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Modeling of ocean-induced ice melt rates of five west Greenland glaciers over the past two decades

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## Key Points:

- How does the ocean control the melting of Greenland glaciers?
- Glaciers dominated by calving processes are less sensitive to ice-ocean interactions
- The study provided a basis for quantifying ice-ocean interaction elsewhere around Greenland

## Supporting Information:

- Supporting Information S1

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## Modeling of ocean-induced ice melt rates of five west Greenland glaciers over the past two decades

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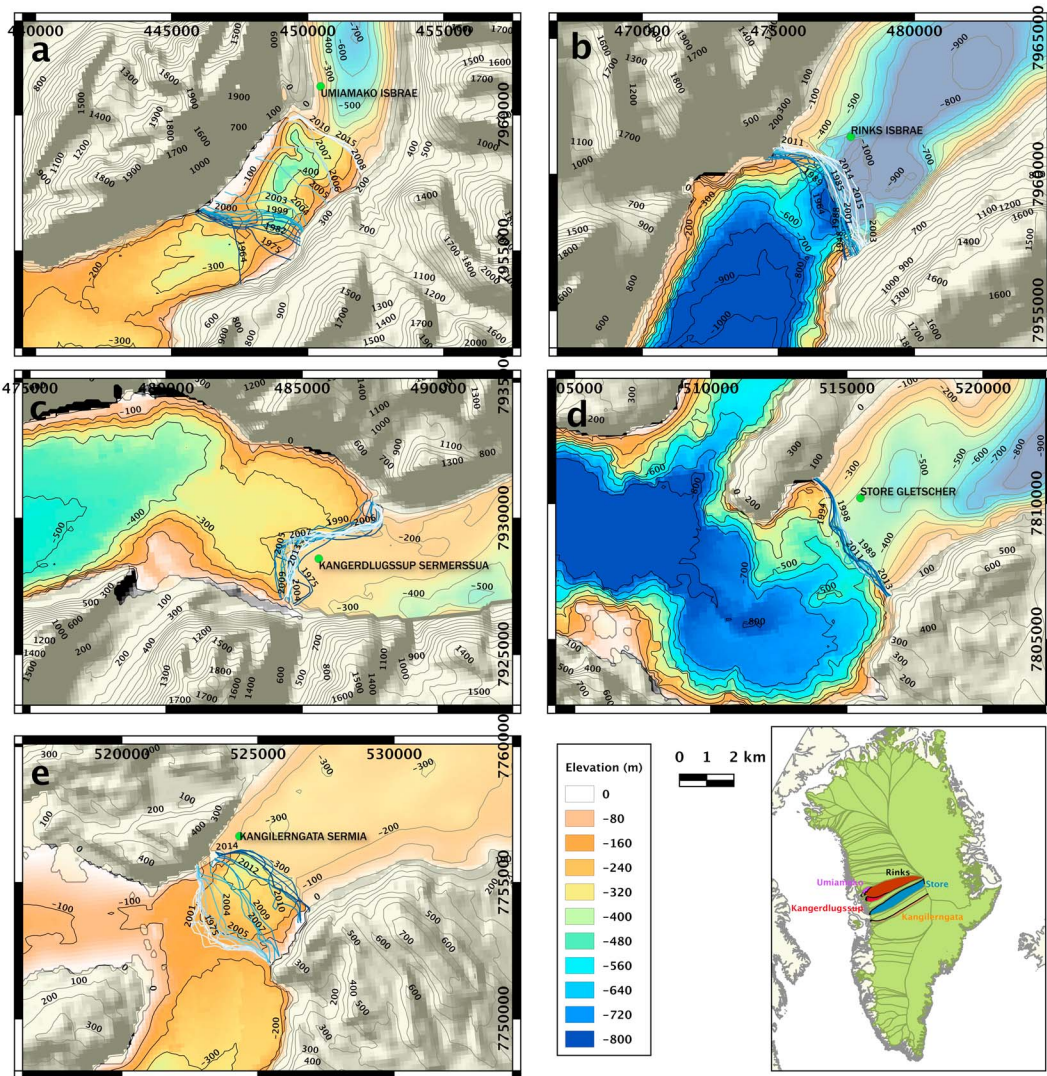
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**Abstract** High-resolution, three-dimensional simulations from the Massachusetts Institute of Technology general circulation model ocean model are used to calculate the subaqueous melt rate of the calving faces of Umiamakko, Rinks, Kangerdlugssup, Store, and Kangilerngata glaciers, west Greenland, from 1992 to 2015. Model forcing is from monthly reconstructions of ocean state and ice sheet runoff. Results are analyzed in combination with observations of bathymetry, bed elevation, ice front retreat, and glacier speed. We calculate that subaqueous melt rates are 2–3 times larger in summer compared to winter and doubled in magnitude since the 1990s due to enhanced subglacial runoff and  $1.6 \pm 0.3^\circ\text{C}$  warmer ocean temperature. Umiamakko and Kangilerngata retreated rapidly in the 2000s when subaqueous melt rates exceeded the calving rates and ice front retreated to deeper bed elevation. In contrast, Store, Kangerdlugssup, and Rinks have remained stable because their subaqueous melt rates are 3–4 times lower than their calving rates, i.e., the glaciers are dominated by calving processes.

### 1. Introduction

The Greenland Ice Sheet has been experiencing an accelerating rate of mass loss over the past few decades from a combination of enhanced surface melt and increased glacier ice discharge [e.g., Rignot and Kanagaratnam, 2006; van den Broeke et al., 2009; Enderlin et al., 2014]. Widespread acceleration of its marine-terminating glaciers and fluctuations in the position of their calving margins has been attributed to the enhanced advection of warm subsurface ocean waters of subtropical origin at the ice-ocean boundary [e.g., Holland et al., 2008; Rignot et al., 2010; Straneo et al., 2010; Rignot et al., 2012; Bevan et al., 2012; Christoffersen et al., 2011]. Yet there has been no quantitative demonstration of the role of subaqueous melt in driving tide-water glaciers into a retreat, in part due to our poor knowledge of the magnitude of the subaqueous melt rates, their spatial pattern, horizontally and vertically, and their seasonal to interannual variability. One hypothesis is that subaqueous melting at the ice front has increased sufficiently in magnitude to counteract the advection of ice from upstream and force the glacier front to retreat. As the glacier retreats, the removal of grounded ice at the ice front will reduce basal resistance, which will enable glacier ice from upstream to speedup. Speedup will cause ice to thin from longitudinal stretching, which will be conducive to further retreat. The retreat rate will depend on the glacier geometry, especially the bed elevation below sea level and whether the bed elevation increases or decreases farther inland [e.g., Post et al., 2011]. When examining the details of the glacier response to ocean thermal forcing, large variations have been reported from one glacier to the next or within the same fjord system [e.g., Box and Decker, 2011; Moon et al., 2012], which suggests a complex interplay between subaqueous melting, iceberg calving, glacier geometry, and other factors [Benn et al., 2007].

Here we consider the recent history of five glaciers in central west Greenland for which we have quality data to reconstruct ocean-induced melt rates. We review their time evolution and response to atmospheric and oceanic forcings. We use three-dimensional, high-resolution simulations of ice melt into the ocean to model their horizontal, vertical, seasonal, and interannual variability. Model forcing is constrained using in situ oceanographic data, reconstructions of ice sheet runoff, novel bed topography, and bathymetry data. We compare changes in subaqueous melt rate through time with ice velocity at the ice front. We conclude on the impact of ice-ocean interaction on the glacier dynamics and in turn on glacier stability.



**Figure 1.** Bed elevation of five Greenland glaciers and bathymetry in the glacial fjords of (a) Umiamako Isbrae, (b) Rinks Isbrae, (c) Kangerdlugssup Sermerssua, (d) Store Gletscher, and (e) Kangilerngata Sermia. Inset shows the location of the glacier basins in Greenland color coded purple, brown, red, blue, and orange from north to south. Bed elevation and bathymetry contours are every 100 m. Bed elevation on grounded ice is shaded with 52% transparency. Ice front positions from 1964 to 2015 are color coded from blue to white, with black labels.

## 2. Data and Methods

### 2.1. Remote Sensing Data

Store Gletscher, Kangerdlugssup Sermerssua, Rinks Isbrae, and Umiamako Isbrae are four tidewater glaciers from the Umannaak region, in west Greenland (Figure 1a); Kangilerngata Sermia drains into Ata Sund fjord, 50 km to the south (Figure 1). The drainage basin, ice flux, and annual speed of the glaciers are listed in Table 1. Store and Rinks are among the fastest glaciers in Greenland, with large drainage basins; the other two glaciers flow 2–3 times slower, with smaller drainage basins [Rignot and Mouginot, 2012].

To conduct our analysis of the glacier evolution, we employ data on ice front position, bed elevation, bathymetry, ice velocity, ocean temperature, and subglacial freshwater discharge from the glaciers. We digitize the ice front positions using airborne imagery from 1964 [Carbournell and Bauer, 1968] and satellite imagery from Landsat-MSS (1976), Landsat-TM (1975–1978), Landsat-5 (1985–1998), Landsat-7 (1999–2012), and Landsat-8 (2013–2015) (Figure 1). The data are from midsummer when the ice fronts are easy to identify but not necessarily at their most retreated position [Howat et al., 2010]. Yearly ice front positions are mapped within 30 m. Short-term variations average about 100 m. We calculate an annual retreat rate,  $q_r$ , in meter per

**Table 1.** Characteristics of Five Greenland Tidewater Glaciers<sup>a</sup>

Glacier	DB km <sup>2</sup>	$Q_f$	$A_f$	$Q_{sg}$			$q_m$		$q_f$	$q_r$	
		Gt/yr Yr	km <sup>2</sup> Yr	W	Yr	Peak	W	S	Peak	Yr	Yr
Umiamako Isbrae	2,304	2.6	0.9	1	16	198	0.6	1.0	2.0	1.3 ± 0.4	0.5 ± 0.8
Rinks Isbrae	33,171	12.7	2.7	13	38	405	1.1	1.6	3.4	8.3 ± 0.5	0.05 ± 0.5
Kangerdlugssup S.	5,000	3.0	1.4	3	39	473	0.6	1.0	2.0	4.5 ± 0.5	0.03 ± 0.3
Store Gletscher	30,507	11.5	1.8	13	99	890	0.8	1.5	3.0	5.2 ± 0.4	−0.02 ± 0.2
Kangilerngata S.	8,004	1.2	0.9	3	57	438	0.4	0.8	1.9	2.0 ± 0.9	0.4 ± 0.8

<sup>a</sup>DB is the drainage basin in square km;  $Q_f$  the ice front flux in 2007 in gigatons ( $10^{12}$ kg) per year;  $A_f$  the mean submerged area in square kilometer;  $Q_{sg}$  the mean subglacial water flux in cubic meter per second in winter (W), annual (Yr), and summer peak (peak); and  $q_m$  the mean subaqueous melt rate in meter per day in winter (W), summer (S), and summer peak value (peak).  $q_f$  and  $q_r$  are, respectively, the mean annual speed and annual retreat rate ( $\pm 1\sigma$ ) for the time period 1992 to 2015.

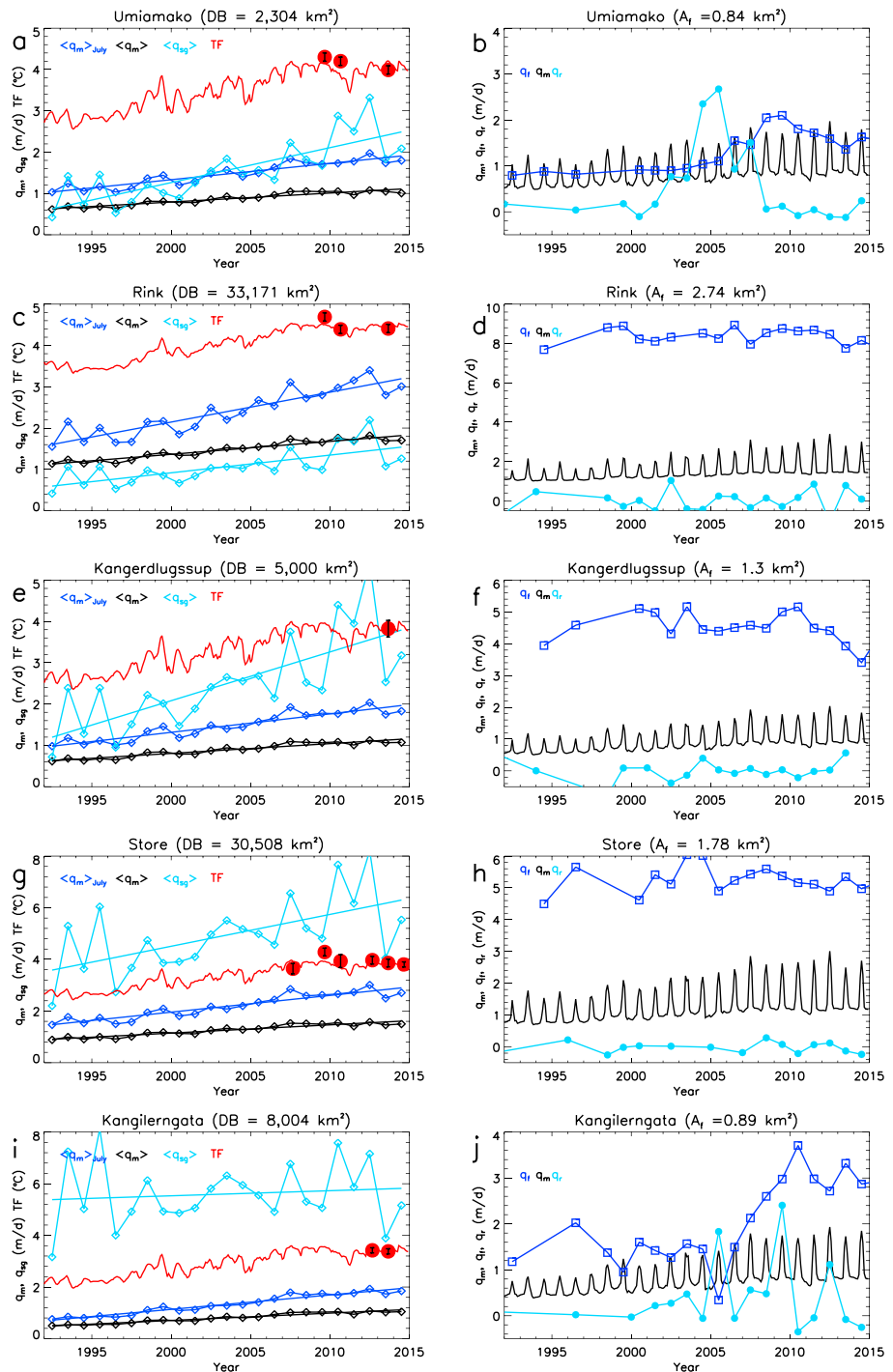
day, by dividing the area of retreat by the glacier width over the 1 year period separating the images. Data separated by a few months or less yield noisy time series of  $q_r$ , because iceberg calving is not a continuous process like ice motion or ice melt by the ocean.

Glacier bed elevation is from *Morlighem et al.* [2014] combining NASA Operation IceBridge (OIB) thickness data and ice motion vectors from *Rignot and Mouginot* [2012] using a mass conservation (MC) approach that minimizes a cost function that measures the disparity between observed and calculated ice thickness. Bed elevation and ice thickness have a horizontal resolution of 350 m, a vertical precision of 50 m, and a posting of 150 m. The MC bed elevation output products are constrained to match the bathymetry data at the ice-ocean margin by including an additional term in the cost function that minimizes the difference between observed bathymetry at the ice front and calculated bed depth at the ice front. Bathymetry was collected using a Reson 8160 multibeam echo sounder operating at 50 kHz in August 2012 and 2013 (Figure 1), with a 1 m vertical precision, at a 25 m posting [*Rignot et al.*, 2016]. For Umiamako, we extended the MC results using a bed solution for year 1996 when the glacier was at a more advanced position. Combined together, these data help constrain water depth,  $h$ .

We use ice velocity data for years 2000 to 2011 from *Joughin et al.* [2010], complemented with ice velocity for year 1989–2015 from Landsat data (*Howat* [2016] and this study), year 1996 from ERS-1 interferometry, and winter 2014 from RADARSAT-2. We calculate a width-averaged ice speed for the glacier fronts,  $q_f$ , in meter per day. The precision of  $q_f$  is 1 m/yr for 2000–2015, 2–3 m/yr for 1996, and 15 m/yr for Landsat data prior to 2013. We generate annual averages in ice velocity to minimize data noise.

## 2.2. ECCO Ocean Thermal Forcing

The amount of ocean heat delivered to the glacier to melt ice is proportional to the difference in temperature between the potential temperature of seawater and the pressure-dependent freezing point of seawater or ocean thermal forcing,  $TF$ . We calculate a vertically averaged TF between 200 m depth and the depth of the seafloor in front of the glacier. Ocean model simulations show that it is approximately over that depth range ( $\pm 50$  m) than ice melts and melt water mixes with ocean waters [*Xu et al.*, 2013]. Ocean temperature is from an ocean simulation carried out by the Estimating the Circulation and Climate of the Ocean, Phase II (ECCO2) project at 4 km horizontal spacing [*Nguyen et al.*, 2012] based on the optimized coarser-resolution configuration of *Losch et al.* [2010] and *Nguyen et al.* [2011]. Due to the coarse resolution of the ECCO2 simulation and the uncertainty of the bathymetry data used by ECCO2, we extract a mean ocean temperature profile for the entire Uummannaq fjord. A comparison of ECCO2 model results with long-term temperature data from the Labrador and Irminger Seas was reported in *Rignot et al.* [2012] and showed a  $+0.4^\circ\text{C}$  bias. Here we use conductivity-temperature-depth (CTD) data collected in years 2007 [*Weinrebe et al.*, 2009], 2008, 2009–2010 [*Chauche et al.*, 2014], 2011, 2012, 2013, and 2014 to compare ECCO2 temperature and observations (Figure 2). We adjust the ECCO2 model results using the CTD data from year 2013 only. The adjustment is  $0.1^\circ\text{C}$  for Umiamako, Rink, and Kangerdlugssup;  $0.2^\circ\text{C}$  for Store; and  $0.7^\circ\text{C}$  for Kangilernata. We calculate a mean difference in TF between ECCO2 and all CTD data of  $0.2^\circ\text{C}$ . If we vary the minimum depth of TF by  $\pm 50$  m, we find



**Figure 2.** Evolution of the subaqueous melt rate  $q_m$  of five Greenland glaciers: (a, b) Umiamako Isbrae, (c, d) Rinks Isbrae, (e, f) Kangerdlugssup Sermerssuam, (g, h) Store Gletscher, and (i, j) Kangilerngata Sermia. Figures 2a, 2c, 2e, and 2g show the annual effective subglacial water discharge (light blue diamond),  $\langle q_{sg} \rangle$ , the summer (July) subaqueous melt rate (dark blue diamond),  $\langle q_m \rangle_{July}$ , the annual subaqueous melt rate (black diamond),  $\langle q_m \rangle$ , and their corresponding linear fit from 1992 to 2015, in meter per day. Ocean thermal forcing, TF, in degree Celsius is red. Solid red circles denote CTD temperature data from year 2007 to 2014. DB is the drainage basin area in square kilometers. Figures 2b, 2d, 2f, and 2h show the monthly ice speeds (dark blue square with thin line in between measurements),  $q_f$ , the monthly subaqueous melt rate (black curve),  $q_m$ , and the yearly glacier retreat rate (light blue circle with thin line in between measurements),  $q_r$ , from 1992 to 2015, in meter per day.  $A_f$  is the ice front submerged area in square kilometers.



an uncertainty of 0.1°C in the CTD values for Umiamak, Rink, and Kangilernata and 0.2°C for Kangerdlugssup and Store. In total, we estimate a mean uncertainty in TF of 0.3°C.

The thermohaline circulation that takes place in front of the marine-terminating glaciers depends on the subglacial freshwater flux,  $Q_{sg}$ , delivered at the glacier base.  $Q_{sg}$  includes (1) ice sheet surface runoff and (2) subglacial melt from geothermal heat and basal friction. Ice sheet runoff is from the Regional Atmospheric Climate Model (RACMO2), with a precision of 20% [Ettema *et al.*, 2009; van Angelen *et al.*, 2012]. In summer, this is the dominant source of subglacial water. In winter, melting of grounded ice dominates. We calculate it using a numerical model of the thermal state of the Greenland Ice Sheet that employs the enthalpy method [Aschwanden *et al.*, 2012]. The model setup is identical to that in Seroussi *et al.* [2013], updated after evaluating the enthalpy framework following Kleiner *et al.* [2015]. The ice sheet runoff and grounded ice basal melt water production are integrated over the glacier surface drainage, i.e., assuming that the subglacial water drainage coincides with the surface drainage, to obtain basin-wide estimates of  $Q_{sg}$  in cubic meter per second. We deduce an effective subglacial freshwater velocity,  $q_{sg}$ , in meter per day, by dividing  $Q_{sg}$  by the submerged area of the ice front, derived from the bathymetry mapping, hence is time dependent, and updated annually (Table 1).  $q_{sg}$  is expressed in the same unit as  $q_f$  and  $q_m$  and differs from the unknown speed of subglacial flow within the channelized system at the ice front.

### 2.3. Ocean Model

We model subaqueous melting of a 150 m wide, vertical, ice face standing in ocean water with subglacial freshwater discharge using the Massachusetts Institute of Technology general circulation model (MITgcm), with a free surface, nonhydrostatic conditions, in three dimensions and at a 1 m spatial resolution sufficient to generate plume turbulence [Xu *et al.*, 2013]. The fjord is rectangular in shape and extends 500 m from the ice front (supporting information). The subglacial discharge is injected at the glacier base via a fixed-size cavity and imposed as a net influx of cold, fresh water mass. Simulations indicate that the exact shape and distribution of subglacial cavities only modulate the area-average melt rate by  $\pm 15\%$  keeping the subglacial water flux constant [Xu *et al.*, 2013]. Slater *et al.* [2015] report a larger variability in melt rates for the more extreme cases of fully channelized and fully distributed systems. Most subglacial water is, however, discharged from 1–2 main channels and a series of smaller channels [Rignot *et al.*, 2015; Fried *et al.*, 2015], which is a configuration similar to Xu *et al.* [2013]. Here we assume an uncertainty of 15% but recognize that the uncertainty could be larger.

From the MITgcm simulations, we calculate the effective ocean-induced melt speed,  $q_m$ , in meter per day, as the horizontally averaged, maximum rate of ice melt.  $q_m$  is not the area-averaged ice melt rate as in Xu *et al.* [2013] because the melt rate is not uniform with depth. The melt rate reaches a maximum value in the lower part of the ice face, hence creating conditions for ice front undercutting [Rignot *et al.*, 2015; Luckman *et al.*, 2015; O'Leary and Christoffersen, 2013]. It is the rate of deep undercutting caused by the ocean waters that controls the migration back and forth of the ice front, not the area-average melt rate or melt rate near the surface. Undercutting has been observed using multibeam echo soundings of calving faces [Rignot *et al.*, 2015; Fried *et al.*, 2015] and confirmed to be a significant driver of iceberg calving for tidewater glaciers [Luckman *et al.*, 2015]. Ice below the region of undercutting will exert negligible basal friction because it is too thin (10–100 m). We calculate  $q_m$  for various values of TF,  $q_{sg}$ , and water depth,  $h$ , and use a simple model fit to match the results (Table S1 in the supporting information). We then find a formulation that unifies all the simulations together using a generalization of the model fit used by Xu *et al.* [2013] as

$$q_m = (A h q_{sg}^\alpha + B) TF^\beta \quad (1)$$

with  $A = 3 \times 10^{-4}$  units,  $h$  is water depth in meter,  $q_{sg}$  the subglacial water flux in m/d,  $\alpha = 0.39$ ,  $B = 0.15$  units, TF is the thermal forcing, and  $\beta = 1.18$  (Figure S1).  $B$  expresses that  $q_m$  is nonzero in the absence of subglacial water flux. TF depends on water depth via the pressure-dependent freezing point of seawater. We find that the model fit agrees with the simulations within 0.2 m/d or 7% (Figure S2).

The uncertainty in  $q_m$  includes a 20% uncertainty in  $q_{sg}$  (scaled by  $\alpha$  in equation (1)) and a 0.3°C uncertainty in TF or 5% (scaled by  $\beta$ ), which amounts to a 14% uncertainty from equation (1). We combine this error quadratically with a 15% error due to the unknown shape and spatial distribution of subglacial channels [Xu *et al.*, 2013], a 5% error in submerged cross-section submerged ice front area, and a 7% uncertainty due to the model fit (Figure S2). We estimate the total uncertainty in  $q_m$  to be 26%.

Given  $q_{sg}$ , TF, and  $h$ , we calculate monthly values of  $q_m$  using equation (1) taking into account changes in position of the ice front and water depth. We then compare the effective melt speed,  $q_m$ , with the average glacier ice front speed,  $q_f$ , and the annual rate of ice front retreat,  $q_r$ , which balance the calving speed,  $q_c$ , as

$$q_r = q_f - q_c - q_m \quad (2)$$

The calving speed,  $q_c$ , here represents the dry rate of iceberg calving that would take place in the absence of ocean melt. Calving by undercutting is included in  $q_m$  since it is the maximum melt rate at depth. If calving and ocean-induced melt balance  $q_f$  exactly, then the glacier front position is steady, i.e.,  $q_r = 0$ . If  $q_m$  increases, for instance, with warmer ocean waters or higher subglacial runoff, the glacier front will retreat ( $q_r < 0$ ) unless  $q_f$  increases.

### 3. Results

We describe the glacier geometry, retreat rate, and subglacial runoff; calculated ocean-induced melt rate; and compare the results of the glacier speed. Umiamakko Isbrae flows down a 4 km wide valley with a 200 m deep sill at the mouth of the fjord where the glacier front was anchored as far back as the 1850s [Weidick, 1968] (Figure 1a). The fjord is up to 300 m deep at the 1964 ice front and up to 350 m deep at the 2012 ice front. Another 5 km upstream, the bed drops to 680 m below sea level. The ice front that retreated 1.5 km from 1964 to 1985 along its southern flank was stable until 2002, retreated by 3.7 km from 2002 to 2008, and has remained stable since.

Rinks Isbrae flows down a 6 km wide valley with steep walls (Figure 1b). The fjord is more than 1000 m deep, with a 650 m deep sill near the ice front. The central part of the glacier is near floatation but is treated here as a near-vertical calving face equivalent. The ice front migrated back and forth with large calving events but remains within 1 km of the 1964 position. Kangerdlugssup is 5 km wide and terminates in waters 350 m deep (Figure 1c). About 6 km inland of the ice front, bed elevation drops by 250 m. The glacier front did not migrate at a significant level between 1964 and 2015. Store is grounded on a 500 m deep sill, several kilometers in length by 5 km in width (Figure 1d). The glacier front has been stable since 1964 and even 1850 [Weidick, 1968].

Kangilerngata Sermia is grounded 350 m below sea level in a 3.5 km wide valley (Figure 1e). We detect a sill at 150 m depth about 3 km from the current ice front, where the glacier was grounded from 1964 to 2004 and even in 1920 [Weidick, 1968]. The glacier front retreated by 3 km in 2005–2010 and has stabilized since. The bed elevation remains 200–450 m below sea level for 30 km inland of the current ice front position.

Subglacial runoff has increased on all glaciers during the observation period, almost linearly, with large inter-annual variations (Figure 2). Runoff flux peaked in 2012 but returned to normal in 2013–2014. Winter (no runoff) subglacial flux  $Q_{sg}$  is 1, 13, 3, 13, and 3  $\text{m}^3/\text{s}$  for, respectively, Umiamakko, Rinks, Kangerdlugssup, Store, and Kangilerngata (Table 1). Summer (month of July)  $Q_{sg}$  is 1–2 orders of magnitude larger. At Umiamakko,  $Q_{sg}$  peaks at 198  $\text{m}^3/\text{s}$  in summer and averages 16  $\text{m}^3/\text{s}$  for the year. The yearly and peak values for Rinks are 38 and 405  $\text{m}^3/\text{s}$ , 39 and 473  $\text{m}^3/\text{s}$  for Kangerdlugssup, 99 and 890  $\text{m}^3/\text{s}$  for Store, and 57 and 438  $\text{m}^3/\text{s}$  for Kangilerngata. Subglacial water flux varies from year to year but increased by 0.7  $\text{m}^3/\text{s}/\text{yr}$  for Umiamakko or 96%, 1.4  $\text{m}^3/\text{s}/\text{yr}$  for Rinks or 81%, 1.7  $\text{m}^3/\text{s}/\text{yr}$  for Kangerdlugssup or 96%, 2.6  $\text{m}^3/\text{s}/\text{yr}$  for Store or 58%, and 1.3  $\text{m}^3/\text{s}/\text{yr}$  for Kangilerngata or 50% in 22 years. During the same period, ocean thermal forcing deduced from ECCO2 increased by 1.5°C from 2.6 to 4.1°C at Store, 1.6°C from 2.1 to 3.7°C at Kangilerngata, 2.7 to 4.3°C at Umiamakko, 2.5 to 4.1°C at Kangerdlugssup, and 1.3°C from 3.4 to 4.7°C at Rinks (Figure 2).

As a result of the increase in  $q_{sg}$  and TF, we find that  $q_m$  increased significantly during the 1992–2014 time period (Figure 2). The melt rate increased from 0.6 to 1.0 m/d in winter due to the increase in TF for Umiamakko. The summer melt rate increased from 0.9 to 2.0 m/d. For Rinks, the rate increased from 1.0 to 1.5 m/d in winter and 1.4 to 3.3 m/d in summer. For Kangerdlugssup, the rate increased from 0.6 to 0.9 m/d in winter and 1.0 to 1.9 m/d in summer. For Store, the winter rate increased from 0.8 to 1.2 m/d and the summer rate from 1.5 to 2.8 m/d. For Kangilerngata, the rate increased from 0.4 to 0.9 m/d in winter and 0.7 to 1.9 m/d in summer. The annual melt rate doubled for Kangilerngata from 0.5 to 1 m/d as the glacier retreated into deeper waters and doubled for the other glaciers: 0.6 to 1.1 m/d for Umiamakko, 1.1 to 1.8 m/d for Rinks, 0.7 to 1.1 m/d for Kangerdlugssup, and 0.9 to 1.6 m/d for Store.

At Kangilerngata, the ice front speed,  $q_f$ , which averaged  $2.0 \pm 0.8$  m/d in 1992–2014, doubled from 1.5 m/d in the 1990s to 2.6 m/d in the last 5 years. At Rinks, Kangerdlugssup, and Store,  $q_f$  remained at, respectively,

$8.3 \pm 0.5$ ,  $4.5 \pm 0.5$ , and  $5.2 \pm 0.3$  m/d, with no temporal trend. At Umiamakko,  $q_f$  averaged  $1.1 \pm 0.4$  m/d in 1992–2014 but increased from 0.8 m/d in the 1990s to 1.8 m/d in 2009 before dropping back to 1.1 m/d in 2013–2014.

Comparing  $q_m$  and  $q_f$ , we find that  $q_f$  for Rinks is 5 times larger than the summer rate  $q_m$ . The increase in  $q_m$  brought it to within 40% of  $q_f$ , but  $q_m$  never exceeded  $q_f$ . For Store and Kangerdlugssup,  $q_f$  is 3 times larger than the summer  $q_m$ . For these three glaciers, the changes in frontal ablation are clearly dominated by calving processes, not by ocean-induced ice melt. In contrast, the summer  $q_m$  for Kangilerngata exceed  $q_f$  around 2002 until 2010. For Umiamakko,  $q_m$  was comparable to  $q_f$  for several years starting in the early 2000s. For these glaciers, changes in frontal ablation are dominated by submarine melting.

The ice fronts of Rinks, Kangerdlugssup, and Store did not migrate more than  $\pm 230$  m since 1992,  $q_f$  did not change, and the increase in  $q_m$  had no impact on the glacier stability. In contrast, Umiamakko and Kangilerngata retreated by 4.2 and 3.3 km, respectively, in 1992–2015. Most of the retreat of Umiamakko took place between 2002 and 2008, when the glacier retreated into a bed about 50 m deeper below sea level. Kangilerngata retreated 2.3 km between 2005 and 2010, from a bed 150 m below sea level to a bed 350 m below sea level.

#### 4. Discussion

The model simulations suggest significant fluctuations in melt rate between summer and winter. This is due to the net decrease in subglacial water flux in winter, which reduces the strength of the thermohaline circulation along the ice face. Winter melt rates are 2 times lower than the summer melt rates. Ice front retreat is therefore more likely to occur in the summer months when the melt rates are the highest, especially since ice front speeds only increase by 8–10% in summer compared to winter [Rignot and Kanagaratnam, 2006]. This implies that subglacial runoff, and therefore higher air temperature, exerts an indirect important control on the rate of subaqueous melt. On the other hand, the relatively high values modeled in winter indicate that ice melt by the ocean is a year-round phenomena, i.e., winter rates should not be neglected.

Few observations of ice melt rate have been obtained in Greenland [Rignot *et al.*, 2010; Motyka *et al.*, 2011; Sutherland and Straneo, 2012; Xu *et al.*, 2013; Enderlin and Howat, 2013; Fried *et al.*, 2015]. Our calculated values are within the range of those obtained in the field in Greenland over a period of a few days, i.e., 0.5 m/d to 4 m/d. Fried *et al.* [2015] quote an average melt rate of 1.6 m/d in July 2013 for Kangerdlugssup versus  $1.8 \pm 0.5$  m/d in our model. For Store, our summer results of  $3.0 \pm 0.8$  m/d for July 2012 are consistent with the hydrographic estimate of  $3.0 \pm 1$  m/d by Xu *et al.* [2013]. For Rinks, Enderlin and Howat [2013] estimate summer melt rates of 2.1 to 2.5 m/d in 2000–2008 from remote sensing versus  $3.4 \pm 0.8$  m/d for July and 1.8 m/d for June–September in our simulation. For Kangilerngata, Rignot *et al.* [2010] estimated  $2.6 \pm 0.5$  m/d in July and August 2008 versus  $1.9 \pm 0.4$  m/d herein.

The model simulations indicate that  $q_m$  has increased significantly over the past decades as a result of an increase in  $q_{sg}$ , TF, and water depth,  $h$ . The change in submerged area had only a small contribution to the change in  $q_m$ . The change in water depth enhanced  $q_m$  by 50% for Kangilerngata and 10% for Umiamakko.

In response to this forcing, Store, Rinks, and Kangerdlugssup have remained stable in ice front position, ice speed, and calving speed. We attribute the lack of retreat to the fact that  $q_m$  remains about 2–3 times lower than  $q_f$  for these glaciers, i.e., calving is the dominant ablation process. For an ice front to retreat ( $q_r < 0$ ), the summer melt rate must at least exceed  $q_f$ , which was not the case during the entire period. In contrast, for the other two glaciers, the model simulation clearly suggests that summer melt rates regularly exceeded the ice speeds. For Kangilerngata, the melt speed exceeded the ice front speed around 2003 and the glacier retreat started in 2005. The glacier retreated in deeper waters, which increased the melt rate. The ice speed nearly tripled during that time and is now twice as large as the ocean-induced melt rate. For Umiamakko, the summer melt rate exceeded the ice speed for several years, but the melt rate is only 1.5 m/d, so that the glacier could only retreat by a cumulative 50–100 m each summer. Its fast retreat started in 2002 as the winter melt matched the ice speed, i.e., enhanced ice melt retreated the ice front year round. The retreat peaked in 2005, at which point the glacier speed doubled, and the glacier has maintained a steady flow and ice front position in waters 50 m deeper.

These simulations indicate that glaciers dominated by calving processes are less sensitive to an increase in ocean temperature or subglacial runoff. For the other two glaciers, ice-ocean interaction dominates and the



glaciers had to retreat, as observed. Hence, the simulations explain why some glaciers retreated while others did not. For the glaciers that retreated, it provides a reasonable estimate of the timing of the retreat, i.e., when the annual melt rates increased above the ice speed. To extend the comparison further, we would need to model the time-dependent ice flow and calving dynamics of the glacier as in *Morlighem et al.* [2016].

If our calculations are correct, we can estimate how much the melt rates need to increase for Rinks, Store, and Kangerdlugssup to retreat. On Store, an increase of 50% would be necessary for  $q_m$  to approach  $q_f$ , whereas on Rinks and Kangerdlugssup  $q_m$  it would need to double. A doubling of  $q_m$  would require a substantial increase in subglacial water discharge and ocean temperature. We conclude that Rinks and Kangerdlugssup are the most stable glaciers among those studied here.

Our methodology is affected by a number of limitations beyond the uncertainty in TF,  $q_{sg}$ , and  $q_m$ . First is the issue of undercutting. The enhanced bending stress caused by undercutting decreases with time as ice viscously adjusts to bending [Cook *et al.*, 2014; Krug *et al.*, 2014], so the impact of undercutting on iceberg calving will depend on the time separation between calving events and in particular should be less for glaciers that do not calve frequently. Second, the ice face is not vertical, e.g., Rinks develops a small floating section. A more horizontal ice face will yield lower melt rates under similar forcings. Third, we have few measurements to evaluate the model, especially in winter where the predicted rates are low yet significant. Fourth, surface thinning may also contribute to glacier retreat but is neglected here because the melt speed from the surface is 1 order of magnitude lower than that at the ice-ocean boundary.

Our approach for representing ice-ocean interaction at ice margins using equation (2) is applicable to other glaciers using reconstructions of ice sheet runoff and ocean temperature from RACMO2, ECCO2, or other models. This makes it possible to quantify the impact of these interactions over a broader range of glaciers and test other atmospheric and ocean models in Greenland. To do this, it will be necessary to collect bathymetry and ocean temperature data in other fjords. In situ observations of melt rates in both summer and winter are critically needed to evaluate and improve these models.

## 5. Conclusions

We present a detailed modeling of subaqueous melt rates of five major Greenland glaciers using reconstructions of subglacial freshwater flux, ocean thermal forcing, bathymetry, and bed geometry for the past two decades. The results reveal a significant seasonal variability in subaqueous melt rate in response to the seasonal variation in subglacial runoff and a significant increase in subaqueous melting in response to the increase in ice sheet runoff and ocean temperature. For three glaciers, this increase has not changed the balance between the calving rate and the rate of ocean-induced melt, so the glaciers have remained stable despite an increase in ice melt. For two others, the calculation unequivocally indicates that the melting speed was sufficient to dislodge the glaciers from their stable positions and trigger a glacier retreat which affected their flow speed and state of mass balance, especially when the glacier fronts retreated to deeper beds. We suggest that the recent and future evolution of other Greenland glaciers may be simulated in the same fashion, using simplified reconstructions of ocean-induced melt rates, and extended in the future with improved modelizations of ice-ocean interaction.

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