Future Projections of Petermann Glacier Under Ocean Warming Depend Strongly on Friction Law

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Abstract Basal friction heavily controls the dynamics of fast-flowing glaciers. However, the best approach to modeling friction is unclear, increasing uncertainties in projections of future mass loss and sea-level rise. Here, we compare six friction laws and evaluate them for Petermann Glacier in northern Greenland, using a higher order three-dimensional ice-sheet model. We model glacier retreat and mass loss under an ocean-only warming until year 2300, while not considering the effects of a future warmer atmosphere. Regardless of the friction law, we find that breakup of Petermann’s ice shelf is likely to occur within the next decades. However, future grounding-line retreat differs by 10s of km and estimates of sea-level rise may quadruple, depending on the friction law employed. A bedrock ridge halts the retreat for four of the laws, and Petermann retreats furthest when applying a Budd or a Coulomb-type “till law.” Depending on the friction law, sea-level contributions differ by 133% and 282% by 2300 for 2°C and 5°C ocean warming scenarios, respectively.

Plain Language Summary Rocks, sediments, and water under glaciers can influence glacier speed and their response to climate change. How to best represent these effects in computer models is unclear, which magnifies uncertainties in our projections of future sea-level rise. We, therefore, compared six ways to represent the under-ice environment in a computer model, and simulate the future of Petermann Glacier, northern Greenland’s largest and fastest glacier. We study how the glacier reacts to the 2°C–5°C warmer ocean expected, if carbon emissions continue, while not considering the effects of a future warmer atmosphere. The experiments indicate that Petermann’s ice shelf, the glacier’s floating extension, will likely breakup within the next decades. Our estimates of future sea-level rise by year 2300 from Petermann Glacier also quadruple, depending on how the under-ice conditions is treated in the model. This dramatically highlights the limitations of current ice-sheet models and projections, and pinpoints the need to improve the representation of the under-ice environment and marine-based ice sheets in further research.

1. Introduction

Fast-flowing ice streams and outlet glaciers have long been known to be influenced by friction between basal ice and glacier beds (Dowdeswell et al., 2016; Stokes & Clark, 1999), specifically the presence of basal deformable till (Alley et al., 1986). However, because glacier beds are largely inaccessible for in situ observations, the physical and hydrological properties and associated frictional effects are poorly known, as well as their spatial and temporal variability. As a result, a diversity of so called “friction laws” in ice-sheet models exist, generally based on theoretical arguments (e.g., Schoof, 2007; Tsai et al., 2015; Weertman, 1957), numerical experiments (e.g., Gagliardini et al., 2007; Helanow et al., 2020), or laboratory experiments (e.g., Budd et al., 1984; Iverson et al., 1998).

Significant progress has been made to infer basal friction through inverse modeling (e.g., Gillet-Chaulet et al., 2012; MacAyeal, 1992, 1993; Morlighem et al., 2010; Pollard & DeConto, 2012). Remotely sensed ice velocities are typically used as a target in an inversion algorithm, where a friction parameter is tuned to minimize data-model misfit. Such assimilation, however, cannot determine whether the friction law is suitable for the particular glacier studied. This would require observations at different time periods of varying basal velocities, stresses, and water pressure (Gillet-Chaulet et al., 2016).
Several friction laws have been compared in both synthetic (Brondex et al., 2017; Gladstone et al., 2017) and real-world settings (Brondex et al., 2019; Joughin et al., 2019; Lilien et al., 2019; Nias et al., 2018). These efforts have focused on West Antarctica, where 50 km wide glaciers are buttressed by large ice shelves. This setting differs from Greenland’s narrower outlet glaciers and ice tongues. Previous studies have also focused on the near-future, even though ice-sheet mass loss will continue for centuries, if global carbon emissions are left unabated (Aschwanden et al., 2019).

Here, we test six friction laws for Petermann Glacier and assess future mass loss and associated sea-level rise under warming ocean waters. Petermann Glacier is the largest (71,305 km$^2$) and the fastest outlet glacier in northern Greenland (Hill et al., 2017; Rignot & Mouginot, 2012) (Figure 1). Petermann and Ryder Glacier, located further to the north, are the only outlet glaciers in northwestern Greenland that have large remaining floating ice tongues (Hill et al., 2017) that buttress upstream ice. Petermann lost nearly 35 km (40%) of its ice tongue in two major calving events in 2010 and 2012 (Münchow et al., 2014). While the 2010 calving event did not have a significant impact on ice flow, the 2012 event caused an ice flow speedup during 2012–2016 (Rückamp et al., 2019). There is concern that further calving events may accelerate regional mass loss from this sector of Greenland.

We exploit the exceptional data set recovered from the Petermann 2015 Expedition (Heuzé et al., 2017; Jakobsson et al., 2018). The overdeepened Petermann Fjord has a maximum water depth of 1158 m and a large glacial sediment deposition forming a sill at the fjord mouth with its deepest passage of about 443 m (Jakobsson et al., 2018). Oceanographic stations have shown that warmer (T > 0°C) and relatively salty (S > 34.6 psu) Atlantic water enters the fjord across the sill (Münchow et al., 2016). The warm water inflow is likely responsible for the recent ice-tongue thinning that led to the large calving events in 2010 and 2012 (Münchow et al., 2014).

In this study, we compare (a) a linearly viscous “Budd” friction law, with a friction parameter found by inversion, and (b) a “till friction law”, with a friction parameter defined ad hoc and justified heuristically. We also present two approaches, (c) till-assimilation and (d) till-assimilation-N-Budd, which assimilates the “till friction law” using two different formulations of the under-ice effective pressure. Finally, we include (e) a Schoof friction law, an increasingly prevalent alternative to simulate ice sheet flow over till, and (f)
a nonlinear Weertman law, which in theory is applicable to sediment-free bedrock but is (still) commonly used in ice-flow modeling.

Previous studies have either neglected calving, or prescribed calving rates (e.g., Hill et al., 2018, 2021; Rückamp et al., 2019), and the most recent research is largely performed with two-dimensional ice dynamics (two horizontal dimensions, vertically integrated), typically using the shallow-shelf approximation (e.g., Brondex et al., 2019; Hill et al., 2021; Joughin et al., 2019; Rückamp et al., 2019). Here, we have an explicit calving model (Morlighem et al., 2016) use three-dimensional higher order ice dynamics, and allow Petermann to respond and adjust to imposed external forcing and involved feedbacks over almost 300 years. In addition, the three “till-friction” laws included (till-elevation, till-assimilation, till-assimilation-N-Budd, see Section 2.2) have previously not been compared (Brondex et al., 2019).

The study is structured as follows. First, we present the data used and the friction laws studied (Section 2) before describing the model, experimental setup, and external forcing (Section 3). The inferred basal conditions and ice flow are outlined in Section 4.1, the modeled present-day behavior in Section 4.2, and future response to ocean warming in Section 4.3. Finally, we discuss the contrasting responses under the different friction laws (Section 5.1), and outline implications for future stability of marine outlet glaciers and model projections (Section 5.2). Detailed context is provided in the supplementary material, including additional figures of the model setup and results, as well as animations of Petermann’s retreat.

2. Data and Method

2.1. Surface and Bed Topography

We use BedMachine v3 (Morlighem et al., 2017) to represent the bathymetry and subglacial topography, which includes the multibeam data from Jakobsson et al. (2018). Observed ice surface elevation as well as the ice/ocean/land mask is taken from the Greenland Ice Sheet Mapping Project (GIMP, 2008, 2010; Howat et al., 2014). Floating ice is inferred from InSAR grounding lines (Howat, pers. comm.). All the above data are contained within the freely accessible BedMachine v3 data product.

2.2. Basal Friction

Using the Ice-Sheet and Sea-level System Model (ISSM; Larour et al., 2012), we simulate present and future behavior of Petermann Glacier under the following six basal friction laws:

2.2.1. Budd Law Inversion (“Budd”)

We use a “Budd” (B) linearly viscous friction law (Budd et al., 1984), and calculate basal drag $\tau_{b,B}$ as

$$\tau_{b,B} = -\alpha^2 N u_b,$$

where $\alpha$ is a friction parameter, $u_b$ is basal velocity, and $N$ the effective pressure. We assume a perfect hydrological connection between the subglacial drainage system and the ocean, defining $N$ as the difference between ice overburden and hydrostatic pressure: $N = \rho_i g H + \rho_w g z_b$, where $\rho_i$, $g$, $H$, $\rho_w$, and $z_b$ are ice density, gravitational acceleration, ice thickness, seawater density, and bed elevation, respectively. By assuming this, we neglect variations in basal pore water pressure, that will act to reduce the effective pressure exerted by the overlying ice. This formulation does thus not account for input of surface melt, which would lead to seasonal variations of $N$.

We find the basal friction parameter $\alpha$ using an adjoint method, minimizing a cost function that quantifies the misfit between modeled and observed (Joughin et al., 2010; MEaSUREs, 2016) ice velocities. The cost function used is a combination of an absolute and logarithmic misfit between modeled and observed ice velocities $u$: 
The absolute and logarithmic terms capture fast-flowing and slow-flowing areas well, respectively. \( c \) is a regularization term to prevent singularities, and \( \gamma_1 = 2000, \gamma_2 = 1, \) and \( \gamma_3 = 6 \times 10^{-7} \) are respective weights for each of the terms in the cost function Equation 2.

### 2.2.2. Till Friction: Pseudo-plastic Power Law (“Till-Elevation”)

The “till friction law” is commonly used for modeling both past (Seguinot et al., 2016), present, and future ice sheets, including Greenland (Aschwanden et al., 2016, 2019; Bueler & Pelt, 2015), and Antarctica (Golledge et al., 2015, 2019).

This friction law assumes flow over a (partly) till-covered bed. The concept builds on laboratory experiments on weak till from the Siple Coast ice streams in Antarctica (Tulaczyk et al., 2000, 2001), where the bed provides little resistance to flow. As meltwater infiltrates basal sediments, or when basal sediments are already saturated and located below sea level, the underlying bedrock gets covered by a water-saturated till. This till is often considered pseudoplastic, that is, no deformation occurs until the basal shear stress (drag) \( \tau_b \) exceeds a certain yield stress \( \tau_c \), given by a Mohr-Coulomb criterion (Terzaghi, 1943)

\[
\tau_c = c_0 + N_{Coulomb} \tan(\phi),
\]

where \( c_0 \) is the till cohesion, \( N_{Coulomb} \) is the effective pressure (Equation 4 below), and \( \phi \) a till friction angle. Cohesion tends to be negligible and is set to 0.

The till friction angle \( \phi \) (Equation 3) represents the ability to withstand basal shear stress and is used as a tuning parameter, sometimes scaled with bedrock elevation (Aschwanden et al., 2016; Golledge et al., 2015, 2019). This scaling is based on the assumption that lower elevations are more likely to contain weak deformable sediments, perhaps of marine origin, while the bed is stronger at higher elevation, where sediments supposedly are less prevalent or absent. This ad hoc approach has some intuitive merit, yet the underlying assumptions have not been validated by observations. For Petermann, observed velocities are to a first order indeed controlled by bedrock elevation (Figure S2), though with considerable scatter.

Following Bueler and Pelt (2015), the effective pressure \( N_{Coulomb} \) is expressed as a function of basal water thickness and till properties:

\[
N_{Coulomb} = \delta P_0 10^{(e_0/C_c) [1-(W/W_{max})]}.
\]

Here, \( W_{max} \) is the maximum water thickness, \( \delta \) is a lower limit of the effective pressure expressed as a fraction of ice overburden pressure \( P_0 \), \( e_0 \) is the till reference void ratio, and \( C_c \) is a till compressibility coefficient (Tulaczyk et al., 2000). Parameter values are listed in Table S1. Basal water thickness \( W = 2 \) is kept fixed in time and space.

Schoof and Hindmarsh (2010) proposed a pseudoplastic friction power law that defines the basal drag \( \tau_{b,T} \) as a function of the yield stress \( \tau_c \) (Equation 3) and the sliding velocity \( u_b \):

\[
\tau_{b,T} = -\tau_c \left| \frac{u_b}{u_0 (1+q)} \right| u_b^{q-1}.
\]

Here, \( u_0 = 100 \text{ m a}^{-1} \) is a threshold velocity above which basal sliding is assumed to occur. A pseudoplasticity exponent \( q = 0.6 \) determines the nonlinearity of the till friction law (see Equation 5). To facilitate
In the third approach, we combine the assimilation method used for Budd with the till friction law. The friction angle $\phi$ is treated as an unconstrained parameter and found such that the basal drag calculated by the till friction law (Equation (5)) matches the initial drag obtained by the Budd inversion (Equation (1)), that is $\tau_{b,T}^* = \tau_{b,B}$. We then insert the inferred $\tau_{b,B}$ into the till friction law (Equation (5)), and solve for the “assimilated” yield stress $\tau_c^*$:

$$\tau_c^* = \frac{1}{|u_b|} \left( \left| u_b \right| (1 - \eta)^{\gamma_b} \right)$$  \hspace{1cm} (6)

To find the friction angle $\phi$, we insert the assimilated yield stress $\tau_c^*$ found in Equation (6) into the Mohr-Coulomb criterion (Equation (3)), and solve for the “assimilated” $\phi^*$:

$$\phi^* = \tan^{-1} \left( \frac{\tau_c^*}{N_{\text{Coulomb}}} \right)$$  \hspace{1cm} (7)

This is a simple way to constrain the friction angle $\phi$ in the till friction law (Equation (5)) and allows for a direct comparison between the Budd and till friction laws.

2.2.4. Assimilated till Friction with Budd’s Effective Pressure (“Till-Assimilation-N-Budd”)

In this approach, we invert for the friction angle $\phi$ as outlined in the till-assimilation approach (Section 2.2.3), but formulate the effective pressure differently. We replace $N_{\text{Coulomb}}$ in Equation (7) with the formulation used in the Budd friction law: $N = \rho g H + \rho_e g z_e$, which depends only on glacier and bed geometry and assumes perfect hydrological connectivity with the ocean (Section 2.2.1).

2.2.5. Assimilated Weertman-Coulomb Friction: Schoof Law (“Schoof”)

We also include a Schoof friction law, which is based on theoretical work (Gagliardini et al., 2007; Schoof, 2005) but has only recently been revisited for modeling studies of real glaciers (Brondex et al., 2019; Joughin et al., 2019; Lilien et al., 2019; Nias et al., 2018). Basal drag $\tau_{b,S}$ is here defined as

$$\tau_{b,S} = - \frac{C_i \left| u_b \right|^{m-1} u_b}{1 + \left( \frac{C_i}{C_{\text{max}}} \right)^{1/m} \left| u_b \right|^m}$$  \hspace{1cm} (8)

where $C_i$ is a friction parameter and $m = 1/n$, where $n = 3$ is the exponent in Glen’s flow law. By construction, the Budd law (Equation (1)) does not include any bounds for the local basal shear stress for a given effective pressure, meaning that at any point of the ice/bed interface, the bed can potentially provide full resistance to the local driving stress.

Conversely, for a given effective pressure, the Schoof law defines an upper limit $\tau_b/N < C_{\text{max}}$ above which excess driving stress has to be balanced by other stress components. This limit, called Iken’s bound, originates from the existence of water-filled cavities at the bed, whose size is bounded by the leeside (downglacier) steepness of bedrock obstacles (Iken, 1981). Basal motion under the Schoof law behaves differently depending on the effective pressure $N$. As for the Budd law but in contrast to for the sill-friction law, we calculate $N$ assuming perfect hydrological connectivity to the ocean (see Section 2.2.1). For larger $N$, basal motion occurs under a typical hard-bedded Weertman regime (Weertman, 1957), for which the basal drag depends on the basal velocity in a power law form $\tau_{b,S} \sim C_i \left| u_b \right|^{\gamma_b}$. Conversely, the Schoof law induces a Coulomb-type plastic flow regime at very low effective pressure $N$, typical for areas of fast flow and near the grounding line, with basal drag proportional to the effective pressure in the form $\tau_{b,S} \sim C_{\text{max}} N$ (Brondex et al., 2017). The latter regime is thus similar to the till friction law described in Sections 2.2.2 and 2.2.3 above.
Following Brondex et al. (2019), the friction parameter $C_s$ is “assimilated” in an analogous manner as for $\phi$ in the till-assimilation law, and we find $C_s$ analytically such that the Schoof friction law (Equation 8) gives the same basal drag as the Budd law, that is, $\tau_{b,s} = \tau_{b,B}$. Solving for $C_s$ in Equation 8 gives:

$$C_s = \left\{ \frac{\tau_{b,B}}{u_b} \right\}^{1/m} \left\{ 1 - \frac{1}{C_{max} N} \right\}^{1/m}.$$  \hspace{1cm} (9)

We set $C_{max} = 0.6$ as spatially constant, and set $C_s$ in ice-free and floating areas to an arbitrary low value of $10^{-3} \text{ MPa}^{-1/3} \text{a}^{1/3}$, following Brondex et al. (2019). The same value for $C_s$ is used for localized grounded areas where $|\tau_{b,B}| > C_{max}/N$, which at Petermann amounts to a few localized areas across a few km from the grounding line.

### 2.2.6. Assimilated Weertman Law (“Weertman”)

Finally, we include a nonlinear Weertman law, which assumes ice flow over a hard bed by a combination of viscous creep and regelation (Weertman, 1957). While there are newer alternative descriptions of friction available (e.g., the friction laws outlined above), this law is still common in the literature, perhaps due to its simplicity. Basal drag $\tau_{b,W}$ is calculated as

$$\tau_{b,W} = -C_w u_b^m,$$  \hspace{1cm} (10)

where $m = 1/n$ and $n = 3$ is Glen’s flow parameter. This law does not include any explicit dependence on the effective pressure $N$, whose influence on flow is instead subsumed into the friction parameter $C_w$. This parameter varies spatially, and we find $C_w$ analytically such that the Weertman friction law (Equation 10) gives the same basal drag as the Budd law, that is $\tau_{b,W} = \tau_{b,B}$. Solving for $C_w$ in Equation 10 we thus get

$$C_w = -\frac{\tau_{b,B}}{u_b^m}.$$  \hspace{1cm} (11)

### 2.3. Submarine and Surface Mass Balance

We use depth-dependent annual submarine melt rates, based on modeling and observations of Petermann’s ice tongue (Cai et al., 2017; Rignot & Steffen, 2008). Present-day melt rates vary linearly from zero at depths shallower than 200 m, to 30 m/a at 600 m depth (Figure 2). These melt rates are used for present-day relaxation (Section 3.3), and are subsequently perturbed in our “future” simulations (Section 3.4). Horizontal melt at the glacier front is prescribed to 30 m/a and is applied if the calving front is grounded (no ice tongue present). No frontal melt is applied when floating ice is present; submarine melt depends on a subglacial discharge plume rising along the shelf draft, and associated melt is close to zero at the front of floating termini. However, throughout our simulations, the ice tongue never disappears completely, so frontal melt is never applied in practice.

We use an average climatic surface mass balance (1979–2014) from the regional climate model MAR 3.5.2 (Fettweis et al., 2017). Our climatic surface mass balance is calculated from monthly means for each year.
3. Model Setup and Experiments

3.1. Ice Dynamics and Calving

Three-dimensional higher order physics (Blatter, 1995; Pattyn, 2003) is used to simulate ice flow on a finite-element mesh constructed for accurate modeling of ice-shelf melt and grounding-line dynamics, which is resolved at subelement resolution and based on a flotation criterion (Seroussi et al., 2014).

The model domain, shown in Figure 1, is delimited to represent Petermann’s entire drainage basin, using the flow directions of observed velocities (Joughin et al., 2010; MEaSUREs, 2016). Offshore of the observed ice-shelf front, the domain is extended to allow for potential ice-front advance. The mostly ice-free fjord walls along the western flank of Petermann’s ice shelf are also included in the domain, as well as local ice caps and tributaries east of the fjord that are connected to Petermann Glacier. The model mesh consists of 255,000 elements distributed over seven vertical layers. The mesh resolution varies from 500 to 10,000 m based on bed topographic gradients, and we refine the mesh resolution to 500 m, where observed velocities exceed 500 m/a. In transient experiments, time steps of 0.01 (3.65 days) to 0.05 years (18.25 days) are needed to maintain numerical stability, depending on the friction law.

We simulate calving front evolution using a tensile von Mises calving law Morlighem et al. (2016), where the calving rate \( c \) depends on the tensile stress:

\[
\dot{c} = \left| \mathbf{u} \right| \frac{\sigma}{\sigma_{\text{max}}},
\]

where \( \sigma \) is the von Mises tensile stress, which depends only on the tensile strain rate (see Morlighem et al. (2016) for details), and \( \sigma_{\text{max}} \) is a stress threshold. We use \( \sigma_{\text{max}} = 1 \) MPa for grounded ice, and calibrate \( \sigma_{\text{max}} \) to 300 kPa for floating ice by reproducing the present-day ice-shelf margin after a 50-year transient relaxation (Section 3.3). Calving is not allowed for ice thickness smaller than 100 m, as this is found to prevent unrealistic calving model behavior along the fjord walls, where calving can occur in the model due to very thin ice, while in reality, ice is unlikely to break off in other areas than at the ice front.

We use an Arrhenius law relating ice temperature to ice hardness. The ice-shelf viscosity parameter, \( B \), is estimated using an inverse method minimizing the absolute misfit between modeled and observed velocities, analogous to the friction inversion described in Section 2.2. Ice viscosity for grounded ice is set spatially invariable, corresponding to an ice temperature \( T_{\text{ice}} = -12^\circ\text{C} \) based on Greenland-wide ISMIP6 experiments (Goelzer et al., 2020).

3.2. Stress Balance Experiments (Fixed Geometry)

In a first set of experiments, ice velocities are calculated on a fixed ice geometry for the six friction laws detailed in Section 2.2. This means a Budd law basal drag inversion (Section 2.2.1), assimilation for the till-assimilation laws (Section 2.2.3 and 2.2.4), and the Schoof (Section 2.2.5) and Weertman (Section 2.2.6) laws, and taking friction parameters at face-value for the till-elevation law. These experiments are time-independent, so no surface mass balance, ocean forcing, or calving model is needed. Model performance is assessed using the root mean square error (RMSE) between modeled and observed (Joughin et al., 2010; MEaSUREs, 2016) ice velocities.

3.3. Transient “Relaxation” Experiments

In the second set of experiments, the ice-sheet model is forced with present-day, fixed atmosphere and ocean forcing (Section 2.3) and the glacier geometry can freely evolve, including the grounding line. While the surface mass balance is prescribed, submarine melt rates are updated continuously according to the evolving geometry of the ice-shelf draft.

These transient simulations first run over 50 years under constant present-day forcing and with a fixed calving front. Subsequently, the calving front is “released”, and relaxation continues for another 50 years of constant present-day forcing. The relaxation aims to eliminate any nonphysical ice flux divergence due to
At the end of relaxation, a quasisteady state with a small model drift is reached (for the modeled Petermann Glacier $|dH/dt| < 1 \text{ m/a}$, $|du/dt| < 5 \text{ m/a}^2$; for adjacent local small glaciers $<2 \text{ m/a}$ and $<10 \text{ m/a}^2$). We also tested to prolong this spin-up phase by 100 years (150 in total) to be closer to steady-state, but the effects on the subsequent future evolution of Petermann’s calving front and grounding line following such a prolonged spinup are negligible.

3.4. Ocean Warming Experiments

Finally, “future” sensitivity experiments are performed by instantly increasing submarine melt rates, associated with an ocean warming in a “best” to “worst case” scenario of 2, 3, 4, and 5°C (Figure 2). We also include no-warming control runs (0°C), where submarine melt rates are kept at present-day values. These experiments start from the relaxed states and are run until year 2300 (280 years). The perturbed melt rates are based on the ocean model MITgcm (Cai et al., 2017) and ocean temperatures are given relative to a 2008 temperature-salinity profile from Johnson et al. (2011). These water properties are consistent with measurements from moorings (Münchow et al., 2016), as well as CTD (conductivity, temperature, depth) casts from the Petermann 2015 Expedition (Heuzé et al., 2017). This expedition was carried out in summer, but our ocean boundary conditions are robust as an annual forcing since negligible seasonal variability of the deep-water properties occurs at Petermann (Münchow et al., 2016).

4. Results

4.1. Inferred Basal Conditions and Ice Flow

We first summarize time-independent “stress balance” experiments (Section 3.2). Further details are provided in Table 1 and Section S1.

Using Budd friction, we infer a moderately weak bed under Petermann’s ice stream ($\tau_b \sim 50–150 \text{ kPa}$; Figure S3e), while the lower ice stream and the grounding zone have weak areas ($\tau_b < 50 \text{ kPa}$) intermixed with strong “sticky spots” ($\tau_b > 150 \text{ kPa}$).

The till-elevation law gives a basal drag under the ice stream 100–200 kPa higher than under Budd (Figure S3h). Below the lower ice stream, the till-elevation law produces a weaker bed than the Budd inversion, while close to the grounding line, the opposite is true. Off the ice stream margins and upglacier of the ice stream onset, the two laws produce basal drag of similar magnitude (±25 kPa). The basal drag under the till-assimilation laws, the Schoof law, and the Weertman law are by construction identical to the basal drag obtained with the Budd inversion (Figure S3e), except for small areas close to the grounding line where Iken’s bound in the Schoof law is exceeded (see Section 2.2.5). Here, the friction coefficient $C_s$ thus has to be set manually, following Brondex et al. (2019). This may explain differences in RMSEs between the Schoof and Budd laws in these fixed-geometry experiments.

The ability to simulate Petermann’s ice flow in these fixed-geometry experiments varies widely depending on the friction law employed (Table 1). The Budd law produce model velocities closest to observations ($\text{RMSE} = 46 \text{ m/a}$; Figure S4a), with velocities within ±50 m/a for the ice stream and the ice shelf, except for a 50–100 m/a too fast grounding zone and similarly too low ice-shelf shear margins. In contrast, the till-elevation law renders the highest model-data mismatch ($\text{RMSE} = 246 \text{ m/a}$; Figure S4b), both for the ice shelf and grounding zone (several 100 m/a too slow) and the ice stream (100–500 m/a too fast). The till-assimilation laws (RMSE = 100 m/a and 88 m/a) alleviates some of the till-elevation issues (Figures S4c and S4d), while Schoof and Weertman performs similarly well (RMSE = 102 m/a and 115 m/a; Figures S4e and S4f). The velocity misfit distribution for three latter laws are similar, with the ice shelf and grounding zone flowing 50–200 m/a faster than observed.

Note that a larger misfit to observations for the till-elevation friction law is expected by construction, as this law is not assimilated against any observations.
Table 1
Overview of Modeling Results for the Compared Friction laws: Budd, Till-Elevation (Till-Elev.), Till-Assimilation (Till-Assim.), Till-Assimilation-N-Budd (Till-Assim-N-Budd), Schoof, and Weertman

|                      | Budd   | Till-elev. | Till-assim. | Till-assim-N-Budd | Schoof | Weertman |
|----------------------|--------|------------|-------------|-------------------|--------|----------|
| **Basal conditions** |        |            |             |                   |        |          |
| $\tau_b$, ice stream (kPa) | 50–150 | 100–200    | = $\tau_b$,Budd | = $\tau_b$,Budd | $\approx\tau_b$,Budd | = $\tau_b$,Budd |
| $\tau_b$, gr. zone (kPa) | <50/>150 | 50–125     | = $\tau_b$,Budd | = $\tau_b$,Budd | $\approx\tau_b$,Budd | = $\tau_b$,Budd |
| **Stressbalance**    |        |            |             |                   |        |          |
| $u$ misfit (RMSE m/a) | 46     | 246        | 100         | 88                | 102    | 115      |
| $u$ misfit, ice stream (m/a) | <50 | +100–500 | <50 | <50 | <50 | <50 |
| $u$ misfit, gr. zone (m/a) | <50 | −100–500 | +50–200 | +50–200 | +50–200 | +50–200 |
| **Transient relaxation** |        |            |             |                   |        |          |
| $u$ misfit (RMSE m/a) | 26     | 73         | 25          | 25                | 18     | 59       |
| $u$ misfit, ice stream (m/a) | −<100 | −100–500 | +100–200 | +100–200 | <50 | <50 |
| $u$ misfit, gr. zone (m/a) | −<100 | −100–500 | +<100 | +<100 | +<100 | +<100 |
| $H$ misfit (RMSE m) | 22     | 62         | 28          | 17                | 22     | 62       |
| $H$ misfit, ice stream (m) | +50–75 | +50–250 | ±<50 | ±<50 | ±<50 | ±<50 |
| $H$ misfit, ice shelf (m) | +<50–200 | +>250 | +100–250 | +100–250 | +100–200 | +100–200 |
| **GL movement (km)** | 9.83   | 10.66      | 11.54       | 9.40              | 10.50  | 10.36    |
| **GL flux (Gt/a)**   | −0.07  | +0.76      | +1.64       | −0.50             | +0.60  | +0.46    |
| **GL flux misfit (%)** | −0.7   | 7.7        | 16.6        | 5.0               | 6.0    | 4.6      |

**Future ocean warming**

|                      | dGL/dt, 2020–2050, 2°C (m/a) | dGL/dt, 2020–2100, 2°C (m/a) | dGL/dt, 2100–2300, 2°C (m/a) | dGL/dt, 2020–2050, 5°C (m/a) | dGL/dt, 2020–2100, 5°C (m/a) | dGL/dt, 2100–2300, 5°C (m/a) |
|----------------------|-------------------------------|-------------------------------|-------------------------------|-------------------------------|-------------------------------|-------------------------------|
| Budd                 | 498                           | 262                           | 43                            | 666                           | 350                           | 16                            |
| Till-elev.           | 102                           | 51                            | 0                             | 137                           | 52                            | 0                             |
| Till-assim.          | 365                           | 136                           | 3                             | 365                           | 147                           | 3                             |
| Till-assim-N-Budd    | 0                             | 123                           | 518                           | 350                           | 194                           | 194                           |
| Schoof               | 365                           | 194                           | 498                           | 194                           | 146                           | 10                            |
| Weertman             | 365                           | 136                           | 365                           | 146                           | 10                            | 2                             |
| GL flux, 2100, 2°C (Gt/a) | 18.09                      | 15.24                        | 18.35                         | 15.13                         | 12.09                         | 5.8                           |
| SLR, 2100, 2°C (mm)  | 1.97                         | 1.24                          | 1.97                          | 1.24                          | 1.24                          | 1.24                          |
| SLR, 2100, 5°C (mm)  | 6.29                         | 4.48                          | 6.29                          | 4.48                          | 4.48                          | 4.48                          |

Shown are inferred basal conditions (fixed geometry; Section 4.1) and results from the transient relaxation (evolving geometry, constant forcing; Section 4.2), including characteristic basal drag ($\tau_b$) and misfits (modeled-observed) of velocity ($u$), thickness ($H$), and grounding line (GL) flux. Note that basal drag $\tau_b$ for the till-assimilation, Schoof, and Weertman laws are (approximately) identical to that of Budd by construction (see Section 2.2). For future ocean warming (evolving geometry, perturbed forcing; Section 4.3), retreat rates (dGL/dt) averaged over three different periods are shown. A retreat rate of zero indicates that a new stable GL position has been reached. The observed GL flux = 9.90 ± 0.48 Gt/a is taken from Wilson et al. (2017). The results for the best-performing friction law within each metric is boldfaced. Retreat rates, GL flux, and sea-level rise (SLR) contributions for respective friction law is shown for the end-member scenarios of ocean temperature increase.
4.2. Transient “Relaxation” Experiments

As the models for the different friction laws relax from their respective stress-balance states, several of the model-data mismatches (Section 4.1) dissipate, perhaps unexpectedly. This could be related to the relaxing surface topography, where any local high stress gradients and associated velocity misfits would be damped once the geometry is released.

The Schoof friction represents Petermann’s flow (RMSE = 18 m/a) and ice thickness (RMSE = 22 m) most closely to observations after the transient relaxation (Figures 3g and 3h), with the Budd (RMSE = 26 m/a and 22 m) and till-assimilation (25 m/a and 28 m with $N_{\text{Coulomb}}$, 25 m/a and 17 m with $N_{\text{Budd}}$) not far behind.

All laws overestimate ice-shelf thickness, suggesting that this mismatch is not mainly related to friction, but likely also to our simple linearly depth-dependent submarine-melting parameterization, or perhaps errors in the bathymetry data in the grounding zone. The performance of each friction law is further detailed in Table 1 and Section S1.

We find that the till-assimilation law simulates flow closer to observations than its till-elevation alternative (Figure S6). While this is expected by construction (Section 4.1), the improvement in performance (Figure S6) highlights where the assimilation is most needed and the till-elevation is furthest from observations. The relative misfit for the till-elevation law is largest (up to 20 times larger) in slow-flowing (non-sliding) areas above sea level (Figure S6b), including adjacent small ice masses with little importance for Petermann’s main flow. However, weak-bed (sliding) fast-flowing areas below sea level control the shelf-ice stream system. This is where the till-assimilation improves model dynamics the most (Figures S6d and S6f); the velocity misfit is reduced from $>500$ m/a ($\sim50\%$–$80\%$ too slow) to $\sim100$–$200$ m/a ($\pm20\%$).

The modeled grounding-line flux under the Budd law (9.83 Gt/a) is the closest to satellite-derived estimates and is within observational errors (9.90 ± 0.48 Gt/a, Wilson et al., 2017). The modeled flux under the
Weertman (10.36 Gt/a), Schoof (10.50 Gt/a), till-elevation (10.66 Gt/a), and the till-assimilation-N-Budd (9.40 Gt/a) laws also agrees reasonably well, while the till-assimilation law overestimates the flux by 16% (11.54 Gt/a; Table 1). Another close model estimate was provided by Hill et al. (2021) (9.85 Gt/a), while Rückamp et al. (2019) simulated the flux to 10.55 Gt/a, further away from the observed flux than the estimate from the Budd friction law in our simulations.

In summary, considering both stress balance (Section 4.1) and transient relaxation (Section 4.2) experiments, and compared to observed ice velocity, ice geometry, grounding-line position, and grounding-line flux, we find that present-day model behavior is closest to observations (lowest RMSE) for the Budd and Schoof friction laws, with till-assimilation and Weertman laws slightly behind, while the till-elevation law shows the poorest performance (highest RMSE). Still, these results are expected by construction, as the inversion methodology applied allows us to get closer to observations than does a calibration without any tuning (till-elevation law).

4.3. Ocean Warming Experiments

In response to a warming ocean and associated stronger subshelf melting, we find that Petermann’s dynamic response (Figure 4), resulting mass loss (Figures 7a–7f), and associated grounding line retreat (Figures 7g–7l) depend greatly on the friction law.

The models suggest that roughly half of Petermann’s ice shelf is lost before 2030, regardless of the friction law and temperature increase (Figure S8). For the laws with acceptable present-day performance (Budd, till-assimilation, till-assimilation-N-Budd, Schoof, Weertman), more than three-quarters of the ice shelf has collapsed by 2050, even under the lowest 2°C scenario (Figures 4a, 4e, 4g, 4i, and 4k). To further investigate ice-shelf sensitivity to ocean warming, we performed additional experiments where we changed the submarine melt rates corresponding to temperature increases of only 0.5°C and 1°C. This weaker forcing delays ice-shelf breakup a little, but still nearly half of the ice shelf is lost by 2050, even for a 0.5°C warming. This indicates that Petermann’s ice shelf currently is in a sensitive state; even a slight temperature increase will likely cause the ice shelf to breakup. Alternatively, it could also indicate that our calving law is too sensitive. While another calving law may lead to different time scales of ice-shelf breakup, there is currently no consensus for which calving law to use. Choi et al. (2018) tested several calving laws for a range of Greenland outlet glaciers and found that the von Mises calving law used here was able to simulate observed ice-front retreat more closely than other laws. However, changing the calving stress threshold ($\sigma_{\text{max}}$ in Equation 12) could affect the timescale of ice-shelf breakup. We constrained the stress threshold used here (300 kPa) by reproducing Petermann’s present-day calving front position closely in the relaxation phase (Section 3.3). Being able to reproduce the current shape of the calving front gives some confidence in our future simulations.

We find that mass loss and resulting sea-level rise estimates may quadruple from one friction law to another (Figures 7a–7f). The Budd and Schoof laws produce glacier behavior closest to reality in our present-day simulations, yet the same two friction laws give significantly different predictions for future mass loss. Across all friction laws, grounding-line retreat differs by 20 km (Figure 6) and the contribution to sea level by ~133%–282% (2.7–6.3 mm sea-level equivalent [SLE] for 2°C, 2.8–10.6 mm SLE for 5°C) by year 2300 (Figures 7a–7f). Under Budd and till-assimilation-N-Budd, stronger retreat, acceleration, and associated 250 m dynamic thinning that propagates 50 km inland (Figures 4 and 8) cause more grounded-ice loss and therefore higher contribution to sea-level rise (6.3–6.9 mm and 4.1–10.6 mm SLE, respectively; Figures 7a–7d). In contrast, Petermann’s future decline under the Schoof and Weertman laws is not as pronounced, with moderate grounding-line retreat (10–15 km) and half the mass loss (2.7–3.5 mm SLE) of that under Budd by year 2300. These findings are consistent with previous studies on synthetic (Brondex et al., 2017) and Antarctic (Brondex et al., 2019) glaciers.

The response to ocean warming is rather subdued with the till-elevation friction, with a stable grounding line despite major ice-shelf loss, no dynamic thinning, and an acceleration of 30%–60% in the grounding zone-ice shelf region (Figures 4c and 4d).

Under till-assimilation (Figures 4e and 4f), Schoof (Figures 4i and 4j), and Weertman (Figures 4k and 4l) friction, grounding-line retreat is 3–5 km less than with Budd (Figures 7g and 7i) and ice-flow acceleration is less pronounced (50%–75%). Despite the modeled retreat and acceleration under the till-assimilation,
Figure 4. Future response of Petermann Glacier to ocean warming. Here, the geometry and velocity responses to a 2°C ocean warming are shown for every 10 years. (a–b) Budd law; (c–d) till-elevation law; (e–f) till-assimilation law; (g–h) till-assimilation-N-Budd law; (i–j) assimilated Schoof law, (k–l) assimilated Weertman friction law. Initial geometry after relaxation (Section 4.1) in upper respective panels are shown in dashed gray, observed velocities in respective lower panels are shown in black. Geometries and velocities are plotted along the dashed flowlines shown in Figure 5.
Schoof and Weertman laws, little inland propagation of dynamic thinning occurs (Figures 4e and 4i). This is in contrast to the large modeled upstream thinning that occurs under the Budd and till-assimilation-N-Budd friction laws.

Grounding-line retreat generally is episodic, asymmetric across-fjord, and paced by fjord bathymetry (Figures 4 and 5), consistent with findings for West Antarctic glaciers (Brondex et al., 2019; Seroussi et al., 2017). For a 2°C warming, mean grounding-line retreat rates are highest during the first 30 years (toward year 2050), with the highest retreat rates of 498 m/a found for Budd (Table 1). Grounding-line retreat rates decline gradually, and 80-year mean retreat rates toward year 2100 vary from 262 m/a for Budd to 51 m/a for the till-elevation law, with intermittent standstills at local bedrock highs (Figures 4 and 5). After 2100, grounding-line retreat continues only under the Budd and till-assimilation-N-Budd laws, while a new stable position has been established for the other laws. Retreat rates for all laws are listed in Table 1.

Figure 5. Grounding-line retreat in response to a +2°C ocean warming from present-day until year 2300. The observed front is shown with white dashed lines. (a) Budd friction law; (b) till-elevation law, (c) till-assimilation law, (d) till-assimilation-N-Budd law, (e) assimilated Schoof law, (f) assimilated Weertman law.
Grounding lines stabilize on the same bedrock ridges regardless of the imposed ocean temperature. Notably, the retreated grounding lines by 2300 vary by only a few km (∼5%–20% of total retreat length) between the 2°C and 5°C scenarios, for respective friction law. Local bedrock highs ∼10–15 km upstream of the present-day grounding line prevent further retreat for the Schoof, Weertman, and till-assimilation laws (Figures 7g and 7h), while with Budd and till-assimilation-N-Budd, retreat continues another 5–10 km upstream under the strongest warming (Figures 7g and 7j). Future ocean warming above 2°C has a bigger impact on ice-shelf frontal retreat than on grounding-line retreat, regardless of friction law, with additional 5–10 km calving front retreat in response to a warming exceeding 2 °C (Figures 7g–7l).

Sea-level rise contributions deviate by less than 10% across the 2°C–5°C scenarios for a given friction law, which is small considering the large temperature range. The exception is the till-assimilation-N-Budd law, where sea-level rise increases by 158% from a 2°C to a 5°C warming.

*Figure 6.* Modeled glacier geometry and grounding line (magenta) by 2300 under +2°C ocean warming. Shown is also the observed front (yellow line) and grounding zone (white lines). (a) Budd friction law; (b) till-elevation law; (c) till-assimilation law; (d) till-assimilation-N-Budd law; (e) assimilated Schoof law; (f) assimilated Weertman friction law.
Our ocean-warming simulations are sensitivity experiments rather than projections. In reality, ocean temperatures will increase gradually over time, and not change instantaneously. For context, we performed additional simulations where submarine melt rates were increased linearly over time, from today until 2300. This was done for all laws, and for the end-member temperatures 2°C and 5°C. A linear temperature increase delays ice-shelf breakup and grounding-line retreat by a few decades, where the ice shelf breaks up between 2050 and 2100, depending on the friction law (not shown here). Ultimately, the final grounding line positions and total sea-level rise by 2300 are still similar to those of the step-change experiments,

Figure 7. Future mass loss and contribution to sea-level rise from Petermann Glacier vary widely under different approaches to basal friction. Shown are responses to 0, 2, 3, 4 and 5°C ocean warming, for (a) Budd friction, (b) till-elevation law, (c) till-assimilation law, (d) till-assimilation-N-Budd law, (e) Schoof friction, (f) Weertman friction. Lower panels (g–l) show corresponding final grounding lines (solid lines) and calving fronts (dashed) by year 2300 for the six friction laws, for each temperature scenario. The observed grounding zone in 2017 is bracketed with black lines (ESA, 2017a) and the observed calving front is shown in white (ESA, 2017b).

Figure 8. Total thickness change from present-day to year 2300 under +2°C ocean warming. Positive (negative) values indicate ice thinning (thickening). Shown is also the modeled and observed fronts (dashed lines) and grounding lines (solid lines). (a) Budd friction law; (b) till-elevation law; (c) till-assimilation law; (d) till-assimilation-N-Budd law; (e) assimilated Schoof law, and (f) assimilated Weertman law.
although this differs depending on the temperature increase and the friction law. We found that the differences between a stepwise and linear temperature increase is substantial by 2100, and more moderate by 2300. For example, the most extreme 5°C warming (Budd law) lead to a sea-level contribution three times higher by 2100 using a step temperature change compared to a linear change, while the difference is only ~15% by 2300. For the more moderate 2°C scenario, the differences are larger; 3.6 times higher by 2100 (1.97 vs. 0.55 mm) for a linear temperature increase, while being 60% lower by 2300 (6.29 vs. 3.67 mm).

The no-warming control runs show that sea-level contributions by 2300 differ by 51% across the friction laws (Figures 7a–7f). Once ocean-warming is applied, the differences increase to 133% (2°C) and 282% (5°C), that is, the differences triple (2°C) and quintuple (5°C) compared to the respective control simulations (Figure 7). For the complementary simulations using a linearly increasing ocean temperature (not shown here), the numbers are smaller but remain substantial; differences across the friction laws by 2300 are 39% and 173% for a 2°C and 5°C ocean warming, respectively.

5. Discussion
5.1. Why the Friction Laws Disagree

Our findings for one of Greenland’s key outlet glaciers illustrate how equivocal current ice-sheet models are with respect to friction, and highlight the uncertainty in projections. While this is consistent with studies of synthetic and Antarctic glaciers (Brondex et al., 2017, 2019; Nias et al., 2018), the current study suggests a more general importance of friction that hold also for the different setting of Greenland’s outlet glaciers. The friction laws compared here are currently employed for future projections and studies of past ice sheets (e.g., Aschwanden et al., 2016; Golledge et al., 2019; Nias et al., 2018; Seguinot et al., 2016; Seroussi et al., 2017). For Petermann, we find clearly contrasting (a) grounding-line retreat, (b) ice acceleration, and (c) inland dynamic thinning (Section 4.3). The differing behavior is mainly a result of the inherent physics of each law and the underlying governing variables (like effective pressure), as well as how friction evolves under changing ice geometry and stress conditions. The transient differences are also influenced by slightly different initial conditions (see Sections 2.2.3 and 2.2.5). In practice, it is not possible to swap the initial conditions across the different friction laws for direct comparison, since each initial model state is internally consistent with the particular friction law. However, the till-elevation law clearly stands out from the rest since the grounding line advances by 10–15 km during present-day relaxation, associated with an upstream thickening of 100–200 m (Section 3.3; Figure S5c). This advanced, thicker initial state likely contributes to the subdued response to ocean warming for this friction law (Figure 4c).

For low effective pressure, as typically prevails close to the grounding line and under fast-flowing ice, the Schoof law tends to Coulomb-type flow (Section 2.2.5). This explains the similar grounding-line retreat under the Schoof and the Coulomb-type till-assimilation law (Figures 5c and 5e). Petermann’s strong response to the imposed ocean warming for Budd and till-assimilation-N-Budd compared to the other laws may be explained by how thinning translates into changes in basal stress conditions. Dynamic thinning brings the lower ice stream closer to flotation, which makes stabilization from local bedrock highs less likely and further grounding-line retreat more likely (Figures 4a, 7a and 5a). As ice thinning occurs, effective pressure drops and the basal shear stress decreases linearly under Budd (cf. Equation 1). This linear relationship does not prevail in the other laws, and for the Weertman law, no link between the effective pressure and basal shear stress exist whatsoever. Moreover, the basal stress under Budd depends on the basal velocity and effective pressure across the entire glacier, and the effective pressure may be low far inland, since it depends on the bed elevation (Section 2.2.1). This allows ice thinning to propagate further inland than with the other laws (Figures 4a and S8a), consistent with Brondex et al. (2019). In contrast, the effect of effective pressure changes (e.g., due to thinning) on the basal stress under Schoof depends on whether these changes occur at low- or high effective pressure (cf. Section 2.2.5, see also Brondex et al., 2017). Ice flow with the Schoof law is Coulomb-like under low effective pressure, conditions that prevail under the ice stream and close to the grounding line, and Weertman-like under high effective pressures (e.g., at bedrock highs).

As the Weertman law does not include any explicit dependence on the effective pressure, this law unsurprisingly shows the most moderate response and predicts the least sea-level rise, in line with previous studies (Brondex et al., 2017, 2019).
Conversely, the assimilated till-friction law with Budd’s effective pressure ("till-assimilation-N-Budd") displays the strongest response to the imposed future ocean forcing and contributes the most to sea-level rise for temperatures exceeding 2°C (Figure 7d), reaching ~8–11 mm SLE by 2300. The grounding line flux by 2300 with till-assimilation-N-Budd is also double that of till-assimilation (~25 Gt/yr vs. 12 Gt/yr), highlighting how important the formulation of effective pressure is (see also Section S1). The till-assimilation-N-Budd friction law responds instantaneously to changes in ice thickness and effective pressure approaches zero at the grounding line (because it is proportional to the height above flotation), whereas till-assimilation depends on basal water thickness, which changes slowly, and the ice overburden pressure remains high. This may explain why the till-assimilation-N-Budd law shows a stronger response than with the Coulomb effective pressure (cf. Equation 4).

With till-assimilation-N-Budd, we have (roughly) \( \tau_b \approx -\alpha \nu b^{0.6} \), which is very close to the Budd law, but we have an exponent \( q \) of 0.6 instead of 1. As the glacier accelerates, the increase in friction will be less pronounced for till-assimilation-N-Budd than for Budd, which explains the slightly higher speed up (Figures 4b and 4h), which in turn renders higher discharge and mass loss with till-assimilation-N-Budd than with Budd (Figures 7a and 7d).

### 5.2. Future Stability and Projections

Our findings show that Petermann’s ice shelf breaks up within decades following our imposed instant ocean warming, and indicate that we cannot rule out an associated grounding-line retreat of 25 km and almost 11 mm of sea-level rise by 2300 forced by the ocean alone (Figures 4 and 7). However, depending on basal friction, retreat may also be stopped by a bedrock ridge some 10–15 km upstream of the present-day grounding line (Figure 4d), rendering less than 3 mm sea-level rise for even a strong 5°C warming (Figure 7d). This bedrock ridge was found to provide a strong control on grounding-line stability also by Hill et al. (2021), who simulated the response of Petermann Glacier due to enhanced basal melt and ice-shelf breakup using a Weertman friction law and prescribed calving rates. They found a sea-level contribution of <1 mm even for their most extreme case involving full ice-shelf collapse and increased basal melt. The moderate sea-level contributions found in our Weertman-simulations are thus consistent with the subdued response found by Hill et al. (2021). However, the maximum melt rates applied by Hill et al. (2021) (50 m/a) were high-end estimates of present-day conditions (Wilson et al., 2017), and are 33% lower than in our 2°C warming experiments (75 m/a). Importantly, the Weertman law used by Hill et al. (2021) consistently lead to the lowest sea-level estimates of all friction laws tested here, supporting findings by Brondex et al. (2019) for West Antarctic glaciers, and should thus be considered minimum estimates.

We cannot, based on our simulations, discount any of these end-member scenarios. The underlying friction laws give results that are similarly close to present-day observations, and present-day conditions can anyway not be used exclusively to identify the most appropriate law. Our simulations suggest that outlet-glacier stability should be considered in the context of local basal conditions. Ice sheet-wide assumptions (e.g., Stearns & Van der Veen, 2018) are not expected to correctly resolve site-specific flow, and can only provide a picture where errors even out, at best.

Finally, our experiments show that the widely used (e.g., Aschwanden et al., 2016, 2019; Golledge et al., 2019), non-assimilated till-elevation friction law performs poorly for present-day Petermann (Figures 3c, 3d and S4–S6), and we recommend not using this law for short-term simulations and projections.

The Budd and Schoof laws show best fidelity and are the standout candidates for projections on short- and long-time scales, though they disagree significantly on future mass loss (Figures 7a, 7e, 8a, and 8e). Further work at other glacier and geologic conditions is needed to improve our confidence in applying either of these laws to future modeling, specifically comparing their performance at the grounding line and upstream areas.

While a validation of the friction laws is not the purpose of this study, testing them against multiple data sets over time could be a viable way forward. This could entail using changes in, for example, ice velocity, surface elevation, calving rates, front positions, and grounding lines over several decades. In reality, there are currently only a handful of glaciers in Greenland where such detailed transient data are available.
Our imposed “future” scenarios are ocean-only sensitivity experiments and not complete future projections. In our experiments, ocean temperatures are increased instantaneously, and we do not consider changes to surface melt, which accounts for 60% of Greenland’s recent mass loss (van den Broeke et al., 2016). Given the rapid reductions seen in the extent of Petermann’s ice shelf over the past decades, our aim is rather to study Petermann’s continued response to warming ocean waters, and to compare the responses across different friction laws. Surface melt is likely to raise Petermann’s sea-level contribution further and be increasingly important over the coming centuries (Aschwanden et al., 2019).

In the “future” experiments, ocean temperatures and associated melt rates are raised based on published model simulations (Cai et al., 2017), without any increase in subglacial discharge. With future surface warming, subglacial discharge is expected to increase as well (e.g., Slater et al., 2016) due to higher surface melt and runoff, leading to more vigorous submarine melt and even higher future mass loss.

6. Conclusion

Basal friction introduces major uncertainties in projections of ice-sheet mass loss and future sea-level rise. We compared six approaches to basal friction: a Budd law, a Schoof law, a Weertman law, and three variations of a Coulomb-type till-friction law. These laws are used in a three-dimensional higher order ice-sheet model, applied to contemporary Petermann Glacier and its future fate in response to a warming ocean.

We find diverse responses under our compared friction laws, but all suggest breakup of Petermann’s ice shelf within decades. Under our imposed ocean-only warming, grounding-line retreat differs by 20 km and the contribution to sea level by ~130% (2.7–6.3 mm SLE for 2°C) by year 2300. The Budd friction law predicts the largest mass loss for a 2°C ocean warming, and Schoof friction the second least, yet these are the two models that most closely represent Petermann’s modern behavior. The Weertman law consistently predicts the lowest contribution to sea-level rise.

A bedrock ridge 10 km upstream of the present-day grounding line stops retreat for four out of six friction laws, while retreat with the Budd and till-assimilation-N-Budd laws continues and amounts to nearly 7–11 mm of sea-level rise by 2300 in the strongest 5°C warming scenario.

Current-generation ice-sheet models employ a diversity of approaches to account for basal friction. Our study illustrates that these models disagree greatly on glacier behavior and response to ocean warming. In fact, our results show that choosing the “wrong” friction law may have a larger effect on future estimates of sea-level rise than an ocean warming of several degrees. This holds regardless of temperature scenario and whether the changes are imposed instantaneously or gradually. The approaches with highest fidelity for modeling modern Petermann Glacier provide diverse estimates for Petermann’s future retreat. This is due to the differing frictional response to changes in effective basal stress. Resolving which approach works best across a wider range of modern glaciers, ideally with data over several decades, appears as a way forward toward more robust future projections. More work on subglacial hydrology and its influence on effective pressure is also required, emphasized here by the diverse responses as a result of how the effective pressure is calculated.

Data Availability Statement

The ISSM code is freely available from the ISSM website (https://issm.jpl.nasa.gov/download). Model scripts used to prepare and launch simulations are publicly available at https://git.bolin.su.se/bolin/akesson-petermann. All underlying data used in this study are publicly available. A list with links to these are provided in the supplementary material.

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