retreat relative to the observed 1996 position, calculated as the average along the fastest moving sector of Thwaites (dashed black line in Fig. 1A). Each line (shaded area) represents the average (standard deviation) along all melt-rate simulations and grouped by individual grounding-zone lengths. (B) Cumulative probability of grounding line retreat beyond Mouginot Ridge before 2100.

Methods

Ice-sheet model configuration

We simulate the past, present, and future dynamics and thermodynamics of Thwaites Glacier, in the Amundsen Sea Embayment sector of West Antarctica, using the Ice-sheet and Sea-level System Model (ISSM)²⁹. The ice sheet model domain covers the grounded and floating portions of Thwaites and Haynes Glaciers, which are discretized using an unstructured triangular mesh (Supplementary Fig. 2A). The two-dimensional mesh contains 89,824 elements generated with a Bidimensional Anisotropic Mesh Generator. The element size varies from 500 m near the grounding zone to capture small-scale spatial features on fast flowing ice to 20 km on slow-moving regions.

We reconstruct and initialize the state of the glacier system by configuring ISSM with conditions representative of the 1990s. We assimilate observational data from that period, where and when available. The model domain is based on a 1996-2001 velocity map derived from ERS-1/2 and RADARSAT-1 SAR data⁴⁵, using a fixed ice front; this is an approximation supported by recent work showing that calving alone is unlikely to be the primary driver of sustained dynamic instability at Thwaites under current forcing^{7,8}. The elevation of the bedrock is from Bedmachine v3.7², using reconstructions based on circumpolar gravity anomalies under floating ice and mass conservation on grounded ice. To reproduce the position of the grounding line observed with double-difference interferometric SAR in 1996, the elevation of the bed is increased by up to 100 m where Bedmachine v3.7 calculates bed elevation via gravity inversion. The correction of 100 m is within the errors of gravity inversion techniques. We define an average grounding line location for 1996 from a set of 10 grounding lines retrieved from ERS-1 and ERS-2 between 1992 and 1996. For the observed position in 2021-2023, we use a long time series of short time ICEYE repeat pass data. The surface elevation is from the Antarctic Digital Elevation Model (DEM) produced from ERS-1 radar altimetry during the 1994–1995 geodetic phase and calibrated with ICESAT 2003-2004 data⁴⁶. The thickness of the ice and the transition from floating to grounded ice are derived from the historical DEM assuming hydrostatic equilibrium (Supplementary Fig. 2B).

We derive ice stiffness for floating ice and basal friction coefficient for grounded ice (Supplementary Fig. 2C) using inversion methods that minimize the discrepancy between the modeled and observed surface velocity (Supplementary Fig. 2D), here the SAR-derived 1996–2001 glacier velocity map⁴⁵. To simulate glacier flow, we use the 2-D Shelfy-Stream Approximation (SSA)²⁹. To model sliding at the glacier base, we select a regularized Coulomb friction law⁴⁷, which represents the relationship between the basal shear stress (τ_b) and the basal ice velocity (u_b) as

$$\tau_b = \frac{C_s |u_b|^{\frac{1}{m}-1} u_b}{(1 + \frac{C_s}{C_{max} N} |u_b|)^{\frac{1}{m}}} \quad (1)$$

where C_s is the friction coefficient, N is the effective pressure and C_{max} is the upper limit of the Iken's bound (τ_b/N)⁴⁸. The use of a regularized Coulomb friction law (with $m = 3$) over a Budd- or Weertman-style friction law was shown to be better able to reproduce the historical speed-up of the Pine Island Glacier¹², located 250 km north of Thwaites Glacier.

To retrieve the rheology of grounded ice, we extrude the domain into a three-dimensional model and use the Blatter-Pattyn Higher-Order stress balance approximation²⁹ with 10 vertical terrain-following levels to initialize the ice temperature based on an enthalpy thermal model⁴⁹. To initialize the thermal model, we solve the higher-order stress balance to obtain a three-dimensional velocity and pressure fields, then run the thermal model to steady state forced by the geothermal heat flux⁵⁰. The resulting temperature field is depth-averaged to inform the ice rheology on grounded ice.

Seawater intrusions at the grounding zone

We simulate the concept of seawater intrusions under grounded ice in ISSM building on²² and²⁵. Basal melt in the km-wide grounding zone decreases linearly from a maximum value at the grounding line $\dot{m}_{GL}$, calculated by hydrostatic equilibrium, to zero at a specified intrusion distance L :

$$\dot{m}_x = \dot{m}_{GL} \left(1 - \frac{x}{L}\right) \quad (2)$$

The basal melt, governed by the law above, is applied to each element upstream of the flotation line (Fig. 1C). The linear decrease in melt with distance is consistent with high-resolution ocean modeling²⁸.

In simulations without a grounding zone, the location of the grounding line within individual elements is tracked using the migration parameterization of the SEM2 sub-element grounding line, where the basal friction is integrated over the grounded portion of an element⁴⁴. We extend the parameterization of SEM2 at the end of the grounding zone, i.e., at the intrusion distance L , effectively simulating the end of the grounding zone as an element with partial melting. This novel implementation is freely available in ISSM v2026.1²⁹.

Hindcast ensemble

After configuring the model, we perform a 6-year ice sheet spin-up (Supplementary Fig. 3) using a constant surface mass balance prescribed by the regional atmospheric model Glacier Energy and Mass Balance (GEMB)³⁶, based on average conditions for the period 1996–2021. The forcing of climate at the ice surface and the absence of a basal melt rate under the ice shelf allow the dynamic model to relax to the imposed initial conditions. The short spin-up is necessary to ensure that i) the initial modeled grounding line matches its observed location in 1996 (Supplementary Fig. 3A) and ii) the modeled loss at the grounding line agrees with the observed mass loss in the mid-1990s (Supplementary Fig. 3B). Note that the bed correction of 100 m under floating ice is also necessary for the model to adjust to the 1996 conditions (Extended Data Fig. 1 and discussion in the manuscript). Without this correction, the spin-up brings the model to a state which is not representative of the historical state. Following the short spin-up, we run a suite of numerical simulations under constant-forcing baseline with an average surface mass balance for 1996–2021³⁶. Finally, we use a fixed time-step of two weeks to ensure numerical convergence.

The goal of the hindcast exercise is to reproduce the three satellite constraints described above (surface velocity³⁰, thickness change³¹, and grounding-line migration³²). It is important to note that the last metric to assess the full state of the glacier is the location of the ice front, as calving could lead to buttress removal and glacier acceleration, although recent studies showed that calving itself is unlikely to be the dominant driver of sustained dynamic instability in Thwaites under current conditions^{7,8}. In the hindcast simulations we investigate the sensitivity of the historical changes to the following parameters: $\theta_i = (\dot{m}, L)$, where $\dot{m}$ is the melting rate under floating ice and L is the length of the grounding zone, following the numerical framework of²⁵. We perform a total of 105 simulations combining $\dot{m}$ and L , under the observational baseline described below.

The ocean-driven melting of floating ice is represented using a depth-dependent linear basal melt parameterization¹⁰, with a maximum melt rate at and below 800 m, linearly decreasing to zero at 200 m. No basal melt is applied to floating ice shallower than 200 m. The maximum melt rate at 800 m ranges from 10 m yr⁻¹ to 150 m yr⁻¹ in increments of 10 m yr⁻¹. This parameterization reproduces both the amplitude of enhanced melting at depth near the grounding line inferred from satellite observations²⁷ (Extended Data Fig. 2) and numerical models⁴², although it is less sophisticated than fully coupled ocean circulation models¹⁰. Melt rates in the eastern sector of Thwaites Glacier have a maximum melt of 10 m yr⁻¹, following robotic observations near the grounding zone⁴³. This assumption is supported by

observations showing that CDW in this region is found at roughly 580 m depth, with robotic observations showing suppressed melting in this area⁴³, consistent with satellite data (Extended Data Fig. 2)²⁷, and ocean modeling results⁴². Because the grounding-line positions from the 1990s to today lie at shallower depths in some areas, CDW - the main origin of melting in the area - would struggle to spill above the current grounding-line location and into the glacier basin. Any such intrusion would most likely occur through the deep bathymetric channel beneath the TMT, but it would take centuries to reach the eastern sector³⁵.

Regarding grounding zone length, we test the sensitivity of glacier dynamics using scenarios without a grounding zone and with lengths ranging from 1000 m to 6000 m, varying in 1000 m increments. These values are consistent with grounding zone lengths inferred from satellite observations in the Amundsen Sea Embayment²⁰ and in other sectors of Antarctica^{34,41}.

Bayesian calibration and sea-level projections

To identify the combinations of melt rate and grounding zone length that are most likely to explain historical trends, we employ a Bayesian calibration approach following the framework of^{15,16}. The Bayes theorem is written as:

$$\hat{P}(\theta_i | y) = \left\| \frac{\mathcal{L}(y | \theta_i) P(\theta_i)}{P(y)} \right\| \quad (3)$$

where $P(\theta_i | y)$ is the posterior distribution, representing the probability of the parameter θ_i conditioned on the observed data y ; $\mathcal{L}(y | \theta_i)$ is the likelihood, quantifying to what extent the set of model parameters θ_i reproduces the observed data y ; $P(\theta_i)$ is the prior distribution, reflecting the initial uncertainty about the parameter before observing the data. Finally, $P(y)$ is the marginal likelihood, serving as a normalization constant.

We begin by generating a uniform (randomly selected) prior ensemble consisting of 10,000 samples of melt rate and grounding-zone length, which are the parameters θ_i . These are drawn uniformly from the hindcast sensitivity grid, with a melt rate $\dot{m} \sim \mathcal{U}(10, 150)$ and a grounding-zone length $L \sim \mathcal{U}(0, 6000)$. To ensure statistical robustness, the model outputs are linearly interpolated from the uniform space θ_i to construct a set of surrogate models f , which reproduce the key observables:

$$f : (\dot{m}, L) \mapsto [R_{\text{west}}, R_{\text{east}}, D, T, S], \quad (4)$$

where R denotes the retreat of the grounding line (for east and west), D denotes the ice discharge at the grounding line, T is the thinning of the glacier, and S is the increase in glacier speed.

For each parameter sample $\theta_i = (\dot{m}, L)$, we evaluate the modeled output (R, D, T, S) using the surrogate model f and compare them with observational values using a Gaussian likelihood function:

$$\mathcal{L}(y | \theta_i) = \frac{1}{\sigma \sqrt{2\pi}} \exp \left[-\frac{1}{2} \left(\frac{f(\theta_i) - y}{\sigma} \right)^2 \right], \quad (5)$$

where y is the observed value of grounding line retreat R , ice discharge D , glacier thinning T , and speed-up S , while σ is the associated uncertainty. Assuming uniform priors, the posterior weights are proportional to the normalized likelihoods:

$$w_i = \frac{\mathcal{L}(\theta_i)}{\sum_{j=1}^N \mathcal{L}(\theta_j)}. \quad (6)$$

342 We perform resampling from the prior ensemble using these weights to obtain posterior distributions
343 $P(\theta_i)$, where θ_i helps to calculate the melt rate $\dot{m}$ and the length of the grounding zone L . These posterior
344 distributions are used to identify the most likely combination of parameters.

345 For each “weighted” simulation, we extend the model until the year 2100 to estimate the contribution
346 to sea-level rise, calculated from the volume of ice lost above flotation in the form of a sea-level equivalent.
347 Using the posterior distribution $P(\theta_i | y)$ obtained from the Bayesian calibration to weight the most likely
348 simulations, we identify the high-density parameter combinations $\theta_i = (\dot{m}, L)$ that are most consistent
349 with the data and extract the corresponding sea-level trajectories.

350 Satellite observations

351 We use a suite of satellite-derived datasets to evaluate the likelihood of model results as a historical trend y
352 in the Bayesian framework described above. Ice surface velocities are obtained from Synthetic Aperture
353 Radar (SAR) interferometry⁴⁵, providing high-resolution ice motion across grounded and floating ice as
354 an average between 1996 and 2001 using ERS-1/2 and RADARSAT-1 SAR data, and between 2020-2022
355 using Sentinel-1a/b, RADARSAT-2, and Landsat-8 data. Glacier speed-up is calculated in reference to the
356 average velocity in 1996-2001.

357 The position of the grounding line between 1996 and 2023 (Supplementary Fig. 1) is retrieved with
358 quadruple differential Interferometric SAR (InSAR). For the 1990s, we use ERS-1 and ERS-2 tandem
359 data. For 2023-2025, we use the ICEYE constellation²⁰ on the main trunk and RADARSAT Constellation
360 Mission (RCM) data on the eastern ice shelf. To quantify changes in ice thickness, we use incorporated
361 altimetry data from ICESat, ICESat-2, and CryoSat³¹. The ice discharge at the grounding line is calculated
362 using the vector velocity and ice thickness as a mass flux at the grounding line, following the approach of
363 ref.¹.

364 Data availability

365 Ice sheet velocity maps are available at ref.³⁰, glacier elevation change is available at ref.³¹, grounding-line
366 locations and grounding zone extensions are available at ref.³².

367 Code availability

368 The Ice-Sheet and Sea-Level System model (ISSM) is freely available on Github (<https://github.com/ISSMteam/ISSM.git>), together with the novel grounding zone parameterization. Code to
369 reproduce the model simulations and model output are available on Zenodo (<https://doi.org/10.5281/zenodo.16551126>).
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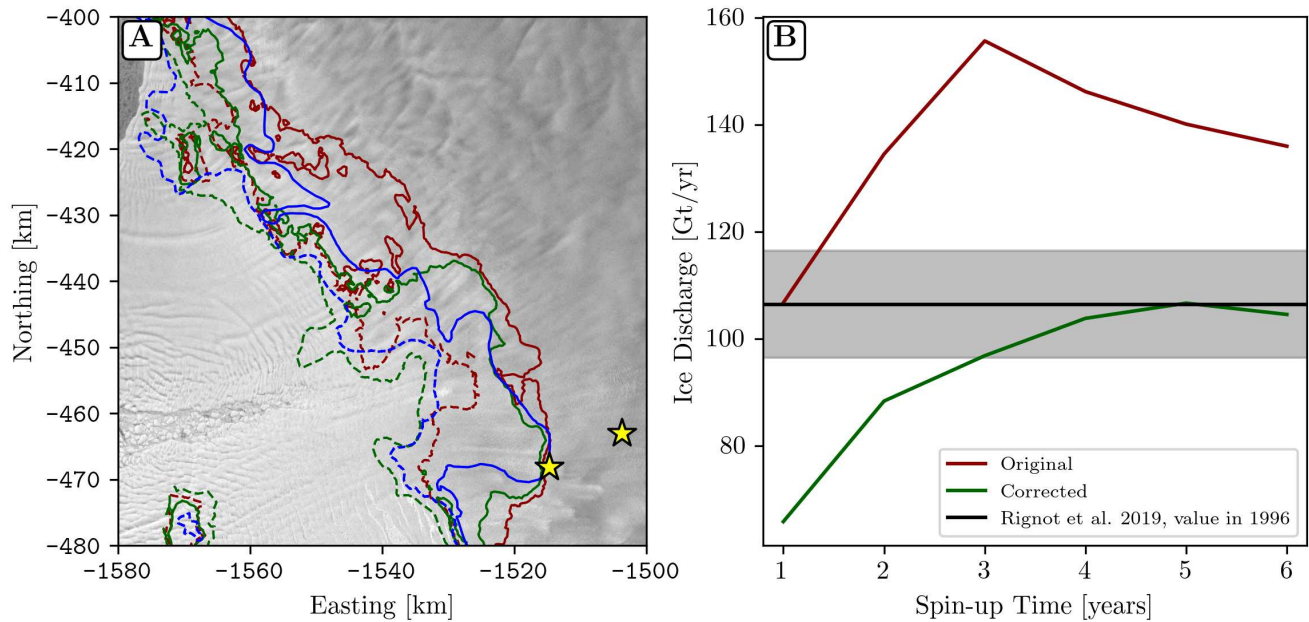
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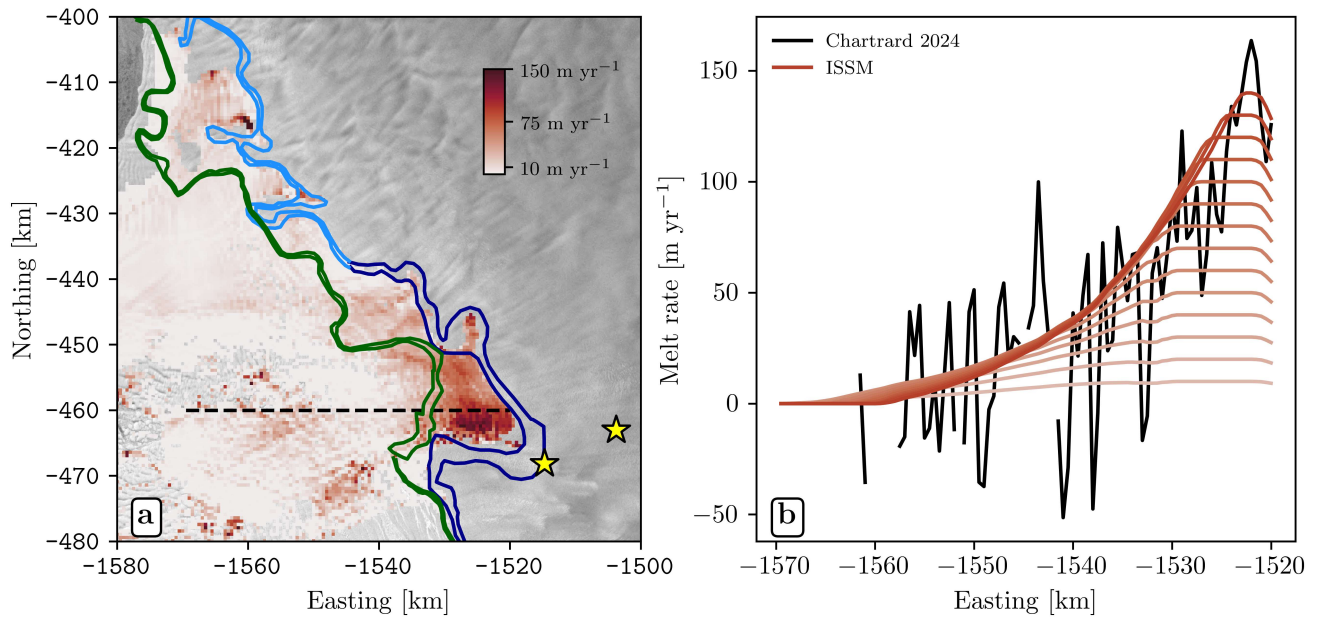
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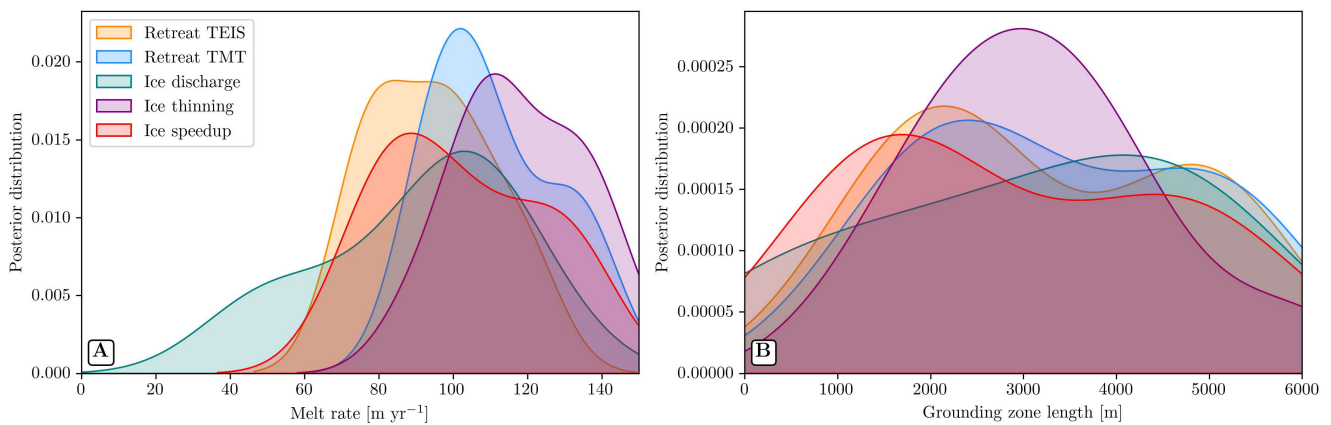
Extended Data



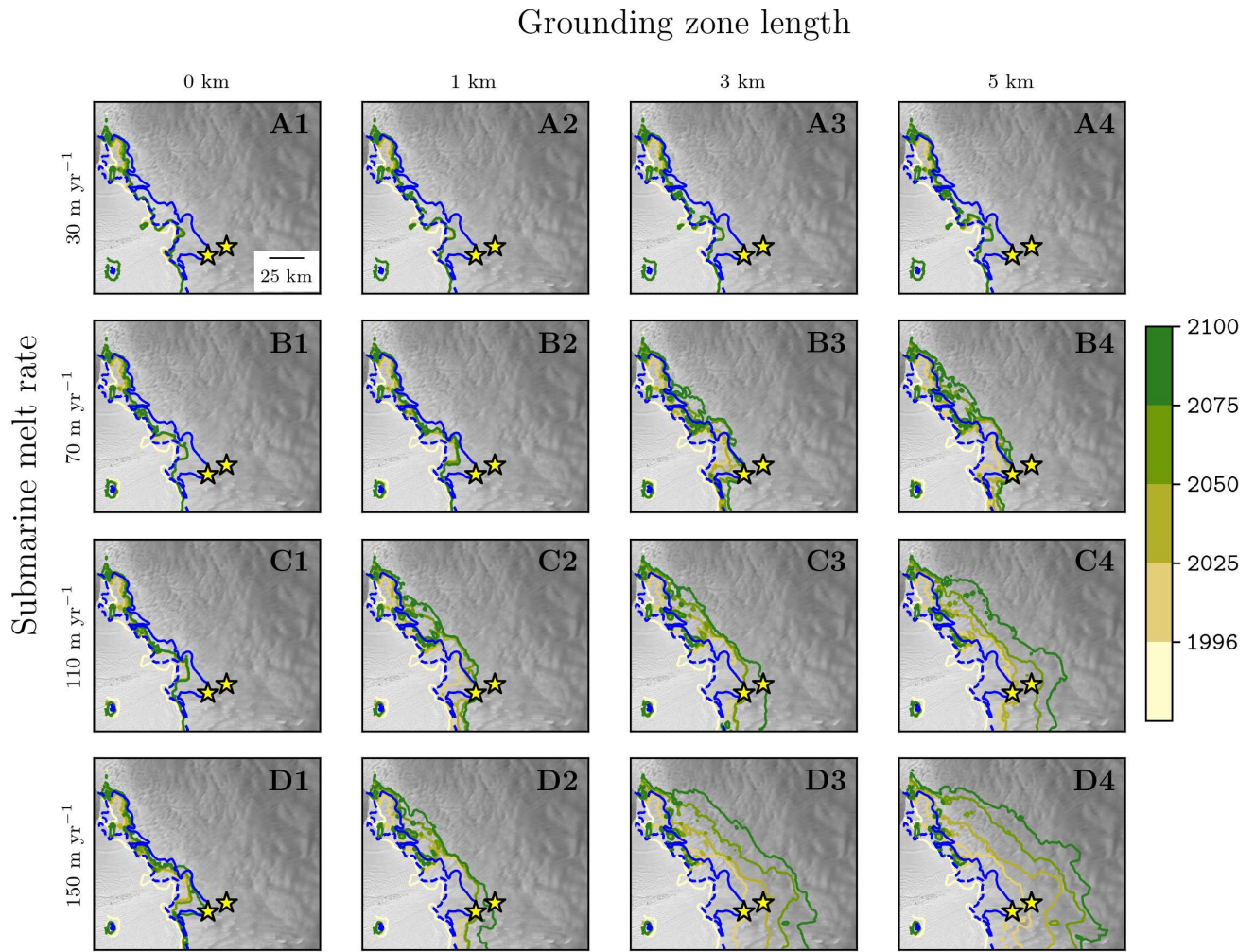
Extended Data Fig. 1: Implications of the 100 m bed correction. A) Modeled grounding line location for maximum a-posteriori simulations (melt rate of 100 m yr^{-1} and grounding zone of 3 km) in 2021 for model run with original bed topography² (solid red line), and with 100 m lifting under floating ice (solid green line). Dashed green and red lines are the grounding line location at the beginning of the hindcast exercise, representative of 1996. Dashed blue line is the observed grounding line location in 1996³², solid blue line is the observed grounding line location in 2023²⁰. Yellow stars are the location of Ridge Mouginot and Ridge South. Background is the RADARSAT-1 mosaic of Antarctica. (B) Ice discharge for simulation with original bed (red line) and with 100 m lifting under floating ice (green line). Black line is observed ice discharge in 1996 and shaded area is the standard deviation¹.



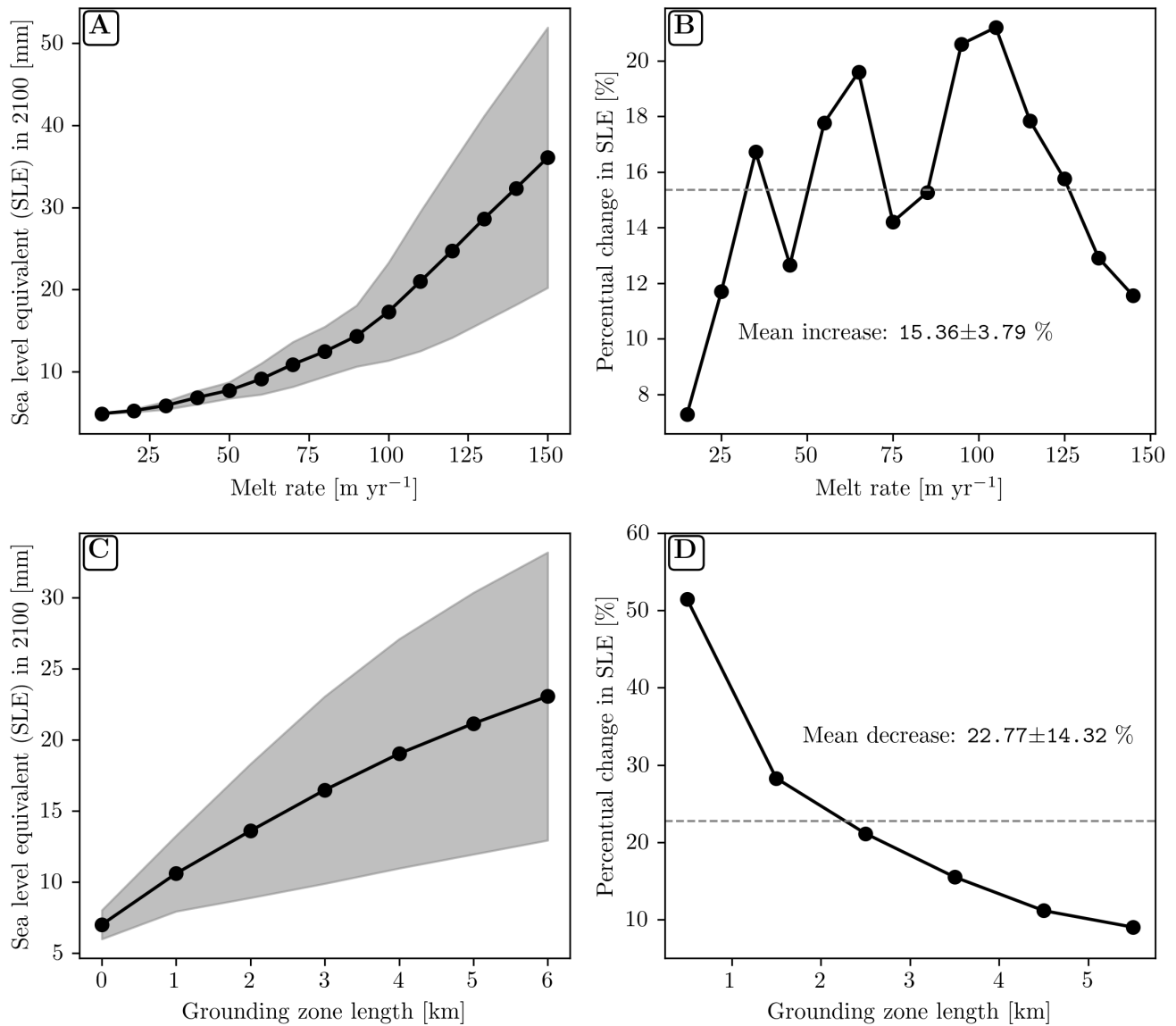
Extended Data Fig. 2: Model-data comparison. (A) Basal melt rate as average between 2010-2021 from satellite observations²⁷. Dark green line is the extent of the grounding zone in 1996 from satellite observations³². Dark and light blue lines are the extent of the grounding zone on Thwaites Main Trunk and Eastern Ice Shelf, respectively²⁰. (B) Observed melt rate (black line) and depth-dependent parameterization in ISSM (red lines) along dashed line in panel (A).



Extended Data Fig. 3 Posterior density function $P(\theta_i | y)$, representing the probability of simulations employing the parameter θ_i to reproduce the observed data y . (A) θ_i is melt rate, (B) θ_i is grounding zone length.



Extended Data Fig. 4: Future grounding line retreat. Color-coded lines are the modeled grounding line location from 1996 to 2100. The dashed and solid blue lines are the grounding line locations observed in 1996 and 2023, respectively³². Panel (C3) is the maximum-a posteriori (MAP) simulation. Yellow stars are the location of Ridge Mouginot and Ridge South. Background is the RADARSAT-1 mosaic of Antarctica.



Extended Data Fig. 5: Sea level equivalent in 2100 as function of melt rate and grounding zone length. (A) Sea level equivalent (SLE) in 2100 for simulations with different melt rates. Black line is the average and shaded area is the standard deviation across simulations with different grounding zone lengths. (B) Percentual change in SLE at each 10 m yr⁻¹ of basal melt rate increase. (C) Sea level equivalent (SLE) in 2100 for simulations with different grounding zone lengths. Black line is the average and shaded area is the standard deviation across simulations with different melt rates. (D) Percentual change in SLE at each 1 km of grounding zone length.

Extended Data Table 1: Tabulated sea-level contribution. Calculated sea level equivalent [mm] of Thwaites glacier in 2100 from model simulations. Mean of the ISMIP6 ensemble³⁸ is 15.2 ± 21.8 mm.

Melt Rate	Grounding zone length						
	0 m	1000 m	2000 m	3000 m	4000 m	5000 m	6000 m
10 m yr ⁻¹	4.8	4.9	4.9	4.9	5.0	5.0	5.0
20 m yr ⁻¹	4.9	5.1	5.3	5.3	5.4	5.4	5.5
30 m yr ⁻¹	5.2	5.5	5.6	5.8	6.2	6.4	6.6
40 m yr ⁻¹	5.4	6.0	6.7	7.1	7.4	7.6	7.8
50 m yr ⁻¹	5.7	7.1	7.7	8.0	8.2	8.5	8.9
60 m yr ⁻¹	5.8	7.9	8.3	8.9	10.1	11.0	11.8
70 m yr ⁻¹	6.0	8.4	9.9	11.7	12.8	13.5	14.0
80 m yr ⁻¹	6.6	9.4	12.4	13.8	14.6	15.0	15.3
90 m yr ⁻¹	7.3	11.6	14.1	15.1	15.5	16.8	20.0
100 m yr ⁻¹	7.8	13.2	15.2	15.8	19.0	23.0	27.2
110 m yr ⁻¹	8.3	14.3	15.8	19.4	25.2	30.2	33.8
120 m yr ⁻¹	8.8	15.1	17.5	25.1	31.4	35.9	39.4
130 m yr ⁻¹	9.1	15.7	22.1	30.7	36.5	41.2	45.2
140 m yr ⁻¹	9.4	16.6	27.0	35.4	41.5	46.5	49.9
150 m yr ⁻¹	9.9	18.1	31.4	39.9	46.5	51.2	55.4

Supplementary Files

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