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Key Points:

- Atmospheric changes will govern East Antarctic Ice Sheet (EAIS) mass balance to 2100
- EAIS is projected to lose mass under RCP2.6 but gain mass under high-emission scenarios by 2100
- Mass loss from Totten Glacier is dependent on the retention of a vulnerable 10-km ice plain

Supporting Information:

- Supporting Information S1

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Aurora Basin, the Weak Underbelly of East Antarctica

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Abstract The East Antarctic Ice Sheet (EAIS) has the potential to raise global sea levels by ~52 m. Here, we model the evolution of select EAIS catchments to 2100 using three basal melt rate parameterizations and force our model with surface mass balance and ocean thermal anomalies from 10 global climate models. While the domain loses mass under low-emission scenarios, it gains ~10-mm sea-level rise equivalent ice volume (SLRe) under high-emission scenarios. The primary region of thinning is within 50 km upstream of Totten Glacier's grounding line. Totten's glacial discharge is modulated by the migration of its grounding line, which is sensitive to brief intrusions of ocean water at temperatures higher than present. Once the grounding line is dislodged, Totten's ice velocity increases by up to 70% of present-day values, resulting in ~6-mm SLRe loss from this sector.

Plain Language Summary Predicting how much the East Antarctic Ice Sheet (EAIS) will contribute to global sea-level rise is critical to the welfare of the global community. In our study, we model a large sector of EAIS to 2100 using a variety of different ocean melting representations and include both ocean temperature and snowfall predictions taken from two suites of climate model output. We find that by 2100, increases in snowfall outweigh increases in ice loss, and this sector of the EAIS will be responsible for a 10-mm decrease in global sea level under the highest climate warming scenarios. We find that Totten Glacier is most at risk to enhanced ocean warming, with the southernmost portion of this glacier acting as an important control on whether it loses ice or remains at its present-day configuration. If this portion of Totten is melted, this glacier may lose enough ice mass to raise global sea levels by 6mm by 2100.

1. Introduction

Ice sheet mass balance is a major component of the global sea-level budget. While the majority of scientific focus has been aimed at Greenland and the West Antarctic Ice Sheet, low-lying sectors of the East Antarctic Ice Sheet (EAIS) have been shown to be vulnerable to warmer-than-present temperatures during the Last Interglacial and Pliocene epoch (Aitken et al., 2016; Cook et al., 2013; DeConto & Pollard, 2016). During these epochs, glacial margins retreated several hundreds of kilometers inland over the Wilkes Subglacial Basin (WSB), caused by elevated Southern Ocean temperatures (Cook et al., 2013). Additionally, large-scale advance and retreat patterns of EAIS within the Aurora Subglacial Basin (ASB) have been inferred from spatial patterns of inland bed erosion, leading to over a 4-m loss of sea-level rise equivalent ice volume (SLRe) from this region during the Pliocene epoch (Aitken et al., 2016). Recent observations show that EAIS contains ~52-m SLRe and has lost 51 ± 13 Gt/year of ice between 2009 and 2017, comprising 20% of the total mass loss of Antarctica during this period (Morlighem et al., 2019; Rignot et al., 2019). This mass loss and acceleration is a dynamic response of the ice sheet to enhanced oceanic thermal forcing at glacial margins (Rignot et al., 2019). Increased basal melting and resultant ice shelf thinning leads to a reduction of the backstress that stabilizes upstream ice, causing accelerated grounded ice discharge (Li et al., 2015; Rignot et al., 2019; Rignot & Jacobs, 2002; Schroeder et al., 2017; Velicogna et al., 2014).

Totten catchment, located in the ASB of East Antarctica, contains 3.5-m SLRe and is one of the few sectors of EAIS undergoing dynamic changes (Greenbaum et al., 2015). Li et al. (2016) reported a mass loss from Totten of 6.8 ± 2.2 Gt/year with an acceleration of 0.55 ± 0.27 Gt/year² between 1989 and 2015. This acceleration in grounded ice flux has been linked to intrusions of warm Modified Circumpolar Deep Water (0–1 °C) into Totten's subice shelf cavity (Greenbaum et al., 2015; Li et al., 2016; Mohajerani et al., 2018; Rintoul et al., 2016; Silvano et al., 2017, 2019). Additionally, a 1- to 3-km retreat of the southernmost portion of

Figure 1. (a) EAIS_{SUB} domain showing the number of models (total 36) that predict at least 1-m grounded ice thinning by 2100. Basin outlines are included, and the associated far ocean temperature and salinity (used by PICO and PICOP) are displayed for each basin. The red box around Totten Glacier shows the domain of (b). The present-day grounding line and ice front are overlaid in black (Morlighem et al., 2019). (b) Bed topography (Morlighem et al., 2019) of Totten Glacier with modeled 2100 grounding lines from the forced PICOP projections as well as the control and present-day grounding lines. The “eastern flank” and “ice plain” regions are boxed in white, and the velocity flowline used in Figure 3 is included as the white dotted line.

Totten’s grounding line was observed between 1996 and 2013, concurrent with changes in subshelf ocean temperature (Li et al., 2015). It remains unclear, however, how EAIS and Totten in particular are going to respond to changes in surface mass balance (SMB) and ocean thermal forcing (TF) over the coming century.

Ice sheet models generally rely on prescribed SMB inputs, which are typically output fields of Regional Climate Models or Atmosphere-Ocean General Circulation Models (AOGCMs). However, resolving Antarctic glacial discharge is challenging because it is largely forced by changes in oceanic heat flux (Rignot et al., 2013). As such, it is necessary to translate changes in ocean heat content at the bottom of ice shelves to basal melt rates. This remains an active field of study in the ice sheet modeling community, and, while ice-ocean model coupling remains the best way to address this question, it is not yet computationally feasible at the scale of EAIS (Seroussi et al., 2017). Thus, we rely on basal melt parameterizations to simulate ice-ocean interactions in our ice sheet model.

In this study, we aim to model mass balance changes of EAIS by 2100 and determine what controls the evolution of its most dynamic catchment. We model six basins in EAIS (hereafter referred to as EAIS_{SUB}) using the Ice Sheet System Model (ISSM, Larour et al., 2012) and three newly developed basal melt rate parameterizations: a nonlocal quadratic melt parameterization (nonlocal, Favier et al., 2019), the Potsdam Ice-shelf Cavity mODEl (PICO, Reese et al., 2018), and PICO coupled to a buoyant plume model (PICOP, Pelle et al., 2019). We force our model with cumulative anomalies in both SMB and TF, taken from six AOGCMs within the Climate Model Intercomparison Project 5 (CMIP5) model ensemble and four AOGCMs from the Climate Model Intercomparison Project 6 (CMIP6) model ensemble. We investigate the modeled response of EAIS_{SUB} mass balance to these given forcings and perform a refined analysis on Totten Glacier, the primary source of mass loss in EAIS_{SUB}. We establish plausible bounds on the contribution of EAIS_{SUB} to global sea-level rise by the end of the century and determine the main controls on Totten’s mass loss.

2. Data and Methods

Contemporary EAIS mass loss has been dominated by the signal from the ASB, which contains several marine-terminating outlet glaciers (Shepherd et al., 2018). In contrast, Victoria Land, Queen Maud Land, and Amery Ice Shelf have displayed little change over the observational period (Lovell et al., 2017; Rignot et al., 2019). We partition a large sector of EAIS into six basins based on Zwally et al. (2012), containing Amery Ice Shelf (Basin 1), Queen Mary Land (Basin 2), ASB (Basin 3), WSB (Basin 4), Oates Land (Basin 5), and Victoria Land (Basin 6, Figure 1a and supporting information Figure S1). We exclude the Filchner-Ronne Ice Shelf, Queen Maud Land, and the Ross Ice Shelf from our study due to their current stability (Rignot et al., 2019) and to partition computational resources to more dynamic basins. For a complete description of the numerical model and its associated limitations, see the supporting information.

We force our ice sheet model with yearly output of SMB and TF from six AOGCMs in the CMIP5 model ensemble (CCSM4, MIROC-ESM-CHEM, NorESM1-M, CSIRO-Mk3-6-0, HadGEM2_ES, and IPSL-CM5A-MR) and four AOGCMs in the CMIP6 model ensemble (CNRM-CM6-1, CNRM-ESM2-10, UKESM-0-LL, and CESM2). These data sets were prepared for the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6, Nowicki et al., 2016). For a complete description of the AOGCM selection process, see the supporting information and Barthel et al. (2019).

We rely on three basal melting rate parameterizations to simulate ice-ocean interactions in our model: non-local quadratic, PICO, and PICOP. The nonlocal quadratic parameterization aims to resolve the positive feedback between subshelf ocean circulation speed and basal melting (Favier et al., 2019), capturing both local melt and subshelf cavity-integrated melt (Jourdain et al., 2017). This feedback is crudely represented by multiplying the basin-averaged TF by the local TF, forming a quadratic dependence of the computed melting rate on the TF at the ice-ocean interface. Basin-scale temperature corrections were computed as to minimize the error between parameterized melt rates and those from Rignot et al. (2013). PICO is a two-dimensional box model that was derived to simulate vertical overturning in the subshelf cavity, solving the transport of heat and salt between boxes by assuming steady state (Little et al., 2009; Reese et al., 2018). Ocean water, with a given basin-averaged temperature and salinity, is entrained into the subshelf cavity and melts the ice shelf. The melt water produced is fresh and buoyant, rising along the base of the ice shelf and inducing the strongest melt near the grounding line. Pelle et al. (2019) coupled PICO to a buoyant plume parameterization developed by Lazeroms et al. (2018), forming PICOP. Here, the subshelf temperature and salinity fields computed by PICO are used as inputs in the plume emulator, along with the full grounding line depth field. We use the best fit overturning strength and effective turbulent heat exchange velocity parameters from Reese et al. (2018) in both PICO and PICOP. Overall, these parameterizations were chosen as to vary in complexity and maximize the diversity of resolved subshelf ocean processes. See Figure S9 for examples of basal melting rate fields computed by each of the parameterizations.

Our simulations span the period from January 2015 to December 2100. We force our model with anomalies in SMB and TF, taken from the AOGCMs listed above under RCP8.5 (CMIP5, Taylor et al., 2012) and SSP585 (CMIP6, Eyring et al., 2016) emission scenarios. We run two additional experiments using forcing fields from NorESM1-M and IPSL-CM5A-MR under RCP2.6 to evaluate the response of our model to lower-emission scenarios. When computing the SMB forcing field, the reference climatology is the present-day SMB field from the Regional Atmospheric Climate Model (RACMO 2.3, van Wessem et al., 2014). When computing the TF field using the nonlocal melt parameterization, the reference climatology is derived from the three observational ocean data sets described in section 2.2.2 of Barthel et al. (2019). When using PICO and PICOP, we use the far-field ocean temperatures for basins 6 through 11 in Reese et al. (2018) and hold salinity constant (Figure 1a).

We first perform three control runs over $EAIS_{SUB}$ for the duration of the aforementioned experimental period to evaluate model drift. One control run is completed for each melting rate parameterization and a 10-year model relaxation period precedes the start of all simulations, as to allow the model to adjust to initial conditions (Figure S3). For the duration of the relaxation period and control runs, SMB and TF fields are held constant at their respective reference climatology fields. In addition to the $EAIS_{SUB}$ -forced simulations, we repeat the experiments on our regional model of Totten catchment using a refined mesh (~ 500 m resolution in the grounding zone) to study the response of this glacier's grounding line and ice dynamics to the applied forcings. Mass balance changes are reported relative to the control experiments (i.e., results should

Figure 2. Change in total mass balance (blue, green, and red bars), SMB (light gray bars), and glacial discharge (dark gray bars) for all 36 simulations in both mm SLRe and Gt (positive-upwards bars indicate EAIS_{SUB} mass gain, and negative-downwards bars indicate EAIS_{SUB} mass loss). Bars are grouped by AOGCM forcing and color coded by the melting rate parameterization used. Emission scenarios are indicated by background shading: CMIP5 RCP8.5 (red shading, left), CMIP5 RCP2.6 (blue shading, center), and CMIP6 SSP585 (yellow shading, right). Means are plotted for each scenario as a black dashed line, and the standard deviation ($\pm 1\sigma$) is visualized as the background shading within the dotted lines. The mean mass balance change per melting rate parameterization is displayed in the legend.

be interpreted as the system's response to additional climate forcing compared to the system under forcings representative of the period spanning January 1995 to December 2014).

3. Results

We observe a varied response of EAIS_{SUB} mass balance to 2100 (Figure 2). Overall, our CMIP5 RCP8.5 and RCP2.6 projections show that EAIS_{SUB} will gain 9.73 ± 10.05 -mm SLRe and lose 1.30 ± 3.35 -mm SLRe by the end of the century, respectively. For the CMIP5-forced projections, the standard deviation in EAIS_{SUB} mass

Figure 3. (a–l) Velocity profiles (m/year) taken along the flowline in Figure 1b are plotted yearly against distance from the ice front (km), with the color of the line coinciding with the year of the projection. Vertical lines correspond to the position of the grounding line along the flowline. The percent velocity increase along this flowline is printed in the top-right corner of each panel. Panels outlined in red display the velocity profiles for projections where Totten's ice plain did not dislodge. (m) Far ocean temperature time series used as input for PICOP in projections of Totten Glacier. The model relaxation period spans the period from January 2005 to December 2014, in which the far ocean temperature was held fixed at 272.42 K (Reese et al., 2018).

balance change associated with choice of AOGCM forcing and melting rate parameterization is 10.29-mm SLRe and 2.37-mm SLRe, respectively. As such, variance in our projections of total mass balance is primarily attributed to the range of AOGCM forcings applied. The standard deviation associated with glacial discharge and total SMB between all CMIP5 experiments is 3.82-mm SLRe and 10.05-mm SLRe, respectively, revealing that significant differences in SMB accumulation between AOGCMs are responsible for the large spread in our projections. Meanwhile, the CMIP6 SSP585-forced models project a mean mass gain of 13.76 ± 5.89 -mm SLRe by 2100. The standard deviation in applied SMB and glacial discharge is 5.17-mm SLRe and 2.44-mm SLRe, respectively.

The SMB contribution dominates the total mass balance signal in our model when forced by MIROC-ESM-CHEM, CCSM4, CSIRO-Mk3-6-0, IPSL-CM5A-MR RCP8.5, and all CMIP6 AOGCMs, leading to substantial mass gains for all melt parameterizations used (Figure 2). These climate models project between 2.5 and 3.35 °C atmospheric warming over Antarctica by 2100, resulting in significant increases in SMB that are concentrated near coastal sectors (Figure S10, Barthel et al., 2019). When using forcings from MIROC-ESM-CHEM, CCSM4, and CESM2, the SMB signal is over four times greater than that of the glacial discharge signal and mass gains exceed 20-mm SLRe by the end of the century. When forced by CSIRO-Mk3-6-0, IPSL-CM5A-MR RCP8.5, CNRM-CM6-1, CNRM-ESM2-10, and UKESM-0-LL, total ice mass gains are on average 10-mm SLRe, resulting from the reduced but still significant SMB contribution. In contrast, for the AOGCMs that project less snowfall across EAIS_{SUB} (NorESM1-M, HadGEM2-ES, and

IPSL-CM5A-MR RCP2.6), total mass balance changes are largely negative or near zero and become more dependent on glacial discharge. Our model predicts mass loss when using forcing fields from NorESM1-M RCP2.6 and HadGEM2_ES. Although HadGEM2_ES projects 2.9 K atmospheric warming over Antarctica by 2100 (similar to IPSL-CM5A-MR), reductions in SMB along the coast between Amery and Totten Glaciers offset interior increases, resulting in only 2.21-mm SLRe SMB gain by 2100 (Figure S10, Barthel et al., 2019). In addition to differences that arise from the applied SMB forcing, we see variation in discharge that arises from the choice of basal melting rate parameterization. When considering the RCP8.5- and SSP585-forced simulations, we find that the use of the nonlocal parameterization, PICOP, and PICO leads to -7.90 -mm SLRe, -5.61 -mm SLRe, and -3.54 -mm SLRe mean discharge by 2100, respectively (Figure 2).

A majority of the projected mass loss is isolated to the ASB (Figures S4–S6), while the rest of EAIS_{SUB} either gains mass or remains in near mass balance by the end of the century. In particular, grounded ice thinning is most substantial along the periphery of Totten's grounding line, with almost all projections showing at least 1-m grounded ice thinning by 2100 (Figure 1a). Significant mass loss from Totten is observed when forced by NorESM1-M (RCP8.5 and RCP2.6), CSIRO-Mk3-6-0, and HadGEM2_ES (Figure S5). By 2100, these AOGCMs project a shelf-wide mean ocean warming of $+0.86$, $+0.23$, $+1.09$, and $+0.86$ °C, respectively, over the present-day average ocean temperature of 272.42 K (Figure 3m, Reese et al., 2018). While the mean ocean warming projected by NorESM1-M RCP2.6 is minimal relative to other AOGCMs, ice discharge increases significantly after 2040 by an abrupt 0.4 °C increase in mean ocean temperature between 2030 and 2040, with minimal projected SMB accumulation to offset the subsequent mass loss (Figure 3m). Glacial discharge remains constant after 2060 at a rate of approximately 18 Gt/year, as the subshelf ocean temperature remains nearly constant at 272.60 K through 2100 (Figure S5). Significant ocean warming is also projected by MIROC-ESM-CHEM ($+0.86$ °C with a temporal pattern almost identical to that of NorESM1-M RCP8.5, Figure 3m); however, large SMB accumulations outweigh the glacial discharge excited by this ocean warming. Our model forced by IPSL-CM5A-MR RCP8.5 and RCP2.6 project Totten to remain in near mass balance through 2100. In the RCP8.5 scenario, we see a significant mean ocean warming of $+0.47$ °C that only results in approximately 1.5-mm SLRe mass loss. This significant ocean warming is projected to occur after 2070, which does not leave enough time for Totten's upstream grounded ice to fully respond to the subsequent ice shelf thinning. As such, we expect that glacial discharge would accelerate in response to this late ocean temperature increase if we extend our projections beyond 2100.

We replicated the experiments with a spatially refined mesh (minimum edge length 500m in and upstream of the grounding zone) on a domain that only contains Totten Glacier to investigate controls on its glacial discharge. We rely on PICOP to compute basal melt rates in these experiments because the spatial distribution and magnitude of computed melt rates match best with observations (Pelle et al., 2019). The ocean temperature is initially set to 272.42 K and evolves in time with anomalies taken from each of the AOGCMs (Figure 3m). In Figure 1b, we observe grounding line retreat along two main sectors of Totten's grounding line: the eastern flank and southernmost portion ("ice plain"). Retreat along the eastern flank is ubiquitous across all experiments with a maximum inland distance of ~ 50 km by 2100. While the distance of upstream grounding line retreat in this sector depends on the ocean temperature applied, the initiation of retreat is temporally uniform and begins immediately in all simulations. The grounding line stabilizes through 2100 along specific topographic highs in this region; however, regions of low-lying bed topography adjacent to these topographic highs are susceptible to melting and allow for ongoing retreat in our simulations under present-day ocean conditions beyond 2100 (Figures 1b and S8, Morlighem et al., 2019).

In contrast, the timing of the initiation of grounding line retreat along the ice plain is dependent on the applied ocean temperature. Figure 1b shows that Totten's grounding line dislodges from this ice plain in every forcing scenario except CCSM4, IPSL-CM5A-MR RCP2.6 (the ice plain only partially dislodges in this projection), CNRM-CM6-1, and CESM2. In all other projections, this region dislodges at various times, occurring as early as 2050 when using HadGEM2_ES forcings. In Figure 3, yearly velocity profiles are extracted along the flowline from Figure 1b and are plotted against distance from Totten's ice front. Significant increases in ice surface velocities coincide with the timing of the dislodging of Totten's upstream ice plain. These velocity increases approach 70% of the initial velocity along the flowline when forced by CSIRO-Mk3-6-0 and extend 50 km upstream of the grounding line, caused by the extreme ocean warming projected by this AOGCM (Figures 3e and 3m). The greatest velocity increases are located right along and just downstream of this grounded ice plain, approximately 100 to 150 km upstream from Totten's ice front. Velocity increases are minimal in both IPSL-CM5A-MR RCP8.5 and NorESM1-M RCP2.6 (Figures 3f and

3g), as the grounding line did not dislodge from the ice plain until approximately 2090, leaving only 10 years for upstream, grounded ice to respond. In the projections in which the ice plain did not dislodge, we observe velocity increases of less than 20% of the initial ice velocity and no distinct period of acceleration. Note that when we extend the simulation period and decrease the ocean temperature to 272.42 K, the grounding line begins to readvance toward the ice plain (see the supporting information).

4. Discussion

Our projections of future $EAIS_{SUB}$ mass balance highlight the large degree of uncertainty that exists in projected climate forcings, parameterized physical processes and the resulting mass balance of this sector in the coming decades. The recent IMBIE-2 multisensor assessment found that between 1992 and 2017, EAIS gained 5 ± 46 Gt/year of ice mass (1.2 ± 11 -mm SLRe by 2100 if this trend remains constant, Shepherd et al., 2018). The high degree of uncertainty was primarily attributed to errors in the observed SMB over the observational period. In contrast, Rignot et al. (2019) reported a mass loss of -57 ± 2 Gt/year (-13.6 ± 0.5 -mm SLRe by 2100) from EAIS over the same period using updated regional SMB models. Outlet glaciers in the ASB dominated mass loss in this assessment (e.g., Totten Glacier, which contributes approximately 7.3 Gt/year between 2003 and 2017, Rignot et al., 2019). Assuming these trends remain constant through 2100, our projections forced by NorESM1-M, CSIRO-Mk3-6-0, HadGEM2_ES, IPSL-CM5A-MR, CNRM-CM6-1, and CNRM-ESM2-1 fall within the error of the IMBIE-2 assessment. In contrast, none of our projections fall within the error bounds of Rignot et al. (2019) due to the overwhelming SMB signal and limited dynamical response of outlet glaciers in both CMIP5- and CMIP6-forced simulations.

Antarctic surface air temperatures are projected to increase by the end of the century in each of the selected CMIP5 and CMIP6 AOGCMs. As such, continent-wide snowfall is expected to increase across Antarctica, as warmer air can hold exponentially more water vapor. Accordingly, projections of Antarctic SMB to 2100 show that if no dynamical ice response is assumed, snowfall is expected to increase by 6–16%, translating to a 20- to 43-mm drop in global sea level (Ligtenberg et al., 2013). While this response is integrated over the entire AIS, it highlights that significant increases in SMB are a distinct possibility across $EAIS_{SUB}$ by the end of the century. This rationale also clarifies why our model forced by CMIP5 RCP8.5 scenarios gain significantly more mass than when forced with the respective RCP2.6 scenarios, regardless of the melt parameterization used (Figure 2). Air temperatures increase, on average, $\sim 4^\circ\text{C}$ more in the RCP8.5 scenarios when compared to their respective RCP2.6 scenarios, resulting in an approximate 100% increase in the water vapor capacity of the air. Meanwhile, basin-averaged ocean temperatures increase, on average, $\sim 0.40^\circ\text{C}$ more in the RCP8.5 scenarios, resulting in an approximate 45.5% increase in total integrated basal mass balance when using PICOP. Although both atmospheric and oceanic temperature increases are significant, the ice dynamic response to this enhanced thermal forcing is minimal, while the impact of the atmosphere is substantial. As such, it is plausible that, aside from select glaciers in the ASB, large-scale $EAIS_{SUB}$ ice dynamics are not particularly sensitive to the increases in subshelf ocean temperatures that are projected at the timescale of this study. Rather, short-term atmospheric changes will govern $EAIS_{SUB}$ mass balance to 2100. Although counterintuitive, this means that lower-emission scenarios (RCP2.6) will lead to greater $EAIS_{SUB}$ mass loss than higher-emission scenarios (RCP8.5 and SSP585) by 2100.

In comparing how $EAIS_{SUB}$ mass change varies between projections forced with the same AOGCM output using different melting rate parameterizations, we find that the nonlocal quadratic melting parameterization generally produces the greatest amount of glacial discharge by 2100, followed by PICOP and PICO. Basal melting rates computed by the nonlocal parameterization depend on the local and basin-averaged thermal forcing, whereas PICOP takes into account the subice shelf geometry (i.e., the depth of the grounding line and basal slopes). These additional dependencies limit the magnitude of basal melting produced by PICOP beneath ice shelves that are simulated to have high melt rates using the nonlocal parameterization (Figure S9). In addition, the spatial distribution of the computed melt rate fields is impacted by these dependencies, driving differences in the modeled patterns of grounding line retreat. In general, melt rates computed by the nonlocal parameterization are more spatially uniform, while those computed by PICOP are channelized, with high melt rates along the plume path and low melt rates elsewhere. These two parameterizations project comparable amounts of total integrated basal mass balance within warm basins. This is exemplified in the NorESM1-M RCP8.5-forced simulation of Totten (Figure S9), where the nonlocal parameterization produces high melt across the ice shelf and grounding line that leads to large-scale retreat of the eastern flank. When modeled with PICOP, however, simulated melt rates are high along the interior of

the ice shelf, leading to ice shelf thinning that propagates upstream and dislodges the ice plain. In colder basins where simulated subshelf melt rates are low, upstream propagation of this interior ice shelf thinning is slow and does not impact grounding line dynamics as readily as melt applied directly at the grounding line. As such, melt rates produced by the nonlocal parameterization generally lead to higher amounts of glacial discharge within the timescale of our study.

In our projections of Totten Glacier, the magnitude of present-day melt rates computed by PICOP agrees well with Gwyther et al. (2014) and Greene et al. (2018), who model maximum basal melting rates of 60 to 80 m/year both along and downstream of the ice plain. Grounded ice thinning along this ice plain varies with the projected magnitude of subshelf ocean temperature increase. We observe maximum thinning rates of 2.35 m/year under the highest ocean warming scenario (CSIRO-Mk3-6-0), 0.46 m/year under the lowest ocean warming scenario (CESM2), and a mean thinning rate of approximately 1.15 m/year that is representative of moderate ocean warming. This mean projection is consistent with Flament and Remy (2012), who observed dynamic thinning near Totten's grounding line on the order of 1.2 ± 0.6 m/year, between 2002 and 2010. Our results also agree with ICESat observations between 2003 and 2008, which measured thinning rates of 1.7 ± 0.2 m/year in this region (Paolo et al., 2015). Li et al. (2015) observed average thinning rates of 0.7 ± 0.1 m/year in this region between 1996 and 2013, resulting in a 1- to 3-km grounding line retreat approaching Totten's ice plain. Immediately following this retreat, a $10 \pm 5\%$ increase in glacier velocity was observed along Totten's main ice trunk (Li et al., 2016), agreeing with our findings that this ice plain acts as an important control on ice dynamics. The observed pattern of grounding line retreat in this study is consistent with the findings of Sun et al. (2016), who demarcated widespread retreat over the eastern flank and minimal retreat along Totten's southern sector, due to the coastal-sloping bed in this region.

In projections with minimal ocean warming (CCSM4, IPSL-CM5A-MR RCP8.5, CNRM-CM6-1, and CESM2), we model grounding line advance downstream of the ice plain, as basal melting rates are too low in this region due to PICOP's plume parameterization (<10 m/year compared to 80 m/year in Greene et al., 2018, Figure 1b). In the other projections, short-term increases in subshelf ocean temperature were sufficient to initiate grounding line retreat along Totten's ice plain (Figure 1). This is exemplified in the projection forced by NorESM1-M RCP2.6, in which Totten's ice plain dislodged after 10 years of ocean temperatures increased 0.4°C above the present-day value of 272.42 K (Figure 3a, Reese et al., 2018). This present-day mean subshelf ocean temperature acts as a threshold in our model; warm ocean water intrusion at temperatures above this results in grounding line retreat past Totten's ice plain, and temperatures below this result in stabilization (Figures S4 and S5). The ability of Totten's grounding line to readvance during cold water intrusions highlights that at present (i.e., within the plausible ocean temperature range to 2100), Totten is not susceptible to Marine Ice Sheet Instability (MISI). During these events, Totten's grounding line readvances downstream over a 50-km prograde portion of bed topography; however, upstream of this, the bed becomes retrograde toward a deep marine basin adjacent to Law Dome, posing a much more significant threat for MISI. As the ice plain sector drains the majority of Totten's ice mass, small changes in grounding line position can induce large changes in the rate of grounded ice discharge. Due to the control that this portion of Totten's grounding line has on the discharge of grounded ice, its inherent sensitivity to brief increases in ocean temperature, and its multimeter sea-level rise potential, scientific focus should remain on this sector of EAIS_{SUB} through the turn of the century.

Overall, the projections presented here display a wide range of plausible contributions of EAIS_{SUB} to global sea-level rise and target Totten Glacier as the main source of future grounded ice mass loss. Determining which glacier(s) will control future changes in mass loss from such a spatially large region of the Antarctic Ice Sheet is key in steering scientific focus to the most dynamic sectors.

5. Conclusions

In this study, we quantified changes in mass balance of the most dynamic catchments in EAIS in response to a range of oceanic TF and SMB scenarios from both CMIP5 and CMIP6. We employ three different basal melting rate parameterizations to represent ice-ocean interactions in our ice sheet model. Differences in the projected SMB fields lead to a wide range of possible contributions of EAIS_{SUB} to global sea-level rise, with an average volume gain of approximately 10-mm SLRe by 2100. Projected increases in SMB dominate the mass balance signal to 2100, leading to low carbon emission scenarios losing more mass than high-emission scenarios. All basins within the domain either are in near mass balance or are projected to gain mass by

2100, except the ASB, where the majority of grounded ice thinning occurs within 50 km upstream of Totten Glacier's grounding line. The extent and timing of grounding line retreat along Totten's ice plain is sensitive to brief changes in ocean temperature. This ice plain acts as an important control of Totten's glacial discharge. In simulations where this ice plain ungrounds, we observe mean ice velocity increases in excess of 70% of the original velocity profile, compared to less than 20% in the projections for which this region does not unground. We find that this sector of Totten is not susceptible to MISI in its current state, as the grounding line readvances and velocity profile stabilizes when ocean temperatures are reduced to near present-day values. However, the inherent sensitivity of Totten's grounding line to short-lived changes in ocean temperature, its control on ice dynamics, and its significant sea-level rise potential render observing this glacier a future scientific priority. As approximately 40% of the global population lives within 100 km of the coastline, resolving plausible bounds on the future contribution of EAIS to global sea level within the coming century and discerning which outlet glaciers are most vulnerable are vital in preparing for the impacts of global climate change.

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