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Submarine melting of glaciers in Greenland amplified by atmospheric warming

In the format provided by the authors and unedited

Supplementary Material for '*Submarine melting of glaciers in Greenland amplified by atmospheric warming*' by Slater & Straneo

S1 Datasets

S1.1 Selecting tidewater glaciers

To define the set of tidewater glaciers considered in this study we start from a list of the 243 fastest flowing glaciers in Greenland provided in Table S1 of [1]. Since the list contains land-terminating glaciers and since we wish to exclude glaciers that are barely marine-terminating, we remove glaciers that are either land-terminating or have a maximum grounding line depth <50 m, removing 69 glaciers. These shallow grounding line depths may be accurate or may result from a lack of observations with which to constrain the ice thickness [1].

We also require ocean properties for each glacier, which are extrapolated to calving fronts from the continental shelf taking account of the fjord bathymetry (section S1.3). We therefore remove glaciers for which the deepest path down the fjord to the ocean involves a point shallower than 25 m, removing a further 41 glaciers. If this fjord bathymetry is accurate then these glaciers are relatively isolated from the ocean and may be more appropriately considered as lake-terminating rather than marine-terminating glaciers, and as such they are excluded from this study. On the other hand, we are aware of at least one case where the fjord bathymetry in BedMachine v3 is inaccurate and is shallower than reality - Sarqardleq Sermia and Sarqardleq Fjord in west Greenland [2]. Until these and other better observations of fjord bathymetry are incorporated into BedMachine, such glaciers are excluded.

Lastly, to focus on the larger glaciers that are most relevant for sea level contribution we remove those with a mean annual subglacial discharge $<2.5\ \mathrm{m^3\,s^{-1}}$ (section S1.2), removing 10 glaciers.

The final set of 123 tidewater glaciers considered in this study (e.g. Fig. 2 of the main article) are, therefore, those that according to present datasets are (i) grounded in more than 50 m of water, (ii) have an open connection to the ocean deeper than 25 m and (iii) have significant subglacial discharge. From the original list of the fastest flowing glaciers in Greenland [1], our final dataset contains 8 of the top 10, 42 of the top 50 and 76 of the top 100. The dataset contains a number of glaciers that either formerly or currently have an ice tongue or shelf (e.g. Jakobshavn Isbrae, Petermann, 79N).

S1.2 Subglacial discharge

We estimate monthly subglacial discharge by combining estimated surface runoff and basal melt rates with hydrological flow routing as in [3].

S1.2.1 Hydrological basins

The hydrological catchment is defined for each tidewater glacier by performing flow routing on the subglacial hydropotential ϕ given by

$$\phi = \rho_w g b + f \rho_i g h \tag{1}$$

in which $\rho_w = 1000\ \mathrm{kg\,m^{-3}}$ is the density of meltwater, $\rho_i = 917\ \mathrm{kg\,m^{-3}}$ is the density of ice and $g = 9.81\ \mathrm{m\,s^{-2}}$ is gravitational acceleration. Bed topography b (m) and ice thickness h (m) are from BedMachine v3 [1]. f is the ratio of subglacial water pressure to ice overburden pressure and we here set $f = 1$. We perform flow routing on ϕ using TopoToolbox [4] and define the hydrological catchment for each tidewater glacier as the area of the ice sheet over which subglacial water would ultimately drain through the calving front of the glacier. As an example, the resulting subglacial hydrological basin for Jakobshavn Isbrae is shown in Extended Data Fig. 1a.

S1.2.2 Subglacial discharge from surface runoff

Ice sheet surface meltwater runoff rates are obtained from the regional climate model RACMO2.3p2, run at 5.5 km resolution and then statistically downscaled to 1 km [5]. The simulation covers the time period 1958 to present but we focus here on the 40-year period 1979-2018. We assume that all of this surface runoff reaches the ice sheet bed, and therefore to obtain the monthly subglacial discharge due to surface runoff, we sum the monthly surface runoff from RACMO over the hydrological catchment for the glacier in question. An example for Jakobshavn Isbrae is shown in Extended Data Fig. 1b.

S1.2.3 Subglacial discharge from basal melting

A number of studies have postulated that significant subglacial discharge occurs even during winter, stemming from basal frictional melting where ice is sliding rapidly over its bed, geothermal heat flux and heat carried by surface waters that enter the subglacial hydrological system [e.g. 6, 7, 8]. The presence of even a small subglacial discharge flux would be sufficient to drive a plume and significant submarine melting [9] and so we require an estimate of subglacial discharge arising from basal melting.

Here, we estimate ice sheet basal melting following the method and dataset of [10]. The subglacial discharge resulting from ice sheet basal melt for each glacier is then estimated by summing the basal melting over the hydrological drainage basin for each glacier, as described for the surface runoff above. For Jakobshavn Isbrae the resulting estimated discharge is substantial at $\sim$50 $\mathrm{m^3\,s^{-1}}$ (Extended Data Fig. 1b), smaller than the 64-105 $\mathrm{m^3\,s^{-1}}$ estimated by [6], but is certainly sufficient to drive a significant plume.

In this study we wish to focus on variability in submarine melting due to atmospheric and oceanic variability, and we therefore consider the flux of basal meltwater to be constant throughout the time period. In reality it may vary due to variability in ice velocity.

The total monthly subglacial discharge is then the sum of the discharge from surface runoff and the (constant) discharge from basal melting (e.g. Extended Data Fig. 1b). The discharge is highly seasonal and is dominated by surface runoff in summer and basal melting in winter.

S1.2.4 Caveats

This method of estimating subglacial discharge makes a number of simplifying assumptions that we believe are justified over the space and timescales considered in this study. By considering only the subglacial hydropotential we are assuming that little supraglacial flow of meltwater runoff takes place. In particular, we are neglecting the possiblity for supraglacial hydrology to route meltwater runoff from one glacier catchment to another. While extensive supraglacial hydrological networks exist around the margins of the ice sheet during summer [e.g 11], the generally high moulin density (e.g. 1 per 25 km^2 below 1800 m.a.s.l in west Greenland according to [12]) limits the distance that can be travelled by surface meltwater runoff before it drains into the subglacial hyrological system. Once into the subglacial hydrological system we have assumed that the subglacial water pressure is equal to the ice overburden pressure ($f = 1$). This choice is supported by limited direct observations of subglacial water pressure [13, 14, 15], but in reality we would expect significant temporal and spatial variability in subglacial water pressure that could redirect subglacial water into an adjacent catchment [16]. Lastly, we have assumed that surface meltwater runoff generated in a given month exits the ice sheet in the same month; that is, there is little delay between surface meltwater runoff production and subglacial discharge of this runoff. Surface meltwater may be stored in surface lakes, but this is thought to be only a small fraction of the total and most of these lakes will drain by the end of the summer [17]. Refreezing of surface meltwater in the snowpack and firn is accounted for within RACMO2.3p2 [5]. Once into the subglacial drainage system, the transit time of meltwater depends on the state of the hydrological system [e.g. 18], but observations from land-terminating glaciers in west Greenland show that by mid-summer transit velocities exceeding $1\ \mathrm{m\,s^{-1}}$ can be expected, indicating little delay of meltwater in the subglacial hydrological system. In sum, process-understanding suggests that the average rate of ice sheet surface runoff during JJA is likely a good estimate of the average rate of subglacial discharge during JJA.

Finally, due to the obvious difficulties with making the relevant observations, we have no direct measurements of subglacial discharge at tidewater glaciers. Estimates can, however, be obtained by analysing the freshwater content of fjord waters close to tidewater glacier calving fronts. Where this has been achieved [19, 20], good agreement is found between these ocean observation-based estimates of subglacial discharge and the regional climate model-based method we have employed in our study.

S1.3 Ocean thermal forcing

In the absence of continuous, long-term oceanographic observations from a significant number of glacial fjords around Greenland, we have used ocean reanalysis and offshore observational products to estimate the temperature of ocean waters influencing Greenland's tidewater glaciers during 1979-2018. These products are ASTE_R1, ORAS5, CHORE and EN.4.2.1 (Table S1).

The Arctic subpolar gyre state estimate (ASTE_R1) is an Arctic-focused state estimate forced by the JRA-55 atmospheric reanalysis and constrained by observed ocean temperature, sea ice concentration and sea level anomaly [21]. Both ORAS5 and CHORE_RL are global ocean reanalyses.

Product	Period available	Resolution	Type	Reference
ASTE_R1	2002-2017	1/3°, 50 z-levels	State estimate	[21]
ORAS5	1979-present	1/4°, 75 z-levels	Reanalysis	[22]
CHORE_RL	1900-2010	1/2°, 75 z-levels	Reanalysis	[23]
EN.4.2.1	1900-present	1°, 42 z-levels	Objective analysis	[24]

Table S1: Summary of the oceanographic products used in this study.

ORAS5 is forced by the ERA-Interim atmospheric reanalysis and assimilates observations of sea surface and subsurface temperature, sea ice concentration and sea level anomaly [22]. CHORE_RL is forced by the ERA-20C atmospheric reanalysis, is relaxed towards observed sea surface temperature and assimilates subsurface profiles [23]. EN.4.2.1 is a regularly updated global compilation of oceanographic profiles. We used version 4.2.1 of the objective analysis, which is a gridded product based on interpolation of the raw profiles [24].

None of the oceanographic products explicitly include glacial fjords around Greenland and we, therefore, estimate ocean properties at tidewater glacier calving fronts based on continental shelf properties and using a similar method to [25] and [26]. For each tidewater glacier we first note the position and the depth of the deepest point of the calving front (call this the grounding line depth), and the position (grid points) of shelf points in the ocean product. We use BedMachine v3 bathymetry [1] to determine if there is a continuous path at the grounding line depth between the calving front and shelf points in the ocean product. If this is the case, and if at least one of these shelf points also reaches to at least the grounding line depth, then we assign the tidewater glacier the ocean properties from this shelf point at the grounding line depth. If multiple shelf points satisfy these conditions then we take the closest. Note that even if a shelf point is connected to the calving front by a continuous path at the grounding line depth according to BedMachine v3, the available profile at the shelf point may not reach to the grounding line depth because the bathymetry in the ocean product can differ from BedMachine v3. If there is no such continuous path at the grounding line depth, for example due to the presence of a sill, then we repeat this algorithm at successively shallower depths until it succeeds in associating shelf properties to the glacier. An example of this process for Helheim Glacier is shown in Extended Data Fig. 2.

By this algorithm, we identify the deepest, closest water on the continental shelf that is able to access the calving front without being impeded by bathymetry. We are essentially extrapolating properties into fjords from the shelf accounting for the presence of sills; the volume below sill depth in a fjord is assumed to fill up with water from the sill depth outside of the fjord. Note that we assign only a single grounding line temperature and salinity to the glacier rather than a full profile - we are assuming that due to upwelling, this grounding line value is the most relevant for submarine melting of the glacier. Since each of the 4 ocean products considered (Table S1) have different resolutions and bathymetry, the algorithm is repeated separately for each product (Extended Data Fig. 2b).

Grounding line water properties are converted to thermal forcing (TF), defined as

$$\mathrm{TF} = T - T_f = T - (\lambda_1 S + \lambda_2 + \lambda_3 z_{gl}) \tag{2}$$

where T is the temperature from the ocean model and T_f is the in-situ freezing point, the latter estimated as a linear function of the ocean model salinity S and the grounding line depth z_{gl} following [9]: $\lambda_1 = -5.73 \times 10^{-2}$ °C, $\lambda_2 = 8.32 \times 10^{-2}$ °C and $\lambda_3 = 7.54 \times 10^{-4}$ °C m^{-1}.

Following this process we have a monthly thermal forcing for each tidewater glacier from each of the 4 ocean products considered (Extended Data Fig. 2c). We combine these 4 time series into a single time series by averaging over the data products to essentially create a multi-model mean. The four ocean products are not entirely independent because all use the raw EN.4.2.1 profiles in some way, but they can differ due to many other factors, and so we consider the multi-model mean the most robust metric of changes in the ocean abutting Greenland's tidewater glaciers.

S2 Submarine melt rate

S2.1 Submarine melt rate parameterisation

We wish to quantify the impact of variability in submarine melting on glacier dynamics while also separating out the variability in submarine melting due to change in subglacial discharge and ocean temperature. Extensive previous modeling and theoretical work has shown that the submarine melting induced by a plume can be parameterised in the form $\dot{m} = \kappa\, Q^{\alpha}\, \mathrm{TF}^{\beta}$, where κ, α and β are constants, Q is subglacial discharge and TF is ocean thermal forcing [e.g. 9, 27, 28]. Note that within a discharge plume, submarine melt rate close to the grounding line is independent of terminus shape [29] and so provided we consider melt rate close to the grounding line (as in this study), we can equally apply this parameterisation at vertical or undercut fronts and at glaciers having an ice tongue or shelf.

In this study we take $\dot{m}$ to be the maximum submarine melt rate in the plume, which is found close to the grounding line [e.g. 27, 29]. To fix the constants κ, α and β we use the standard buoyant plume model that has been described in (among many other papers) [9] and [28]. Following observational work suggesting that a line plume geometry is most appropriate for plumes in Greenland [30, 31], we consider a line plume model of width w. The subglacial discharge per unit width is then $q = Q/w$. We run the line plume model for 10 values of q between 0.1 and 2 m^2/s, TF between 0.1 and 7 °C and grounding line depth z_{gl} between -850 m and -100 m. In these simulations the ocean is taken to be unstratified; the introduction of stratification, as is seen in Greenlandic fjords, does not significantly change the maximum melt rate because the maximum melt rate is found close to the grounding line where the fjords are typically very weakly stratified. The maximum melt rate achieved in these simulations is plotted in Extended Data Fig. 3.

We fit the submarine melt rate parameterisation to these simulations using least squares, obtaining $\dot{m} = 0.732\, q^{0.31}\, \mathrm{TF}^{1.19}$ (Extended Data Fig. 3d). The exponents of q and TF are very similar to those obtained in previous studies [e.g. 9]. Note that the maximum submarine melt rate depends very weakly on grounding line depth once thermal forcing is assumed to be constant (Extended Data Fig. 3c) and so grounding line depth does not appear explicitly in the parameterisation.

To apply the parameterisation when given a total subglacial discharge Q we have to assume a value for the line plume width w. This is rather uncertain, but undercut notches in calving fronts are found to be a few hundred metres wide [30] while [31] have estimated a similar width based on observations of the mixing between subglacial discharge and ambient fjord waters. In this study we therefore assume a width $w = 200$ m. The final melt rate parameterisation is then

$$\dot{m} = 0.732\, q^{0.31}\, \mathrm{TF}^{1.19} = 0.732\, (Q/200)^{0.31}\, \mathrm{TF}^{1.19} = 0.142\, Q^{0.31}\, \mathrm{TF}^{1.19} \qquad (3)$$

that is, $\kappa = 0.142$, $\alpha = 0.31$ and $\beta = 1.19$. Due to a lack of observations with which we can constrain parameters in the line plume model or the assumed width, there is clearly significant uncertainty in κ and the absolute melt rates predicted. Note, however, that the exponents α and β are not affected by this uncertainty and so while we should not read too much into the absolute melt rates, we can have more confidence in the relative variability in melt rate driven by changes in subglacial discharge or ocean thermal forcing.

By inserting monthly subglacial discharge (section S1.2) and monthly thermal forcing (section S1.3) into Eq. 3, we are able to estimate monthly submarine melt rate at each of Greenland's tidewater glaciers over the time period 1979-2018.

S2.2 Splitting out variability due to atmosphere and ocean

To assess the relative importance of subglacial discharge and thermal forcing variability, we estimate submarine melt rate allowing (i) subglacial discharge to vary but maintaining thermal forcing in its 1979-1993 state, (ii) allowing thermal forcing to vary but maintaining subglacial discharge in its 1979-1993 state and (iii) allowing both factors to vary. Since both thermal forcing and subglacial discharge vary seasonally, we create a baseline seasonal cycle for each by taking the average value over each month in the period 1979-1993. By 'maintaining thermal forcing in its 1979-1993' state we then mean that the thermal forcing time series for each year in 1979-2018 is just the baseline seasonal cycle.

We can then derive expressions for the percentage changes shown in Fig. 3 of the main article. We express the subglacial discharge and ocean thermal forcing as a sum of their baseline seasonal cycle (denoted by subscript '0') and an anomaly (denoted by primes)

$$Q(t) = Q_0(t) + Q'(t) \qquad \text{and} \qquad \mathrm{TF}(t) = \mathrm{TF}_0(t) + \mathrm{TF}'(t) \tag{4}$$

By definition, Q_0 and TF_0 have seasonal variability but no interannual trend. The percentage changes shown in Fig. 3e-h of the main article are smoothed versions of

$$100\% \times \frac{Q'}{Q_0} \qquad \text{and} \qquad 100\% \times \frac{\mathrm{TF}'}{\mathrm{TF}_0} \tag{5}$$

When ocean thermal forcing is held constant in its 1979-1993 state, the submarine melt rate is given by

$$\dot{m} = \kappa Q^\alpha\, \mathrm{TF}_0^\beta = \kappa Q_0^\alpha \left(1 + \frac{Q'}{Q_0}\right)^\alpha \mathrm{TF}_0^\beta = \dot{m}_0 \left(1 + \frac{Q'}{Q_0}\right)^\alpha \tag{6}$$

where $\kappa = 0.142$, $\alpha = 0.31$, $\beta = 1.19$ and $\dot{m}_0 = \kappa Q_0^\alpha \mathrm{TF}_0^\beta$ is the 1979-1993 baseline submarine melt rate. The fractional change in submarine melt rate arising due to subglacial discharge is therefore

$$\frac{\dot{m} - \dot{m}_0}{\dot{m}_0} = \left(1 + \frac{Q'}{Q_0}\right)^\alpha - 1 \approx \alpha \frac{Q'}{Q_0} \tag{7}$$

where the last approximation holds only when Q' is small compared to Q_0 and is not used in Figs. 3a-d of the main article.

Similarly, when subglacial discharge is held in its 1979-1993 state, the fractional change in submarine melt rate arising due to ocean thermal forcing is

$$\frac{\dot{m} - \dot{m}_0}{\dot{m}_0} = \left(1 + \frac{\mathrm{TF}'}{\mathrm{TF}_0}\right)^{\beta} - 1 \approx \beta \frac{\mathrm{TF}'}{\mathrm{TF}_0} \tag{8}$$

where again the last approximation holds only when TF' is small compared to TF_0 and is not used in Figs. 3a-d of the main article.

When both subglacial discharge and ocean temperature vary, the fractional change in submarine melt rate is

$$\frac{\dot{m} - \dot{m}_0}{\dot{m}_0} = \left(1 + \frac{Q'}{Q_0}\right)^{\alpha} \left(1 + \frac{\mathrm{TF}'}{\mathrm{TF}_0}\right)^{\beta} - 1 = \alpha \frac{Q'}{Q_0} + \beta \frac{\mathrm{TF}'}{\mathrm{TF}_0} + \text{higher order terms} \tag{9}$$

If the perturbations in discharge and thermal forcing are small compared to their 1979-1993 states then the change in melt rate due to both discharge and ocean temperature is approximately the sum of the changes due to each variable individually. In this case, since $\alpha = 0.31$ and $\beta = 1.19$, a fractional change in thermal forcing has around 4 times more impact on submarine melting than the same fractional change in subglacial discharge. Note also that all of the fractional changes in melt rate are independent of the coefficient κ that is poorly constrained by observations.

S3 Glacier model

S3.1 Model description

We use a simple two-stage glacier model adapted from [32] to estimate the impact of variability in submarine melt rate on glacier retreat, ice discharge and mass loss. Our intention here is to provide a simple translation of submarine melt variability to glacier mass variability. As such, we do not attempt to model in detail the evolution of specific glaciers, rather we ask how an idealised glacier would evolve when subject to the median submarine melt rate in the given region.

Following [32] the two-stage model consists of an glacier interior of length L, ice thickness H and surface mass balance P. The glacier interior delivers an ice flux Q to the glacier grounding line/terminus zone, which has ice thickness h_g, while the grounding line/calving flux is Q_g. Imposing mass conservation for the interior and terminus regions results in [32]

$$\frac{dH}{dt} = P - \frac{Q_g}{L} - \frac{H}{h_g L}\left(Q - Q_g\right) \tag{10}$$

$$\frac{dL}{dt} = \frac{1}{h_g}\left(Q - Q_g\right) \tag{11}$$

Following [32], the interior flux is estimated by assuming a dominant balance between the driving stress and the basal friction, giving

$$Q = \nu \frac{H^{2n+1}}{L^n} \tag{12}$$

where $n = 3$ is the Glen's flow law exponent and ν is a constant of proportionality that will here be treated as a tuneable parameter (it can be expressed in terms of the basal slipperiness, so that tuning ν is equivalent to tuning the basal slipperiness). Boundary-layer analysis of the grounding zone [e.g. 33] suggests that the grounding line flux can be expressed as a power of the grounding line ice thickness. To this we additionally account for the removal of ice due to submarine melting at a rate $\dot{m}$ and the potential amplification (θ) of calving due to submarine melting (sometimes called the calving multiplier effect). The grounding line/calving flux is therefore written

$$Q_g = \Omega h_g^\gamma + (1+\theta)\dot{m}h_g \tag{13}$$

where Ω and γ are constants that in general depend on the choice of sliding law and whether there is an ice shelf providing buttressing to the grounding line. Here, we follow [33] in taking $\gamma = (m+n+3)/(m+1)$, where $m = 1/n$ is the exponent of the Weertman sliding law. With $n = 3$ this gives $\gamma = 19/4$ so that calving flux is a sensitive function of ice thickness. The coefficient Ω can be expressed in terms of the basal slipperiness and the ice softness [33], but is here treated as a tuneable parameter (once again this is equivalent to tuning the ice softness). The model is closed by assuming that the ice thickness in the terminus zone h_g is the flotation thickness

$$h_g = -(\rho_w/\rho_i)b_L \tag{14}$$

where b_L is the water depth at the terminus (note $b_L < 0$), $\rho_w = 1028\ \mathrm{kg\,m^{-3}}$ is the density of ocean water and $\rho_i = 917\ \mathrm{kg\,m^{-3}}$ is the density of ice.

In sum, the only differences compared to the model presented in [32] are the addition of the submarine melt rate and calving amplification to the calving flux (Eq. 13). The model applies equally to glaciers with an ice shelf provided that the ice shelf provides little buttressing to flow, which is a simplifying assumption but affects very few of the glaciers in the dataset.

S3.2 Running the model

To initialise the glacier model, we assume that the glacier is close to steady state during 1979-1993. Limited terminus position observations [e.g. 34] and ice discharge time series [e.g. 35] during this time period together with relatively constant submarine melt rate (Fig. 3 of the main article) suggest this is a reasonable assumption. Suppose, for a given ice sheet region, the mean submarine melt rate during 1979-1993 is $\dot{m}_0$. The glacier model then has 6 remaining free parameters (Table S2), which can be taken as initial ice velocity $v_0 = f_0\,(1+\theta)\,\dot{m}_0$, initial calving front ice thickness h_{g0}, initial inland ice thickness H_0, initial glacier length L_0, bed slope b_x and calving rate multiplier θ.

Once these quantities have been defined, the value of Ω follows from Eq. 13

$$v_0 h_{g0} = \Omega h_{g0}^\gamma + (1+\theta)\dot{m}_0 h_{g0} \tag{15}$$

the value of ν follows from Eqs. 11 & 12

$$v_0 h_{g0} = \nu \frac{H_0^{2n+1}}{L_0^n} \tag{16}$$

the surface mass balance P is given by

$$P = \frac{v_0 h_{g0}}{L_0} \tag{17}$$

Parameter	Values
Initial ice velocity factor f_0	$1, 5$
Initial calving front ice thickness h_{g0} (m)	$200, 700$
Initial inland ice thickness H_0 (m)	$1000, 3000$
Initial glacier length L_0 (km)	$50, 200$
Bed slope b_x	-$[5, 20] \times 10^{-4}$
Calving rate multiplier θ	$0, 1$
Simulations	$2^6 = 64$

Table S2: Parameter values used in the simple glacier model. Each parameter set is used 3 times (varying discharge, varying ocean temperature and varying both) for each ice-sheet region.

and the bed topography is given by

$$b(x) = -\frac{\rho_i}{\rho_w} h_{g0} + b_x(x - L_0) \tag{18}$$

Rather than trying to model a specific glacier, we instead test 2 values of each of the 6 free parameters, giving a set of $2^6 = 64$ simulations for each submarine melt rate time series (Table S2). Through varying these 6 parameters, the 64 simulations cover differing sizes of glacier, differing bed slopes, differing fraction of terminus ice flux that is accounted for by submarine melting and differing sensitivity of calving to submarine melting. Together these parameters essentially determine the response time of the glacier [32] while $1/f_0$ determines the initial fraction of the ice flux that is removed by submarine melting and calving driven by submarine melting. Note that by assuming a steady state exists during 1979-1993, we require $f_0 \geq 1$, as otherwise mass loss due to submarine melting would exceed the ice flux.

With this initialisation, the model is run forwards from 1979 to 2018, forced by monthly submarine melt rate from each glacier in the dataset (section S1.1). We consider three submarine melt rate time series per glacier, consisting of the cases when (i) the atmosphere varies, (ii) the ocean varies, and (iii) both vary. To isolate the response of the glacier to submarine melting, we hold the surface mass balance constant throughout.

In total we then have 64 simulations for each of 123 glaciers and for each of 3 submarine melt scenarios. These are then grouped into ice sheet regions for the analysis (e.g. Fig. 4 of the main article). As such, these simulations are intended to capture the response of an idealised glacier in a given region to submarine melting, sampling variability in this response due to differing glacier characteristics and due to variability in submarine melting within a region.

S3.3 Interpreting the results

An example for a single simulation for South Greenland is illustrated in Extended Data Fig. 4. The submarine melt rate (Extended Data Fig. 4a) is highly seasonal, resulting in seasonality in ice discharge (Extended Data Fig. 4b) and terminus position (Extended Data Fig. 4c). Since we are here interested in the annual to decadal sea level contribution, we smooth over the seasonality with a 4-year lowess filter. The cumulative sea level contribution per unit width of glacier is proportional to

$$\mathrm{SLR} = \int_{1979}^{t} \frac{d(\mathrm{VAF})}{dt'}\,dt' = \int_{1979}^{t} (PL - Q)\,dt' \tag{19}$$

as shown in Extended Data Fig. 4d.

In order to facilitate comparison across the whole ensemble, we normalise the ice discharge, retreat and sea level contribution (Extended Data Fig. 4e-g). We emphasize that different ensemble members can have larger or smaller sea level contributions under the same submarine melt rate forcing; for example, glaciers with a larger initial ice flux typically show larger sea level contribution. Here, however, we are interested only in how sea level contribution at a given glacier changes due to atmospheric or ocean-sourced variability in submarine melting. As such, the ice discharge is normalised ($\tilde{Q}$) to its mean 1979-1993 value

$$\tilde{Q} = \frac{Q}{v_0 h_{g0}} \tag{20}$$

The retreat is normalised ($\tilde{L}$) as

$$\tilde{L} = \frac{L - L_0}{R_b} \tag{21}$$

where R_b is the total retreat between 1979 and 2018 obtained in the simulation with both atmospheric and oceanic variability in submarine melting. The dynamic sea level contribution is normalised ($\tilde{\mathrm{SLR}}$) as

$$\tilde{\mathrm{SLR}} = \frac{\mathrm{SLR}}{\mathrm{SLR}_b} \tag{22}$$

where SLR_b is the total 1979-2018 sea level contribution in the simulation with both atmospheric and oceanic variability in submarine melting.

Due to this normalisation, all ensemble members which account for both atmospheric and ocean-sourced variability in submarine melting have a normalised sea level contribution equal to 1 in 2018 (e.g. Extended Data Fig. 4g and Fig. 4 of the main article). Due to differing glacier response times or glacier-to-glacier variability in submarine melt rate within a region, ensemble members may however have different normalised sea level contribution earlier in the time period, giving rise to an envelope of sea level contribution (Fig. 4 of the main article).

Similarly, simulations accounting separately for the atmosphere or ocean-sourced variability in submarine melting may have different normalised sea level contribution depending on the relative importance of the atmosphere and ocean to the submarine melt time series in question (Extended Data Fig. 4). Note that the sum of the sea level contributions due to atmosphere and ocean individually is not necessarily equal to the sea level contribution due to both atmosphere and ocean. This arises due to (i) non-linearity in the submarine melt rate parameterisation and (ii) non-linearity in the glacier response to submarine melting.

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