Continued retreat of Thwaites Glacier, West Antarctica, controlled by bed topography and ocean circulation

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The Amundsen Sea sector is experiencing the largest mass loss, glacier acceleration, and grounding line retreat in Antarctica. Enhanced intrusion of Circumpolar Deep Water onto the continental shelf has been proposed as the primary forcing mechanism for the retreat. Here, we investigate the dynamics and evolution of Thwaites Glacier with a novel, fully-coupled, ice-ocean numerical model. We obtain a significantly improved agreement with the observed pattern of glacial retreat using the coupled model. Coupled simulations over the coming decades indicate a continued mass loss at a sustained rate. Uncoupled simulations using a depth-dependent parameterization of sub-ice-shelf melt significantly overestimates the rate of grounding line retreat compared to the coupled model as the parameterization does not capture the complexity of the ocean circulation associated with the formation of confined cavities during the retreat. Bed topography controls the pattern of grounding line retreat, while oceanic thermal forcing impacts the rate of grounding line retreat. The importance of oceanic forcing increases with time as Thwaites grounding line retreats farther inland.

**Keypoints:**

- Thwaites glacier likely to continue losing mass at a similar rate over the next decades
- Grounding line retreat governed by bed topography but retreat rate influenced by oceanic conditions

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Coupled ice-ocean models produce more realistic estimates of glacier retreat rates than ice models driven by parameterized melt.
1. Introduction

Thwaites Glacier (TG), in the Amundsen Sea Embayment (ASE) sector of West Antarctica, has long been identified as a potentially unstable, marine-based glacier [Hughes, 1981; Rignot, 2001] due to its retrograde bed sloping inland [Lindstrom and Hughes, 1984; van der Veen, 1985]. This has been referred to as West Antarctica's weak underbelly [Hughes, 1981]. The glacier is 120 km wide, discharged about 126 Gt/yr of ice into the ocean in 2014 [Mouginot et al., 2014], with a mass loss of 50 Gt/yr [Medley et al., 2014; Sutterley et al., 2014]. Portions of its bed are grounded more than 2000 m below sea level [Holt et al., 2006] and the entire basin contains an ice volume equivalent to 0.4-m of global sea level rise.

Observations of ice motion [Mouginot et al., 2014] and grounding line dynamics [Rignot et al., 2014] over the past three decades have revealed complex patterns of change. After almost a decade of near stability, the ice flow of Thwaites main trunk began to accelerate in 2006, affecting its entire basin, and leading to a flux increase of 33% [Mouginot et al., 2014]. The grounding line retreated 14 km over the main trunk between 1996 and 2011 and 1 to 9 km along the sides [Rignot et al., 2014]. The eastern part of the ice shelf, hereafter referred to as Thwaites eastern ice shelf, is buttressed by a large ice rumple, part of an east-west ridge about 40 km seaward of the present day grounding line [Rignot, 2001; Tinto and Bell, 2011; Millan et al., 2017]. The observed changes in this sector have been different than over the rest of the basin, with a doubling in ice velocity between 1996 and 2006, followed by a large slowdown until 2008, as the eastern ice shelf dissociated.
from the main floating tongue of TG [Mouginot et al., 2014]. Extensive dynamic thinning has been observed, reaching 4 m/yr over the coastal area [Pritchard et al., 2009]. Several studies have reported that the intrusion of warm, salty Circumpolar Deep Water (CDW) into Pine Island Bay through submarine glacial troughs initiated ice shelf thinning and glacier speed-up [Jacobs et al., 2011; Dutrieux et al., 2014]. Oceanographic observations in this area remain sparse but suggest a large interannual variability in ocean heat. For instance, the rate of ice shelf melt dropped by 50% in January 2012 during cold ocean conditions compared to warm ocean conditions in January 2010 [Dutrieux et al., 2014]. Ocean models have been used to investigate the intrusion of CDW and sub-ice-shelf basal melting assuming a fixed ice shelf cavity shape [Schodlok et al., 2012; Nakayama et al., 2014; St-Laurent et al., 2015]. Several studies, however, reported that basal melting is sensitive to the ice shelf cavity shape [Schodlok et al., 2012; Goldberg et al., 2012a]. Furthermore, other glacier modeling studies have shown that sub-ice-shelf melt rates exert an important control on ice dynamics on short time scales and that an increase in ice shelf melt may have already destabilized the West Antarctic Ice Sheet [Joughin et al., 2010; Parizek et al., 2013; Favier et al., 2014; Seroussi et al., 2014b; Joughin et al., 2014; Nias et al., 2016]. Simulating the ocean and ice streams as a coupled system is therefore essential to accurately capture the evolution of fast-changing outlet glaciers in West Antarctica. Previous studies, however, have focused on simplified geometries and idealized experiments [Goldberg et al., 2012a; Sergienko et al., 2013; De Rydt and Gudmundsson, 2016]. Here, we aim to address this challenge with actual glacier and ice shelf geometries.
We investigate the dynamics and evolution of TG over a 50-year period in response to varying ocean conditions using a coupled ice-ocean model. We employ the finite element Ice Sheet System Model (ISSM) coupled to the Massachusetts Institute of Technology general circulation model (MITgcm) to perform a set of simulations. We assess the evolution of TG over the next five decades and investigate the impact of changes in ocean conditions on ice shelf melt rates and glacier dynamics. We compare the results from coupled ice-ocean simulations with standalone ice sheet results with parameterized sub-ice shelf melt rate. We first describe the ice sheet and ocean models and the approach used to couple them. We detail the experiments performed and the results. We finally discuss the feedback mechanisms between ice dynamics and ocean circulation and how this impacts the evolution of TG over the coming decades.

2. Data and Methods

2.1. Ice flow model

We use the finite element software ISSM (Larour et al., 2012) to run simulations of TG ice flow using a two-dimensional shelfy-stream approximation (MacAyeal, 1989) (see supplementary material for stress balance approximation impact). Basal friction follows a linear viscous law and ice rigidity is based on ice temperature (Cuffey and Paterson, 2010). Depth-averaged temperatures are used to compute ice rigidity over the grounded part of the domain; ice temperature is assumed to be in steady-state and is kept constant throughout the simulation. If the ice sheet is not exactly in a thermal steady-state, this does not impact simulations conducted on decadal to centennial time scales (Seroussi et al., 2013). The model is initialized using data assimilation to match observed surface...
velocities of 2007 [Rignot et al., 2011]. Basal friction is inferred on grounded ice and ice rigidity is inferred on floating ice [Morlighem et al., 2010]; these variables remain unchanged during the simulations. Grounding line position is determined by a floatation criterion: ice is floating if its thickness is smaller than the floatation height and grounded otherwise. We use a sub-element parameterization in order to accurately capture the position of the grounding line and integrate basal friction accurately over the grounded part of the domain [Seroussi et al., 2014a].

The ice domain covers the catchment basin of TG (Fig.1). The horizontal mesh is made of anisotropic triangles, which size ranges from 500 m near the grounding line to 15 km close to the divides, resulting in a 44,500 element mesh (see supplementary material for resolution impact). P1 linear finite elements are used and the mass transport equation is stabilized using a streamline upwinding method. Two week time steps are chosen to satisfy the CFL (Courant-Friedrichs-Lewy) condition. At each time step, ice velocity, surface and base elevation and grounding line position are updated. The ice front position remains fixed during the simulations.

Structural collapse of marine-terminating ice cliff linked to atmospheric warming and hydrofracturing of ice shelves does not happen on TG until 2060, even in the most extreme warming scenarios [DeConto and Pollard, 2016]; this process is therefore unlikely to happen over the next few decades. The length of Thwaites main tongue fluctuated over the past decades [MacGregor et al., 2012], suggesting that its ice front is prone to variations; which is not included in our simulations.

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Bedrock elevation under grounded ice, and surface elevation for the ice streams and ice shelves are from Bedmap2 [Fretwell et al., 2013]. Bedrock elevation derived using mass conservation [Morlighem et al., 2011, 2013] and described in Rignot et al. [2014] are used upstream of Thwaites grounding line and nested into the Bedmap2 grid using kriging at the transition between the two datasets. Bathymetry under the floating ice shelves and in the open ocean is from IBCSO [Arndt et al., 2013], which includes gravity-derived bathymetry in front of TG [Tinto and Bell, 2011]. Surface mass balance and surface temperatures averaged over the 1979-2010 period from RACMO2 [Lenaerts et al., 2012] are applied at the surface of the domain, and the geothermal flux used to compute the ice steady-state temperature comes from Shapiro and Ritzwoller [2004].

2.2. Ocean model

The ocean model is the MITgcm [Marshall et al., 1997], which includes a thermodynamic sea ice model [Losch et al., 2010] and an implementation of sub-ice shelf melting and freezing parameterized as a diffusive flux of temperature and salinity [Jenkins et al., 2001; Losch, 2008]. The model is based on the global configuration that is part of the Estimating the Circulation and Climate of the Ocean, Phase II (ECCO2) project [Menemenlis et al., 2008; Fenty et al., 2015]. The ocean model domain includes all the Amundsen Sea Embayment, from Abbott to Getz ice shelves, and extends at least 100 km beyond the outer edge of the continental shelf (Fig.1).

The ocean grid is extracted from the 18-km Cube Sphere of Menemenlis et al. [2008], downscaled to a 2-km horizontal grid spacing, and includes 46 vertical levels, which constrains the time steps to 115 seconds. The initial and boundary conditions are interpo-
lated linearly from ECCO2 integration [Menemenlis et al., 2005], and atmospheric forcing is from the Japan Meteorological Agency and Central Research Institute of Electric Power Industry 25-year reanalysis [Onogi and others, 2007]. The same forcing from 1992 is repeated indefinitely during the simulations, as the objective of this study is to understand the interaction between the ice and the ocean with a realistic ocean forcing. The exchange of heat and fresh water at the base of the ice shelf is parameterized with constant turbulent exchange and friction velocity coefficients [Holland and Jenkins, 1999], as in Hellmer et al. [2012], Dinniman et al. [2012], and Schodlok et al. [2012, 2016].

2.3. Ice-ocean coupling

The ocean and ice models are operated on different grids with different resolutions to ensure appropriate simulation of the relevant processes. An asynchronous coupling is used to couple the ice and ocean models; information is exchanged between the two models every month through a MATLAB interface and the ocean model is restarted every month. The averaged sub-ice shelf melt rate over the last month is provided by the ocean model to the ice sheet model, while the geometry of the ice shelf cavity is provided by the ice sheet model to the ocean model. In areas where the ice shelf cavity opens, either vertically following ice shelf thinning or horizontally following grounding line retreat, ocean properties (e.g., heat, salt) from the closest grid cell are extrapolated to the new ocean cells. This scheme does not conserve mass, heat or salt budget, which is not critical for this study as we focus on the impact of ocean conditions on ice sheet dynamics and not detailed characteristics of ocean properties. Data are bilinearly interpolated from one grid to the other.
The ocean model is run as a standalone component for 5 years before being coupled to the ice sheet model in order to reduce the effect of initial conditions.

2.4. Experiments

The coupled ice-ocean model is run for 50 years under two different scenarios. In a first conservative experiment, referred to as 1992F, we model the glacier response to the ocean and atmospheric forcing of year 1992 repeatedly for 50 years. Year 1992 was a relatively cold year with the second lowest amount of melting on Pine Island Glacier over the last three decades [Schodlok et al., 2012]. In a second experiment, referred to as 1992F+, we test the impact of a change in ocean temperature on the glacier response by increasing the initial and boundary conditions by 0.5°C. As a prior study [Schmidtko et al., 2014] reported a warming of 0.1°C per decade in the ASE sector, the 1992F+ experiment represents a rather extreme scenario of ocean warming over the next 50 years. In a third experiment, we employ a standalone ice flow model of TG, referred to as 1992UC, forced with a depth-dependent parameterization of the sub-ice-shelf melt rate based on the 1992 climatology (see supplementary material).

3. Results

The simulated evolution of TG agrees with observations [Pritchard et al., 2009; Mouglinot et al., 2014; Rignot et al., 2014] and is consistent with prior modeling studies [Joughin et al., 2014; Parizek et al., 2013]. We attribute the agreement between our model and observations to improved model physics and a more accurate description of bed topography. All three experiments show an acceleration of TG (Fig.S4), increased mass loss, and a sustained grounding line retreat during the simulation period (see supplementary...
They suggest that TG will continue experiencing rapid grounding line retreat at about 1 km/yr, ice flow acceleration, and a mass loss increasing at an average rate of 1.2 and 1.6 Gt/yr\(^2\) for 1992F and 1992F+, respectively, over the next four decades.

The results of the 1992F experiment (Fig.1) show that the ocean model captures the intrusion of CDW onto the continental shelf toward the ice shelf cavities. The modeled thermocline, where water temperature increases from -0.8 to 0.8°C over about 200 m, lies around the 370-m depth on the continental shelf, which is consistent with year 2009 observations [Dutrieux et al., 2014]. The ice shelf melt water rates estimated for Pine Island and Thwaites ice shelves are, respectively, 88 and 81 Gt/yr, which agrees with melt estimates derived from remote sensing observations for the 2003-2009 period [Rignot et al., 2013; Depoorter et al., 2013]. The modeled ocean-induced melt rates have significant seasonal variations (Fig.2), with summer-to-winter changes of 20 Gt/yr, or 25% of the total melt water production in the 1992F experiment. The ice shelf melt rate is highest close to the grounding line, at about 60 m/yr, and decreases to less than 10 m/yr toward the ice front (Fig.3) [Schodlok et al., 2012; Nakayama et al., 2014].

Over the 50-year integration period, the model projects that TG will cause the equivalent of 12.1 and 13.3 mm of eustatic sea level rise for the 1992F and 1992F+ experiments, respectively (Fig.2). In both cases, the mass loss is faster over the last 20 years than over the first 30 years, with an annual ice loss equivalent to 0.21 and 0.29 mm of sea level rise for, respectively, the first 30 and last 20 years of the 1992F experiment. The 1992UC experiment shows a similar mass loss, equivalent to 12.9 mm of eustatic sea level rise over the 50-year integration period. The contribution to sea level rise of TG increases by only...
1.2 mm between the two coupled experiments, 1992F and 1992F+, as the ice shelf is mostly unconfined, but the difference between the two scenarios increases during the last 10 years, as a confined ice shelf develops along the main trunk of TG and provides buttressing to the grounded ice. Increasing the ocean temperature by 0.5°C has therefore a significant impact on the melt rate but a limited impact on the cumulative volume change. The ocean-induced melt rate increases by 30% on average and the additional volume loss is 9% (Fig.2). For comparison, the 1992UC shows an increase in the sub-ice-shelf melt rate from 78 to 132 Gt/yr, over the 50-year simulation, or 69%, which is three times higher than in the coupled experiments. The additional volume loss is 7%.

The coupled system allows the melt rate to quickly adjust to ice shelf thinning and grounding line retreat [De Rydt and Gudmundsson, 2016]. The melt decreases during the first 10 years of simulation, in response to the thinning of the eastern ice shelf, which causes the ice shelf base to rise above CDW currents, before it starts increasing again (Fig.2). The melt rates averaged over the first, 20th, and 50th years of the 1992F simulation are shown in Fig.3; newly ungrounded areas experience large melt rates, generally over 40 m/yr. Prior studies using simplified geometries indicated that the ocean circulation takes a couple of years to adjust to changes in ice shelf cavity and the glacier flow takes decades to adapt to changes in melt rates [Goldberg et al., 2012a, b]. Our results, with advanced modeling techniques and realistic geometries, reveal similar time scales.

The pattern of grounding line retreat is complex, with periods of rapid retreat alternating with periods of relative stability as the grounding line retreats over highs and lows in topography. We project significant retreat in three areas: over the main trunk (Fig.4...
point C), at the easternmost part of the eastern shelf (point A, 15-km retreat during the simulation period), and on the westernmost part, close to Haynes Glacier (point D, 28-km retreat). The grounding line is comparatively stable in between these areas in the coupled experiments (Fig.4a and b). The grounding line retreats by 20 km over the main trunk before stabilizing on a ridge (point C) in the bedrock topography after 30 and 25 years in, respectively, the 1992F and 1992F+ scenarios. The retreat rate varies from 0.5 to 1.5 km/yr during this period, which is consistent with the rate derived from satellite observations for the period 1992-2011 [Rignot et al., 2014]. After 30 years (for 1992F) and 24 years (for 1992F+) of simulation, the grounding line retreats into a deeper bed on the eastern part of the main trunk (point B). This additional retreat is more pronounced in the 1992F+ scenario and extends 17 km eastward relative to 1992F (Fig.4a and b). Overall, the 1992F+ scenario causes the grounding line to retreat 100 to 200 m/yr faster than the 1992F scenario (Fig.S5), however, the pattern of retreat is similar and the grounding line stabilizes over the same topographic features. The grounding line retreat in the 1992UC experiment is significantly faster than in the coupled simulations; the ungrounding until the ridge at point C (Fig.4c) takes 12 years in this simulation, i.e., a doubling of the retreat rate in this area, which is higher than observed [Rignot et al., 2014]. A large retreat of 25 km also takes place between points A and B, principally during the last 15 years of the simulation, while in the coupled experiments only a minor retreat takes place in this sector.

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4. Discussion

The results presented here reflect important improvements in observations (e.g., bedrock topography derived with a mass conservation algorithm [Morlighem et al., 2011], inclusion of the eastern shelf ice rise), ice sheet modeling (e.g., sub-element parameterization of grounding line position [Seroussi et al., 2014a]), and coupling of ice sheet and ocean models for the first time in a non-idealized ice-ocean system. Ice/ocean coupling, in particular, appears to be critical. The 1992UC forced with parameterized sub-ice-shelf melt rate calibrated from an ocean simulation, leads to significantly faster grounding line retreats in the eastern part of the domain, which differs from observed grounding line retreat [Rignot et al., 2014]. The parameterized melt rate overestimates the melting close to the grounding line, where the water column is shallow and ocean circulation is weak. Comparison of melt rates at the beginning and end of the simulation (Figs.S1 and S2) shows that the depth-dependent parameterization does not hold as the ice shelf cavity evolves because it overestimates how easily the ocean heat is transferred to the sub-ice-shelf cavity. Using such a parameterization therefore overestimates the retreat rate of rapidly-changing glaciers over the short 50-year integration period, highlighting the importance of using coupled models for century-scale simulations of glaciers in the West Antarctic Ice Sheet.

Limitations of our simulations include the absence of tides and constant heat and salt transfer coefficients in the ocean model. Observations [Jenkins et al., 2010] show a velocity dependence of these coefficients. Using this velocity dependence, however, does not allow the reproduction of sub-ice shelf melting rates observed by remote sensing because...
of the coarseness of the ocean circulation model: the weak ocean circulation simulated close to the grounding line causes low melt rates in these areas, while remote sensing observations of ice shelves [Rignot et al., 2013; Depoorter et al., 2013] show that the highest melt rates are located close to the grounding line. Using constant transfer coefficients allows capturing this pattern (Fig.3). Dansereau et al. [2014] compared spatial patterns and magnitude of sub-ice shelf melt rates derived with velocity-dependent and velocity-independent transfer coefficients; they concluded that uncertainties remain regarding both spatial patterns and magnitude of melt rates and that additional observations were needed to better calibrate these parameters. Tidal circulation has the largest effect on cold ice shelves [Makinson et al., 2011], while the effect of tides is weak for ice shelves bathing in CDW, as currents under the ice shelf are mainly controlled by meltwater-driven flows [Dutrieux et al., 2014]. A uniform increase in ocean temperature is unrealistic, and several studies suggest that changes in the circulation of CDW mainly influence the depth of the thermocline [Dutrieux et al., 2014] but this remains beyond the scope of this paper. The calving front remains fixed in time and the rigidity of the ice shelf does not evolve with time. Our scenario therefore presents conservative estimates, as Jeong et al. [2016] reported an increase in ice shelf rift formation, which resulted into ice shelf weakening and calving happening further upglacier than previously observed for Pine Island Glacier. Large calving events could cause a more rapid and unstable retreat, especially if the eastern ice shelf were to break up.

Our numerical experiments support the hypothesis that the grounding line of TG displays an irreversible character [Joughin et al., 2014] due to the bed topography: the
grounding line stabilizes temporarily over the same bedrock ridges in all experiments. The oceanic forcing, however, provides a critical modulation of the retreat rate. In warmer ocean conditions, the grounding line takes less time to reach the stabilizing ridges because ice shelf melting is higher. As new, confined ice shelves form during the retreat, ice shelf melting will be less efficient than in unconfined cavities used to develop parameterized ice shelf melt rate. A coupled ice-ocean model produces lower and more realistic estimates of ice shelf melt. Accurate knowledge of oceanic conditions beneath the ice shelves and precise bed topography along the areas of retreat, similar to the high-resolution bed topography of the ASE \cite{Rignot2014}, and fully-coupled ice/ocean models are therefore critical to realistic projections of glacier evolution over the upcoming decades.

5. Conclusions

Simulations of TG performed using a fully coupled ice-ocean numerical model are able to accurately reproduce the pattern of acceleration and grounding line retreat observed over the past decade. Our simulations suggest that these changes will continue over the coming decades as the grounding line retreats into deeper ground. The bed topography is a primary control of the grounding line retreat, which implies that high-resolution reconstruction of the bed topography, as used herein, is essential to model the glacier flow. We find that thermal forcing from the ocean controls the rate of retreat, as warmer ocean waters accelerate the retreat by 100-200 m/yr, and this impact increases as confined ice shelves develop. The melt rate quickly adjusts to changes in the ice shelf cavity, which implies that coupled ice-ocean models are essential to better project the evolution of this rapidly changing glacier. In contrast, uncoupled ice models employing a depth
parameterization of sub-ice shelf melt derived from ocean models may overestimate the rate of grounding line retreat.

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Goldberg, D. N., C. M. Little, O. V. Sergienko, A. Gnanadesikan, R. Hallberg, and M. Oppenheimer, Investigation of land ice-ocean interaction with a fully coupled ice-


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Figure 1. Potential ocean temperature (°C) at 513-m depth simulated with MITgcm over the ocean, and ice velocities (m/yr) over the grounded ice. Grey areas are oceanic regions with bathymetry above 513 m and white areas are covered by land ice. The solid black line represents the 2008 ice front, white lines show the 2011 grounding lines, and black dashed line the ice model domain.
Figure 2. Evolution of total melt rate (left) under the floating part of TG and change of ice volume above floatation (right) over Thwaites model domain (dashed black contour on Fig.1). Blue, red, and green curves represent, respectively, the 1992F, 1992F+, and 1992UC experiments.
Figure 3. Average melt rate (m/yr) under the floating ice simulated by the ocean model for year 1 (left), 20 (middle), and 50 (right) for the 1992F experiment. The black line marks the ice domain contour and blue lines indicate water column thickness (m) under the ice shelf.
Figure 4. Evolution of grounding line position over the 50-year simulations for the 1992F (left), 1992F+ (middle), and 1992UC (right) experiments. Background shows the bed topography (m). The green and purple lines represent, respectively, the 1996 and 2011 observed grounding line positions [Rignot et al., 2014].