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Ocean-Ice Shelf Interaction in East Antarctica

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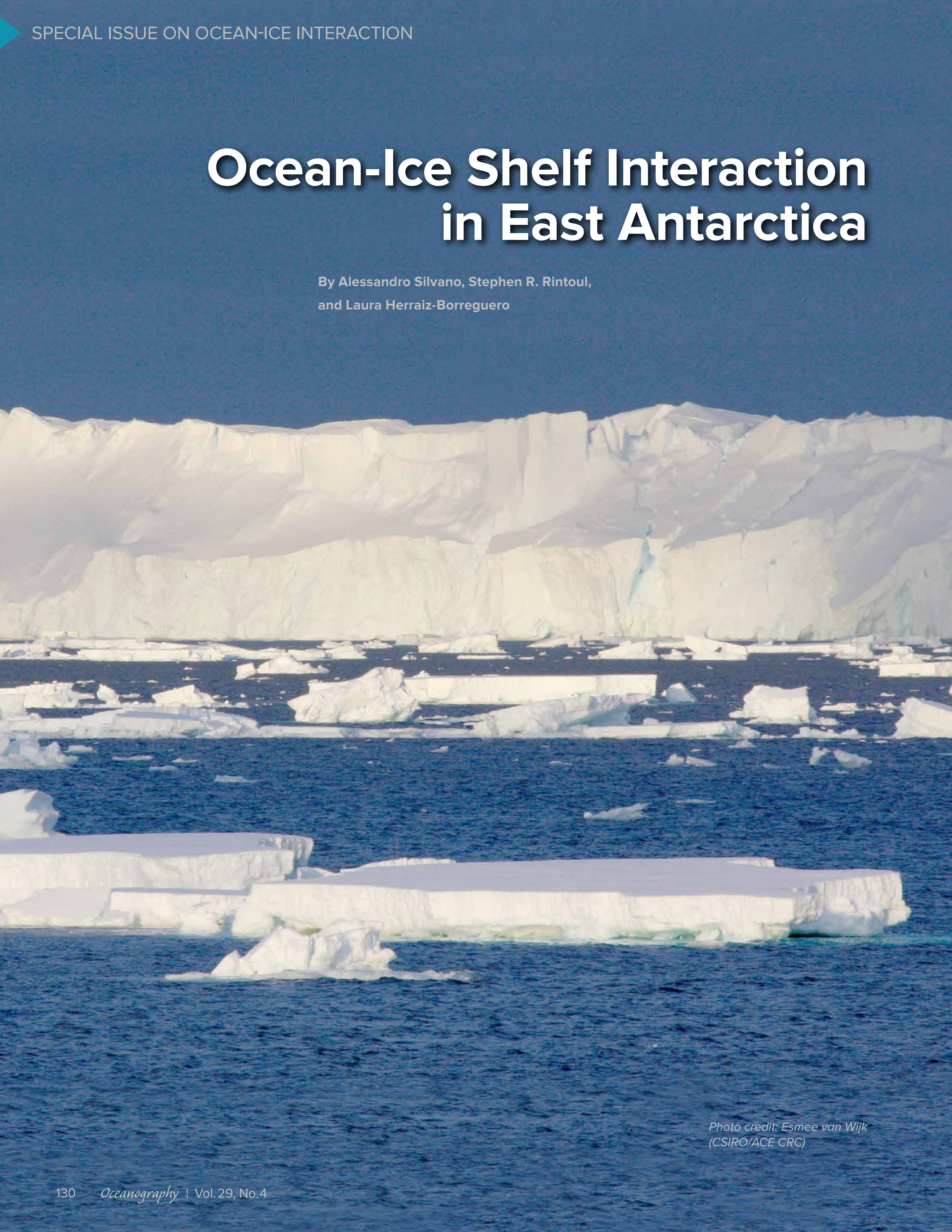

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> "Given the large volume of marine-based ice in East Antarctica, it is critical to develop a better understanding of the potential vulnerability of the [East Antarctic Ice Sheet] to changes in the ocean."

ABSTRACT. Assessments of the Antarctic contribution to future sea level rise have generally focused on ice loss in West Antarctica. This focus was motivated by glaciological and oceanographic observations that showed ocean warming was driving loss of ice mass from the West Antarctic Ice Sheet (WAIS). Paleoclimate studies confirmed that ice discharge from West Antarctica contributed several meters to sea level during past warm periods. On the other hand, the much larger East Antarctic Ice Sheet (EAIS) was generally considered to be relatively stable because of being largely grounded above sea level and therefore protected from ocean heat flux. However, recent studies suggest that a large part of the EAIS is grounded well below sea level and that the EAIS also retreated and contributed several meters to sea level rise during past warm periods. We use ocean observations from three ice shelf systems to illustrate the variety of ocean-ice shelf interactions taking place in East Antarctica and to discuss the potential vulnerability of East Antarctic ice shelves to ocean heat flux. The Amery and the Mertz are "cold cavity" ice shelves that exhibit relatively low area-averaged basal melt rates, although substantial melting and refreezing occurs beneath the large and deep Amery Ice Shelf. In contrast, new oceanographic measurements near the Totten Ice Shelf show that warm water enters the sub-ice-shelf cavity and drives rapid basal melting, as is seen in West Antarctica. Totten Glacier is of particular interest because it holds a marine-based ice volume equivalent to at least 3.5 m of global sea level rise, an amount comparable to the entire marine-based WAIS, and recent glaciological measurements show the grounded portion of Totten Glacier is thinning and the grounding line is retreating. Multiple lines of evidence support the hypothesis that parts of the EAIS are more dynamic than once thought. Given that the EAIS contains a volume of marine-based ice equivalent to 19 m of global sea level rise, the potential for ocean-driven melt to destabilize the marine-based ice sheet needs to be accounted for in assessments of future sea level rise.

BACKGROUND

Global mean sea level rose by 0.19 ± 0.02 m between 1901 and 2010 in response to global warming (Church et al., 2013). Ocean thermal expansion and ice loss from glaciers have been the dominant contributors to sea level rise over the past century (Church et al., 2013). The amount of ice stored in ice sheets, equivalent to ~7 m of global sea level rise for Greenland (Dowdeswell, 2006) and ~58 m for Antarctica (Fretwell et al., 2013), is vast compared to projected sea level contributions from glaciers (0.04–0.23 m) and thermal expansion (0.10–0.33 m) by the end of the twenty-first century (Church et al., 2013). Melt of even a small fraction of the water stored in the ice sheets would therefore have a substantial impact on future sea level rise, with widespread consequences for society, particularly in coastal regions. Satellite measurements show that the Greenland and Antarctic Ice Sheets have made positive and growing contributions to sea level rise in the last two decades (Rignot et al., 2011), and the ice sheets are expected to make the largest contribution in centuries to come (Rignot et al., 2011; Dutton et al., 2015). The response of the ice sheets to continued warming of the climate remains the largest single source of uncertainty in projections of long-term sea level rise.

Ice sheet stability depends on the balance between gains (from snowfall) and losses (from iceberg calving and melting). The presence of floating ice shelves around the margin of Antarctica contributes to the stability of the Antarctic Ice Sheet. Ice shelves form where ice streams flow off the continent into the ocean and start to float. Back stresses generated when a flowing ice shelf interacts with the seafloor or side walls restrains the drainage of glacial ice into the ocean and thus "buttresses" the ice sheet (Dupont and Alley, 2005). Thinning or collapse of an ice shelf reduces the buttressing effect and increases the discharge of ice into the ocean. Ice shelves thin when the inflow of ice from the continent is insufficient to balance the loss of ice to melt and iceberg calving. Melting from above by a warm atmosphere and melting from below by a warm ocean can both contribute to thinning of ice shelves. For most of Antarctica, air temperatures remain below freezing year-round at present (Picard and Fily, 2006), so basal melting by the ocean makes the dominant contribution. This means the future of Antarctic ice shelves, and the grounded ice sheets buttressed by the ice shelves, is strongly tied to the surrounding ocean. Ice shelves whose grounding lines (the boundaries between the floating ice and the grounded ice) are located well below sea level are potentially more sensitive to ocean forcing. Because the freezing temperature is lowered by pressure, the thermal forcing available for melting is higher for deeper grounding lines for a given water temperature. Moreover, ice sheets grounded on bedrock that slopes upward toward the sea are particularly vulnerable (Weertman, 1974; Schoof, 2007). In this geometry, a retreat of the grounding line increases the thickness at the ice front and therefore the ice discharge, producing a self-sustaining retreat that is hard to reverse (Joughin and Alley, 2011). This process is called marine ice sheet instability. Several studies suggest that marine ice sheet instability is already underway in West Antarctica (Favier et al., 2014; Joughin et al., 2014; Rignot et al., 2014). If this is the case, a threshold has been crossed that implies a commitment to meters of sea level rise from West Antarctica alone in centuries to come.

Observations from recent decades support the hypothesis that the ocean controls the stability of the Antarctic Ice Sheet. Antarctica as a whole is losing mass, with the largest losses in the Amundsen and Bellingshausen Sea sectors in West Antarctica (Rignot et al., 2008; Harig and Simons, 2015; Wouters et al., 2015). Here, the ice loss has been attributed to thinning of the buttressing ice shelves as a result of increased basal melting by ocean heat flux (Shepherd et al., 2004; Pritchard et al., 2012). Oceanographic evidence confirms that the most rapid mass loss, grounded ice thinning, and grounding line retreat have occurred where relatively warm ocean waters reach sub-ice-shelf cavities, driving rapid basal melting (e.g., Jenkins and Jacobs, 2008; Jacobs et al., 2011). The rate of ice loss and thinning of floating ice shelves has increased over the past two decades (Rignot et al., 2011; Paolo et al., 2015). Consistent with recent observations of dynamic behavior in response to ocean forcing, studies of past climate suggest the West Antarctic Ice Sheet has waxed and waned many times in the past (Scherer, 1991; Naish et al., 2009; Pollard and DeConto, 2009).

East Antarctica, on the other hand, has long been thought to be more stable. Most of the East Antarctic Ice Sheet (EAIS) was understood to be grounded on bedrock well above sea level, and the marine-based parts of the EAIS were believed to be isolated from warm Southern Ocean waters. But recent studies have overturned these assumptions. For example, inferences of past sea level from proxy data and models conclude that the EAIS retreated and made substantial contributions to sea level during past warm periods in Earth's history, suggesting the EAIS is more dynamic than previously thought (Williams et al., 2010b; Young et al., 2011; Cook et al., 2013; Patterson et al., 2014; Pollard et al., 2015; Aitken et al., 2016). New measurements from airborne sensors show that large sectors of the EAIS rest on bedrock well below sea level (Figure 1a; Ferraccioli et al., 2009; Young et al., 2011; Roberts et al., 2011; Jamieson et al., 2016). In fact, the marine-based EAIS hold a volume of ice equivalent to 19 m of global sea level rise, five times larger than the marine-based ice sheet in West Antarctica (Fretwell et al., 2013). While early studies indicated no significant ice loss in East Antarctica (e.g., Rignot, 2002), more recent glaciological observations show that some regions of the EAIS have lost mass in the last two decades. In particular, some aspects of recent change observed at Totten Glacier are similar to those seen in West Antarctica, including thinning of the grounded ice (e.g., Pritchard et al., 2009; Flament and Remy, 2012; Harig and Simons, 2015; Li et al., 2015) and retreat of the grounding line (Li et al., 2015). While the rates of change at Totten Glacier are not as large as those seen at Pine Island Glacier and other glaciers in West Antarctica, these observations come as a surprise, given that little warm water was thought to reach the continental shelf in this region, which lies well south of the core of the Antarctic Circumpolar Current (ACC). Until recently, no oceanographic data from the Totten ice front were available to test the hypothesis that ocean-ice shelf interaction could explain the dynamic behavior of this sector of East Antarctica.

Given the large volume of marine-based ice in East Antarctica, it is critical to develop a better understanding of the potential vulnerability of the EAIS to changes in the ocean. New modeling studies of the response of the Antarctic Ice Sheet to continued high emissions of greenhouse gases underscore the urgency of this task, showing that mass loss from Antarctica may be larger and more rapid than previously thought, with a large contribution from the EAIS (Golledge et al., 2015; DeConto and Pollard, 2016).

In this review, we summarize what is known about ocean-ice shelf interaction in East Antarctica. The region has received much less attention than West Antarctica, but recent studies have started to fill these gaps and provide new insights into the role of the ocean in driving changes in ice shelves. We focus on the three largest and deepest marine-based sectors of the EAIS: Lambert Graben drained by the Amery Ice Shelf, Aurora Basin mostly drained by Totten Glacier, and Wilkes Basin near the Mertz Glacier (Figure 1a).

OCEAN-ICE SHELF INTERACTION

Ice shelves form the interface between the Antarctic Ice Sheet and the surrounding Southern Ocean. Interaction between these two systems affects both the ice and the ocean. The ocean supplies heat to ice shelves, driving basal melting and influencing ice sheet dynamics, while freshwater input from ice shelf melting modifies ocean properties. The rate of ice shelf basal melting is determined by the ocean heat transport entering the ice shelf cavities and by the dependence on pressure of the seawater freezing temperature, which decreases ~0.75°C for every 1,000 m increase in depth (Foldvik and Kvinge, 1974). The heat transport depends on the temperature of the water (heat content) and on how fast ocean heat is transported to the base of the ice shelf. Thus, the rate of melting at the ice base depends on how fast heat and salt can be transported across the boundary layer to the ice-ocean interface. This, in turn, depends on other factors that are difficult to measure in the ice shelf cavity, such as ice shelf basal roughness, ocean circulation and tides, and turbulence (which, in its turn, depends on heat and salt gradients across the boundary layer). Direct measurements of the ocean beneath an ice shelf can only be made by drilling boreholes through the thick ice shelf or by sending underwater autonomous vehicles into the cavity. While such measurements have been made at a few ice shelves, we still have very few observations of the ocean circulation and heat transport beneath ice shelves. However, a strong correlation has been found between basal melting (inferred by satellite measurements) and ocean temperature measured at the front of the ice shelf (Rignot and Jacobs, 2002). For this reason, ocean temperature provides a good proxy for ice shelf basal melting. The highest basal melt rates occur near deep grounding lines due to the pressure dependence of the seawater freezing temperature. An increase of melt in this region will reduce the ice thickness and increase the gradient of ice thickness change near the grounding line, resulting in a reduction of buttressing to ice discharge and a rise in global sea level (Rignot and Jacobs, 2002). The deepest grounding lines in the world are located in East Antarctica, up to 2,100 m and 2,400 m deep for the Totten and Amery Ice Shelves, respectively (Fretwell et al., 2013).

How does the ocean melt an ice shelf? The schematic in Figure 2 illustrates three modes of basal melting. Mode 1 is typical of so-called "cold cavity" ice shelves. In winter, waters on the continental shelf are cooled to the surface freezing point. As the ocean freezes to form sea ice, salty brine is left behind in the water column. The decrease in temperature and increase in salinity both act to increase the density of the surface shelf water. The density may increase to the point where surface waters will sink to the seafloor, forming Dense Shelf Water (DSW; also called High Salinity Shelf Water). DSW is generally formed in coastal polynyas, where the combination of strong katabatic winds and a physical barrier that blocks inflow of sea ice (e.g., an ice tongue or grounded icebergs) acts to create an area

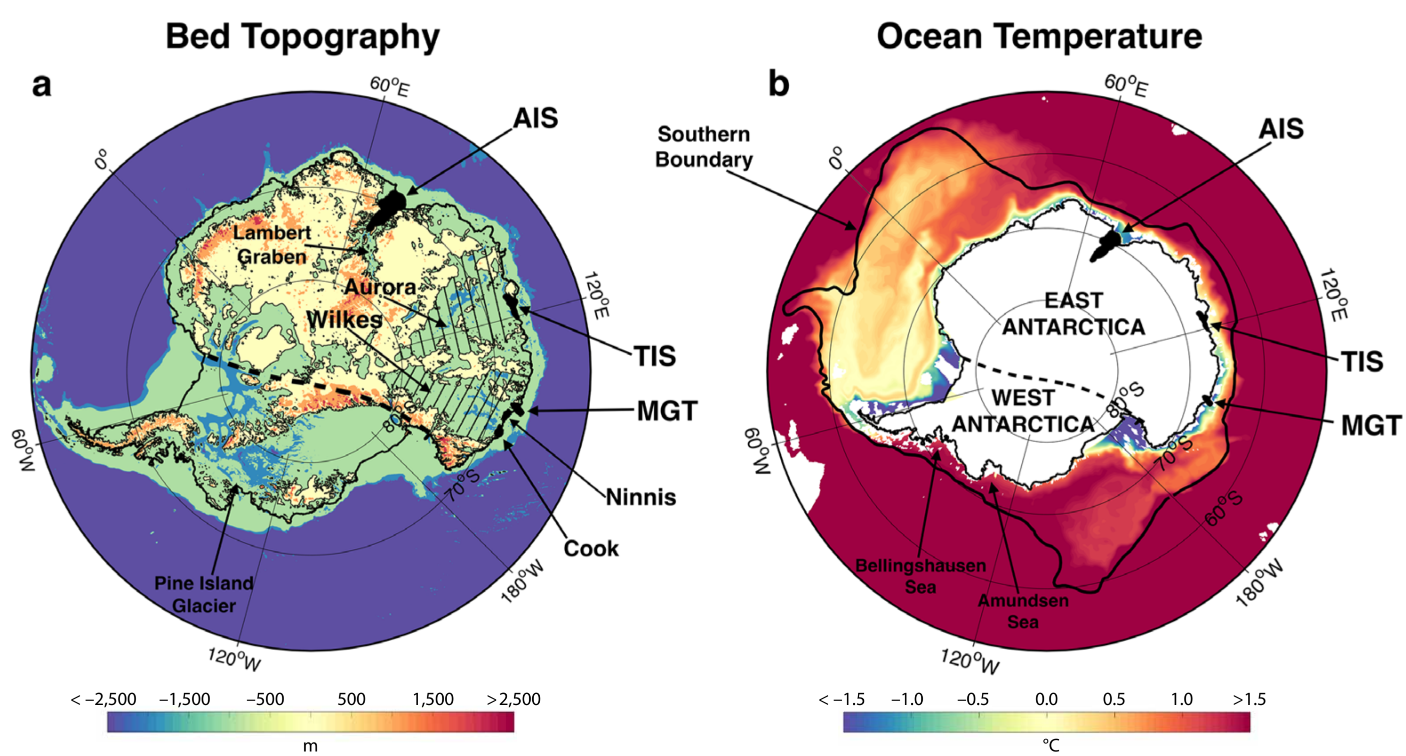


FIGURE 1. Map of Antarctica. (a) Bed topography with coastline (thick black line) and 0 m bathymetric contour (light black line) overlaid (Fretwell et al., 2013). Highlighted are the Amery Ice Shelf (AIS), Totten Ice Shelf (TIS), Mertz Glacier Tongue (MGT), and Ninnis and Cook Ice Shelves. In the Aurora and Wilkes Basins of East Antarctica (dashed areas) and in West Antarctica, the Antarctic Ice Sheet sits on bedrock as much as 2–3 km below sea level. (b) Ocean temperature at 438 m depth, close to the core of the Circumpolar Deep Water in the Antarctic Circumpolar Current, based on the six-year (2005–2010) mean from the Southern Ocean State Estimate (Mazloff et al., 2010). The heavy black line marks the southern boundary of the Antarctic Circumpolar Current (Orsi et al., 1995).

of low ice concentration and hence strong air-sea interaction. Polynyas are characterized by elevated sea ice production because sea ice is continuously produced and advected away by the wind. The continental shelf often gets deeper toward the grounding line of an ice shelf, allowing DSW to spread along the seafloor to the grounding line (Figure 2). Due to the pressure dependence of the freezing point, the DSW is warmer than the local freezing point and therefore melts the underside of the ice shelf. A buoyant mixture of DSW and fresh meltwater upwells and flows out of the cavity along the ice shelf base, sustaining an overturning circulation within the sub-ice-shelf cavity (Mode 1 in Figure 2). Where the temperature of the mixture of meltwater and DSW is less than the surface freezing point (i.e., "supercooled"), this water mass is called Ice Shelf Water (ISW). The formation of ISW by ocean-ice shelf interaction is generally a feature of cold cavity ice shelves. As the mixture of DSW and meltwater rises, it may reach a level where it is cooler than the in situ freezing point. This can result in formation of frazil ice crystals that may accrete to the base of the ice shelf, forming what is called marine ice. Marine ice layers stabilize the ice shelf, protecting it from fracturing and melting (Khazendar et al., 2009; Kulessa et al., 2014). Recently, marine ice has been identified as an important source of iron (Herraiz-Borreguero et al., 2016b), a limiting nutrient essential for phytoplankton growth. Icebergs with a visible marine ice layer are often known as jade icebergs due to their green color.

Mode 2 circulation dominates in "warm cavity" ice shelves (Figure 2). At these ice shelves, relatively warm Circumpolar Deep Water (CDW) spreads from offshore across the continental shelf to the sub-ice-shelf cavities. Where CDW mixes with surrounding waters and cools as it crosses the continental shelf, the resulting water mass is called modified CDW, or mCDW. CDW, which is the most voluminous and warmest mid-depth water mass found in the Southern Ocean (Worthington, 1981), is transported eastward around the continent in the ACC. The southern limit or "boundary" of this current (see Figure 1b) is located close to the West Antarctica coast in the Amundsen and Bellingshausen Seas (Orsi et al., 1995). In this sector, the southern edge of the ACC reaches the shelf break and allows warm (up to 1°C–1.5°C) and relatively salty (34.5–34.7) CDW to reach the continental shelf (Figure 1b). DSW is not formed in the Amundsen and Bellingshausen Seas, so the CDW fills the bottom layer and can reach the deepest regions of the ice shelf cavities and drive rapid basal melting. The most prominent example of a "warm cavity" ice shelf is Pine Island Glacier in West Antarctica, where inflow of CDW reaches the grounding line with temperatures in excess of 3°C above the local freezing point (Jenkins et al., 2010). In the Weddell and Ross Sea sectors, large clockwise gyres of relatively cold water isolate the continental margin from the warm waters of the ACC (Figure 1b), and the ice shelf cavities there are filled with DSW. East Antarctica was long thought to be similarly isolated from warm offshore waters because the axis of the ACC was far to the north. However, the southern boundary of the ACC approaches the East Antarctic coast between 60°E and 140°E (Figure 1b), providing a reservoir of ocean heat near the continental shelf break. Due to a lack of observations, it was not known whether warm offshore waters crossed the continental shelf and reached the ice shelf cavities in these locations. As discussed in the next section, recent measurements show that in some locations, relatively warm water does cross the shelf break to drive rapid basal melting, demonstrating that East Antarctica is more vulnerable to changes in ocean heat fluxes than previously thought.

Atmospheric heating during summer months leads to warming of the surface waters. Mode 3 is related to the transport of warm surface waters to the calving fronts of ice shelves by tides, eddies, Ekman transport, and ocean currents (Figure 2). There is usually less melt at an ice front than near the grounding line due to the suppression of the freezing point with depth. However, in some cases where the grounding line is relatively shallow, basal melt rates driven by warm surface waters can rival those observed at the grounding line. Examples where Mode 3 plays a primary role are the Ronne-Filchner and

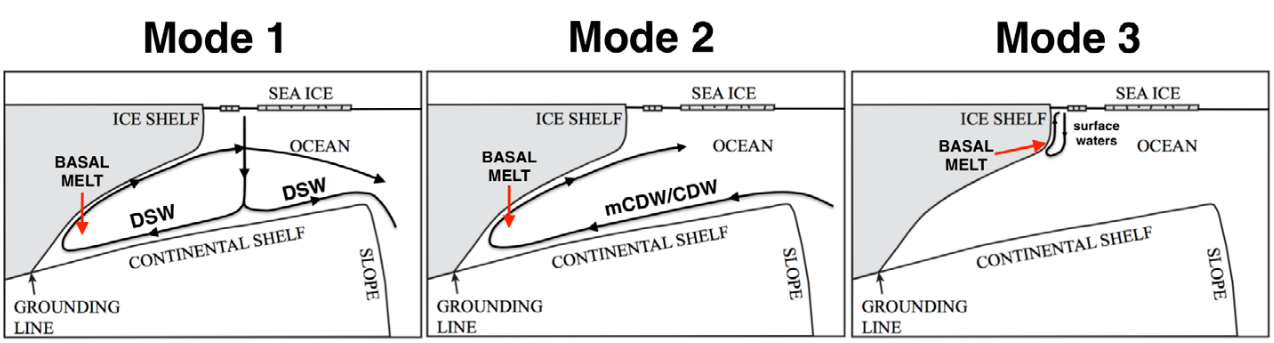


FIGURE 2. Three modes of ice-shelf basal melting (from Jacobs et al., 1992). Mode 1 is driven by cold Dense Shelf Water (DSW), Mode 2 by warm (modified) Circumpolar Deep Water (mCDW/CDW), and Mode 3 by surface waters. The outflow from the cavity is a mixture of glacial meltwater with DSW (Mode 1) or mCDW (Mode 2); the mixture is called Ice Shelf Water (ISW) when its temperature is below the surface freezing point.

Fimbul Ice Shelves in the Weddell Sea (Makinson and Nicholls, 1999; Joughin and Padman, 2003; Hatterman et al., 2012) and the Ross and McMurdo Ice Shelves in the Ross Sea (Arzeno et al., 2014; Stern et al., 2013).

OCEAN-ICE SHELF INTERACTION IN EAST ANTARCTICA

The ice shelves most susceptible to basal melting, and therefore the systems with the greatest potential to contribute to sea level rise, are those with deep grounding lines. Here, we focus on the three outlet glaciers and related ice shelves at the margins of the large marine-based sectors of the EAIS (Figure 1a). Totten Glacier drains more ice from East Antarctica than any other glacier and forms the Totten Ice Shelf (TIS) where it reaches the sea. The Amery Ice Shelf (AIS) is fed by several glaciers, including Lambert Glacier, the largest glacier in the world, which drains about 16% of the area of East Antarctica (Allison, 1979). Mertz Glacier is near Wilkes Basin, the largest marine-based drainage system in East Antarctica, and forms the Mertz Glacier Tongue (MGT) where it starts to float. Ice shelf systems all have different geographic and oceanographic settings; the three examples below illustrate the variety of ocean-ice shelf interactions in East Antarctica.

Totten Ice Shelf

Totten Glacier on Sabrina Coast is the major outlet of the Aurora Basin and drains a marine-based ice volume equivalent to 3.5 m of global sea level rise (Greenbaum et al., 2015). Strong retreat and mass loss of Totten Glacier made substantial contributions to global sea level change during past warm periods in Earth's history such as the Pliocene, when global temperatures were comparable to those predicted for the end of this century and atmospheric CO_2 concentrations were similar to those of the present (Williams et al., 2010b; Pollard et al., 2015; Aitken et al., 2016). Future climate projections under continued high anthropogenic greenhouse emissions indicate mass loss from Aurora Basin will contribute several meters to sea level rise in the coming centuries (Golledge et al., 2015; DeConto and Pollard, 2016). Recent glaciological observations reveal that during the last two decades, Totten Glacier has been retreating, thinning, and losing mass (Li et al., 2015, 2016). The basal melt rate at Totten Ice Shelf is 10–18 m yr^{-1}, the highest among the major ice shelves

> "Estimates of future sea level rise need to take the potential contribution from a dynamic [East Antarctic Ice Sheet] into account."

in East Antarctica and only exceeded by Pine Island and neighboring ice shelves in the Amundsen–Bellingshausen Seas (Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015). However, no oceanographic measurements had been made near the front of TIS so it was not known whether ocean-ice shelf interaction was responsible for the recent changes in the glacier and ice shelf. Temperature profiles collected in 1996 and 2007 revealed the presence of relatively warm (0°C–0.5°C) mCDW near the seafloor on the outer continental shelf (Bindoff et al., 2000; Williams et al., 2011), but until recently, heavy sea ice had prohibited access to the calving front to confirm that warm water reached the ice shelf cavity.

The Totten Ice Shelf calving front was finally reached in January 2015 by RSV *Aurora Australis*. Offshore winds opened a narrow and short-lived coastal lead that allowed the ship to occupy a section along the calving front of the glacier (Figure 3a). Measurements of water depth, temperature, salinity, and velocity were collected. A narrow 1,100 m deep trough was found at the ice front, almost twice as deep as suggested by earlier measurements (Fretwell et al., 2013; Greenbaum et al., 2015). A thick layer of relatively warm mCDW (–0.4°C) was present at the bottom of the trough (Figure 3b; Rintoul et al., 2016; recent work of author Silvano and colleagues). If the warmest water observed at the calving front were to reach the grounding line at 2,100 m depth, it would be more than 3°C above the local freezing point, approaching that observed near Pine Island Glacier (Jenkins et al., 2010).

Velocity measurements confirm that the warm water was flowing strongly into the cavity at the time of the section (Rintoul et al., 2016). The ocean heat flux into the cavity was consistent with that required to support the large multiyear mean basal melt rates inferred from independent glaciological data (Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015), supporting the hypothesis that ocean heat flux drives rapid basal melting of Totten Glacier and suggesting the snapshot obtained during the voyage was broadly representative. Water properties observed at the ice shelf calving front provide further evidence of active basal melting. Fresher water in the upper 500 m on the western side of the calving front (Figure 3c) reflects the outflow of a buoyant mixture of mCDW and glacial meltwater. Using the approach of Jenkins

(1999), author Silvano and colleagues recently used oxygen and salinity data to show that the water flowing out of the cavity contains up to 5 ml l^{-1} of glacial meltwater, about one-third of that observed at the front of Pine Island Glacier.

Oceanographic observations show Totten Glacier is a "warm cavity" ice shelf, similar to the rapidly melting ice shelves in the Bellingshausen and Amundsen Seas in West Antarctica (Figure 3d). No DSW has been observed on the Totten continental shelf in summer or winter (Williams et al., 2011), consistent with observations of weak sea ice formation in Dalton Polynya east of the TIS (see Figure 3a; Tamura et al., 2016). As a consequence, warm mCDW, the densest water on the shelf, has access to the ice shelf cavity, as found in the Amundsen and Bellingshausen Seas, although waters in the bottom layer are about 1°C cooler near Totten Glacier compared, for example, to Pine Island Glacier. The water properties confirm substantial meltwater in the outflow from the Totten cavity. As large basal melt rates have been observed since at least the early 2000s (Rignot and Jacobs, 2002; Depoorter et al., 2013; Liu et al., 2015; Rignot et al., 2013), it is likely that mCDW has been entering the cavity for at least this time period.

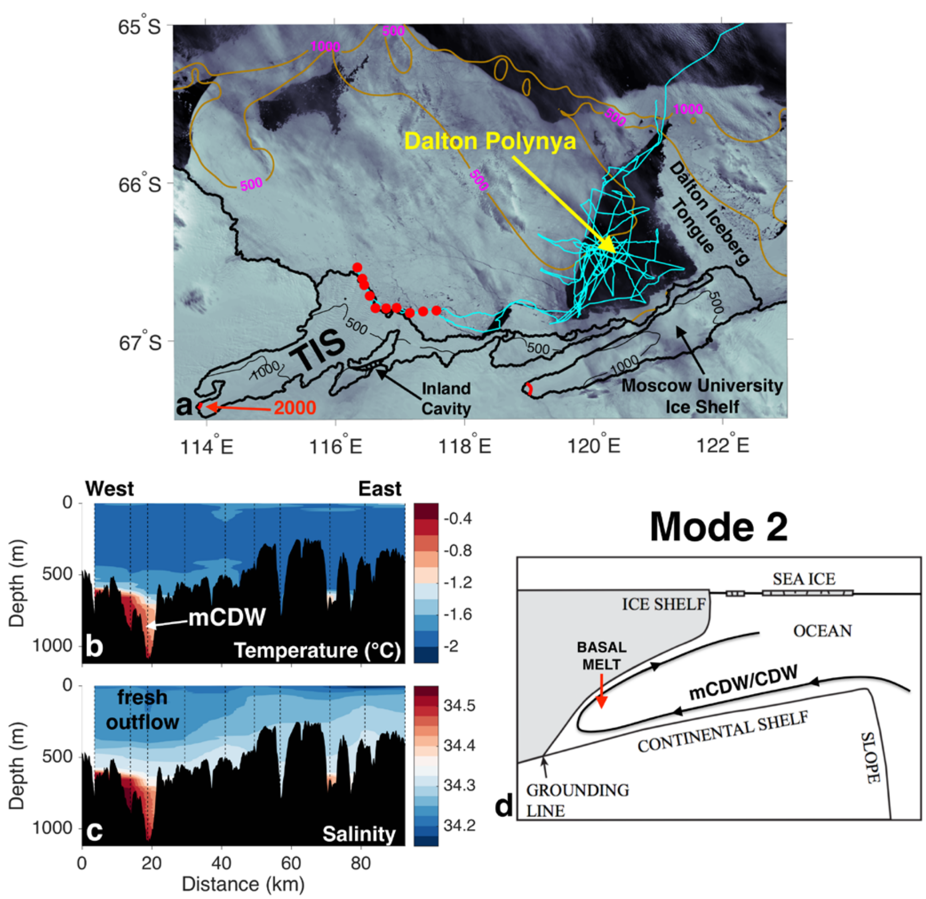


FIGURE 3. Totten Ice Shelf. (a) Satellite image of Totten Ice Shelf (TIS) and surroundings on January 22, 2015 (Scambos et al., 1996a). The bathymetry (Arndt et al., 2013) and the coastline (Fretwell et al., 2013) are shown by brown and black lines, respectively. The red dots mark the locations of the stations undertaken at the Totten calving front in summer 2015. The small dashed area indicates an inland cavity that connects the main trunk of TIS with the adjacent fringing ice shelf (Greenbaum et al., 2015). East of TIS is the Moscow University Ice Shelf where no data are available at the calving front. A series of grounded icebergs and fast ice (Dalton Iceberg Tongue) blocks inflow of sea ice from the east driven by westward coastal currents and allows the formation of the Dalton Polynya, east of TIS. (b) Vertical section of in situ temperature (°C) measurements along the calving front in 2015 (locations of stations shown by red dots in Figure 3a). (c) As in (b) for salinity. (d) Schematic showing that basal melting of TIS is dominated by Mode 2, where warm modified Circumpolar Deep Water (mCDW) enters the sub-ice-shelf cavity to drive relatively rapid melt.

Mertz Glacier Tongue

Wilkes Basin contains a volume of marine-based ice equivalent to 3–4 m of global sea level rise (Mengel and Levermann, 2014). Like Totten Glacier in Aurora Basin, the Wilkes Basin ice sheet has retreated and contributed to sea level rise during past warm periods (Williams et al., 2010b; Cook et al., 2013; Patterson et al., 2014). Also like Totten, models suggest that continued high emissions of greenhouse gases could drive mass loss from Wilkes Basin, resulting in several meters of sea level rise in coming centuries (Mengel and Levermann, 2014; Golledge et al., 2015; DeConto and Pollard, 2016). The two main ice streams draining the Wilkes Basin, Cook and Ninnis Glaciers (see Figure 1a for the location of the Cook and Ninnis Ice Shelves), appear to be largely in balance at the present time based on satellite measurements of gravity and surface elevation (e.g., Velicogna and Wahr, 2013). The potential for the ocean to destabilize the ice sheet in this sector remains poorly understood. Both Cook and Ninnis Glaciers are almost inaccessible due to heavy sea ice conditions, and no measurements have been made near the calving fronts of these glaciers. The Wilkes Land Expedition on USCGC *Polar Star* in 1985 showed that water as warm as –1.2°C crossed the shelf break between 150°E and 155°E, but the warm intrusions did not extend more than halfway across the continental shelf (Foster, 1995). Measurements made by the Russian R/V *Ob* in the 1950s about 120 km north of the Ninnis ice front showed no evidence of warm mCDW (Gordon and Tchernia, 1972).

Mertz Glacier lies about 100 km west of Ninnis Glacier. While Mertz Glacier does not drain the main Wilkes Basin, it is well studied in comparison to Ninnis and Cook Glaciers, and its proximity makes it a useful analogue for the oceanographic conditions typical of the area's coast. Prior to a major calving event in 2010, the Mertz Glacier Tongue extended about 120 km northward from the coast (Figure 4a, cyan line) and blocked the general east-to-west movement of sea ice along the continental shelf. The combination of strong katabatic winds and restricted inflow of sea ice resulted in formation of a strong and active polynya to the west of the MGT as well as others in coastal embayments further west. Over the period 1992–2013, Mertz Polynya was the fourth most active Antarctic polynya in terms of sea ice production (Tamura et al., 2016). Formation and export of large volumes of sea ice results in the production of DSW, which flows down the slope and mixes with surrounding waters to form the local variety of Antarctic Bottom Water (Gordon and Tchernia, 1972; Rintoul, 1998). Calving of the glacier tongue has reduced the polynya's sea ice production and hence the density of shelf water and bottom water formed in the region (Shadwick et al., 2013; Nihashi and Ohshima, 2015).

Active formation of DSW in Mertz Polynya has a significant impact on ocean-ice shelf interaction, as well as on bottom water formation. Figure 4b shows temperature along the MGT front in January 2015 (i.e., post-calving). Mertz is primarily a cold cavity ice shelf, like those found at other locations around the Antarctic margin where polynyas produce large volumes of DSW. The DSW, with relatively high salinity and temperature close to the surface freezing point, enters the cavity and flows downslope to the grounding line (Figure 4c). At the 1,200 m deep grounding line of Mertz Glacier, water at the surface freezing point is about 0.9°C warmer than the local freezing point and drives basal melting. The mixture of inflowing DSW and glacial meltwater forms Ice Shelf Water (ISW), with the coldest and strongest outflow of ISW found on the western side of the ice front (Figure 4b). While the presence of ISW with temperatures cooler than −2°C confirms that glacial melting occurs at depth and the meltwater exits the cavity, the relatively weak thermal forcing is consistent with low estimates of basal melting (1–2 m yr^{-1}) inferred from glaciological measurements (Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015). No warm mCDW was observed at the ice front during the 2015 voyage (Figure 4b). However, mCDW has been observed over the continental shelf in this region, with highly modified CDW extending as far south as the offshore end of the MGT prior to calving (Rintoul, 1998; Williams et al., 2010a; Lacarra et al., 2011; Snow et al., 2016). The mCDW reaching the continental shelf is lighter than the DSW and is found at mid-depth there, in contrast to the Totten continental shelf where mCDW is the densest and deepest water mass present. In this way, Mertz Polynya conditions the ocean stratification and the interaction of the ocean with the ice shelf, as also seen in modeling studies (Cougnon et al., 2013). Calving of the MGT has reduced sea ice formation and DSW formation in Mertz Polynya, and may therefore have altered basal melt rates, but no studies have yet been published quantifying the impact of the calving event on ocean-ice shelf interaction.

Amery Ice Shelf

The Lambert Glacier system is the major outlet of the MacRobertson Land sector, draining 16% of the area of East Antarctica (Fricker et al., 2000). It contains an ice volume equivalent to 4–6 m of sea level. Lambert Glacier and several tributary glaciers supply the Amery Ice

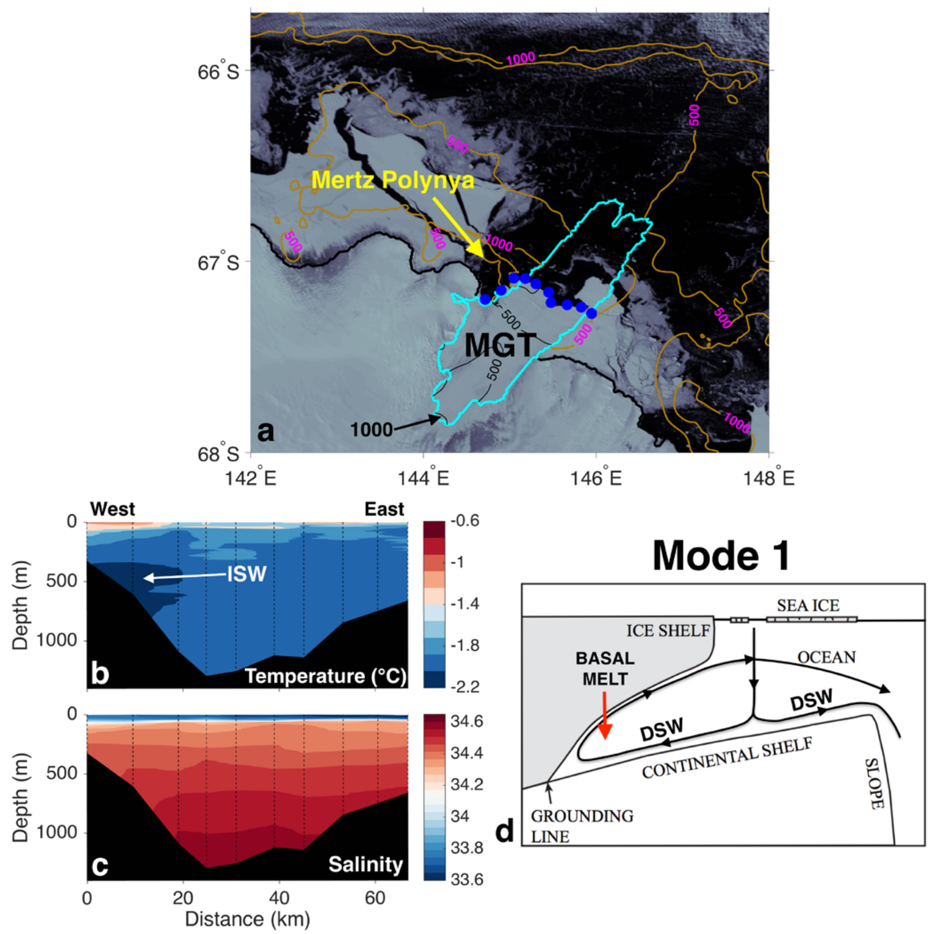


FIGURE 4. Mertz Glacier Tongue (MGT). (a) Satellite image of the MGT and surroundings on December 2, 2014 (Scambos et al., 1996b). The bathymetry (Arndt et al., 2013) and coastline (Fretwell et al., 2013) are shown by brown and black lines, respectively. The blue dots represent the locations of stations undertaken along the calving front in summer 2015, and the cyan line indicates the ice shelf boundary prior to a major calving event in 2010. (b) Vertical section of in situ temperature (°C) recorded at the calving front in 2015 (blue dots in Figure 4a). ISW = Ice Shelf Water. (c) As in (b) for salinity. (d) Schematic showing that basal melting of the MGT is dominated by Mode 1, where cold Dense Shelf Water (DSW) formed in the Mertz Polynya flows into the sub-ice-shelf cavity and drives melt at the grounding line.

Shelf, the third largest embayed ice shelf in Antarctica, with an area of 62,000 km^2. The deepest part of the southern grounding line is at 2,500 m (Fricker et al., 2001). Craven et al. (2009) estimated that about 80% of the continental ice entering the AIS is lost to basal melting (~43.2 Gt yr^{-1}). A layer of marine ice up to 190 m thick on the northwestern side of the AIS base constitutes as much as 9% of the ice shelf volume (Fricker et al., 2001) and is thought to stabilize the AIS (Khazendar et al., 2009; Kulessa et al., 2014). Like Totten Glacier and the glaciers draining Wilkes Basin, evidence from paleoclimate studies (Passchier, 2011) and simulations of the future response of the ice sheet (Golledge et al., 2015; DeConto and Pollard, 2016) suggest the Amery Ice Shelf is also dynamic.

The Amery Ice Shelf-Ocean Research (AMISOR) project measured ocean properties beneath the ice shelf continuously from 2001 to 2012. Outside the ice shelf cavity, a few oceanographic expeditions were conducted in the 1980s and in 2001 and 2002. However, not until the appearance of "seal oceanographers" in 2010 was it possible to draw a picture of the spatial and seasonal variability of Prydz Bay water masses (Herraiz-Borreguero et al., 2015, 2016a; Williams et al., 2016).

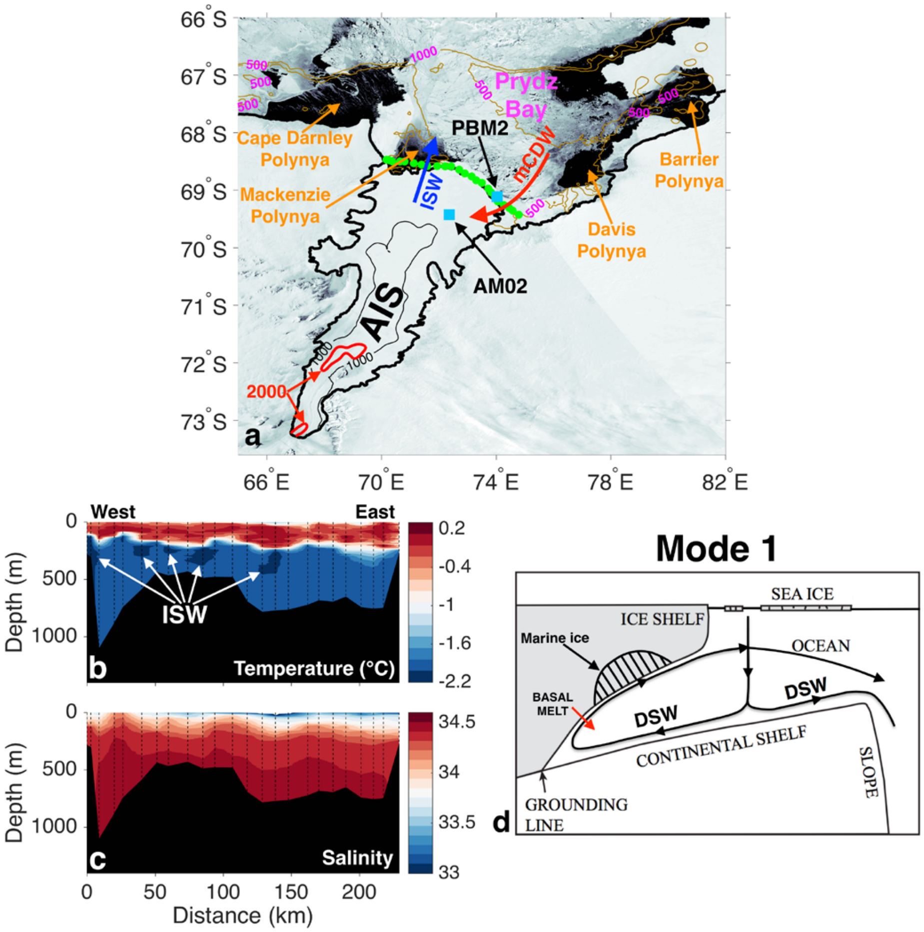


FIGURE 5. Amery Ice Shelf. (a) Satellite image (NASA Worldview data, NASA EOSDIS, https://earthdata.nasa.gov/labs/worldview) of the Amery Ice Shelf (AIS) and surroundings on October 30, 2014, when the Cape Darnley, Mackenzie, Davis, and Barrier Polynyas were clearly visible. Red and blue arrows depict the inflow of mCDW in winter in the east and the outflow of ISW in the west. The bathymetry (Arndt et al., 2013) and coastline (Fretwell et al., 2013) are shown in brown and black lines, respectively. The light blue squares show the location of one borehole site, AM01, and one Australian mooring, PBM2, deployed 2001–2002. The green dots represent the locations of stations undertaken along the calving front in summer 2001. (b) Vertical section of in situ temperature (°C) recorded at the calving front in 2001 (green dots in Figure 5a). (c) As (b) for salinity. (d) Schematic showing that Mode 1 dominates basal melting at the AIS, driven by cold DSW formed in the Mackenzie, Davis, and Barrier polynyas.

Both DSW and mCDW drive basal melting beneath the AIS (Herraiz-Borreguero et al., 2015). Cold and salty DSW occupies the deepest part of the water column in the bay and near the AIS front (Figure 5b,c). This DSW is created during active sea ice formation at three polynyas—Barrier, Davis, and Mackenzie (Williams et al., 2016; Figure 5a)—and enters the AIS cavity (Herraiz-Borreguero et al., 2013, 2016a). When DSW reaches the deep grounding line where it is ~2°C above the in situ freezing point, it drives rapid basal melting (up to 25 m yr^{-1} near the grounding line; Wen et al., 2010). The total basal melt of the AIS is 33.5 ± 24.4 Gt yr^{-1} (area-averaged basal melt rate of 0.6 ± 0.4 m yr^{-1}). ISW formed by mixing of glacial meltwater with DSW rises along the base of the ice shelf and freezes to form frazil ice crystals that accrete to the ice shelf base to form the thick layer of marine ice found on the western side of the cavity (Fricker et al., 2001), with most the marine ice formed in winter (Herraiz-Borreguero et al., 2013). The lower part of the marine ice layer is hydrologically connected to the ocean below (Craven et al., 2009), and so potentially subject to rapid melting. The circulation in the AIS cavity is dominated by this Mode 1 circulation, with inflow of DSW and outflow of ISW (Figure 5d).

However, inflow of mCDW on the eastern side of the AIS cavity drives basal melting during austral winter (Herraiz-Borreguero et al., 2015). CDW upwells into Prydz Bay where it cools very rapidly and deepens on its way to the AIS (Herraiz-Borreguero et al., 2015). MCDW occupies the eastern flank of Prydz Bay during summer (Herraiz-Borreguero et al., 2015; Williams et al., 2016) and is not observed at the ice-shelf front until April (Figure 6a; Herraiz-Borreguero et al., 2015, 2016a). Very cold mCDW (−1.7°C) is observed at borehole AM02 (~80 km from the ice shelf front) in mid-winter (Figure 6a), causing up to 2 ± 0.5 m yr^{-1} (23.9 ± 6.5 Gt yr^{-1}) of basal melting in a limited area of the northeastern flank of the ice shelf base

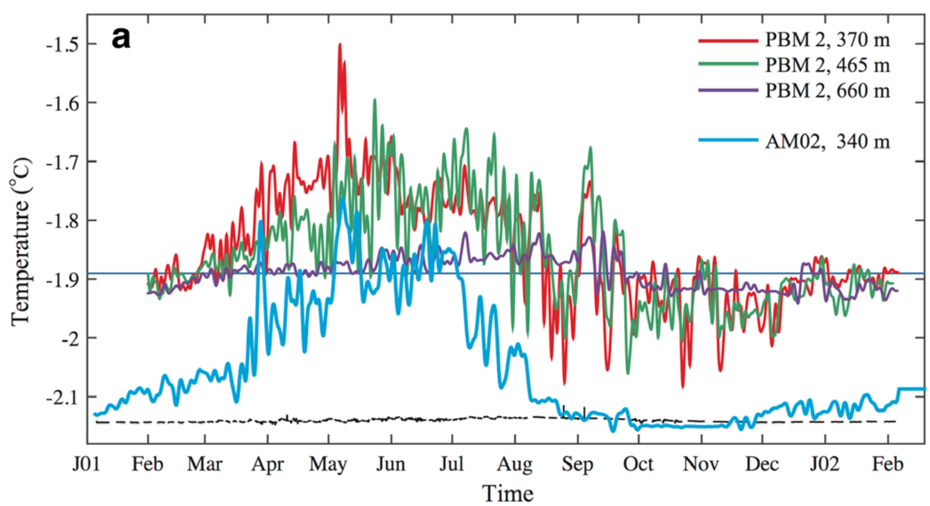
(Herraiz-Borreguero et al., 2015). The interaction of the mCDW with the AIS also results in the formation of a distinct variety of ISW. This "warm" circulation is characteristic of Mode 2 and occurs in conjunction with Mode 1 circulation (Figure 6b).

Interaction of the ocean with the Amery Ice Shelf likely influences formation of Antarctic Bottom Water (AABW). Recent studies show how ISW controls the formation rate and thermohaline properties of DSW in Prydz Bay (Herraiz-Borreguero et al., 2016a; Williams et al., 2016). Prydz Bay DSW has, in turn, been linked to the formation and interannual variability of AABW formed at Cape Darnley, a persistent polynya immediately west of Prydz Bay (Figure 5a; Couldrey et al., 2013; Ohshima et al., 2013; Williams et al., 2016).

VARIETIES OF OCEAN-ICE SHELF INTERACTION IN EAST ANTARCTICA

The three ice shelves discussed above illustrate the different mechanisms driving basal melting in East Antarctica. The basal melt rate is strongly influenced by the water properties and stratification near the front of the ice shelf, which in turn are influenced by a number of processes, including sea ice formation and melt, coastal circulation, winds, intrusions of offshore waters, and freshwater input from ice shelf basal melt. The water properties and the stratification where water flows into the cavities of the Amery, Totten, and Mertz Ice Shelves are shown in Figure 7; a profile from the Pine Island Glacier calving front is included to provide a comparison to West Antarctica. Amery and Mertz Ice Shelves are similar in many ways. Both are primarily cold cavity ice shelves dominated by Mode 1 circulation. Active polynyas produce large volumes of DSW, which enters their cavities to drive basal melting. The mixture of cold DSW with glacial meltwater at depth produces ISW. More importantly for ocean-ice shelf interaction, the presence of DSW restricts intrusions of warmer and lighter mCDW to mid-depth and prevents the warm water from reaching the grounding lines of Amery and Mertz Ice Shelves. The inflow of mCDW sets up a secondary Mode 2 circulation during winter, and it makes a relatively small contribution to basal melting of the Amery Ice Shelf when compared to Totten and Pine Island Glaciers. There are also some clear differences between the Amery and the Mertz. The grounding line of Amery Ice Shelf is much deeper than that of Mertz, resulting in stronger thermal forcing and melt at the grounding line. Marine ice is widespread on the northwestern underside of the large Amery Ice Shelf. Mertz Polynya is more active than the polynyas adjacent to Amery (Tamura et al., 2016), and ISW is less widespread at Mertz. As a result, the Mertz region produces saltier (Figure 7b) and denser shelf water that contributes directly to AABW formation, while the fresher and lighter DSW from the Amery region contributes to AABW formation at Cape Darnley Polynya.

The water properties and stratification at Totten Glacier are very different from those observed at Amery and Mertz (Figure 7). The relatively weak Dalton Polynya does not produce enough brine to convert the fresh surface waters on the Totten continental shelf into DSW. In the absence of DSW, mCDW occupies the deepest part of the water column and has access to the Totten Ice Shelf cavity. The warm cavity Totten Ice Shelf therefore experiences high basal melt rates driven by the inflow of mCDW (i.e., Mode 2). The outflow from the Totten cavity is relatively warm and low in oxygen compared to the outflow from Amery and Mertz, reflecting the mixture of glacial meltwater with warm mCDW rather than cold DSW (recent work of author Silvano and colleagues). These characteristics are similar to, but not as extreme as,



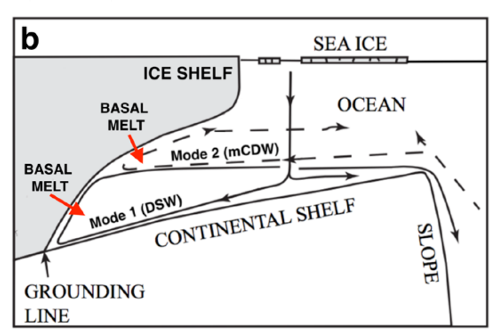


FIGURE 6. Temporal variability at the Amery Ice Shelf. (a) In situ temperature (°C) between January 2001 and February 2002 at the eastern calving front of the Amery Ice Shelf (PBM2) and in the eastern sub-ice-shelf cavity (AM02). See Figure 5a for the locations of PBM2 and AM02. The 40 h Butterworth filtered time series are shown. From April to August, relatively warm mCDW (green and red lines) reaches the eastern Amery Ice Shelf to drive basal melting at mid-depths. Cold DSW is found year-round in the bottom layer (purple line) and drives basal melting near the grounding line. (b) Schematic showing that Mode 1 (cold DSW) dominates basal melting at the Amery Ice Shelf near the grounding line, while Mode 2 contributes to basal melting at mid-depths when relatively warm mCDW enters the eastern cavity in winter.

those observed at the front of Pine Island Glacier, where warm ocean waters reach the grounding line and drive rapid melt (Jenkins et al., 2010).

Comparison of the four ice shelves illustrates the strong influence of stratification on ocean-ice shelf interaction. The stratification, in turn, is strongly influenced by polynya dynamics. Where active polynyas produce DSW, access of warm water to the ice shelf cavity is restricted and area-averaged basal melt rates are low. Where DSW is not formed, because sea ice production is low and/or stratification is strong, mCDW that crosses the continental shelf can access the ice shelf cavity and area-averaged basal melt rates are high. Ocean temperatures are, of course, not the only factors affecting basal melt rates. For a given ocean temperature, melting will be more rapid at deeper grounding lines due to the pressure dependence of the freezing point. The melt rate also depends on a number of other factors, including cavity geometry and the surface characteristics of the underside of the ice shelf. Ocean heat flux to ice shelf cavities, and hence basal melting, can also vary in time in response to atmospheric variability (Dutrieux et al., 2014) and changes in the shape of the ice shelf cavity (Jacobs et al., 2011). Though most oceanographic observations have been made in summer, the winter data that are available from Amery Ice Shelf show that seasonal evolution of water properties and circulation can influence ocean-ice shelf interaction and basal melt rates (Herraiz-Borreguero et al., 2015), highlighting the need for year-round observations near ice-shelf calving fronts.

The similarities and differences among ice shelves can also be illustrated by comparing total and area-averaged basal melt rates inferred from glaciological measurements (Table 1). The warm cavity ice shelves in West Antarctica have high basal melt rates, both in total and on an area-averaged basis. (These ice shelves also show the largest acceleration in the rate of volume loss [Paolo et al., 2015]). The cold cavity ice shelves in East Antarctica (e.g., Amery and Mertz) have relatively low area-averaged melt rates. Totten, in contrast, has the largest area-averaged basal melt rate of all East Antarctic ice shelves (for ice shelves larger than 1,000 km^2). The nearby Moscow University Ice Shelf also has a relatively large area-averaged basal melt rate compared to other East Antarctic ice shelves. Amery and Shackleton have relatively low rates of melt on an area-averaged basis, but lose substantial mass to basal melting. The Shackleton Ice Shelf system is important because it drains Denman Glacier, and deep connections between Denman Glacier and Aurora Basin make this glacier potentially susceptible to marine ice sheet instability (Roberts et al., 2011). However, lack of oceanographic data prevents an assessment of the susceptibility of this part of the EAIS to ocean-driven ice loss.

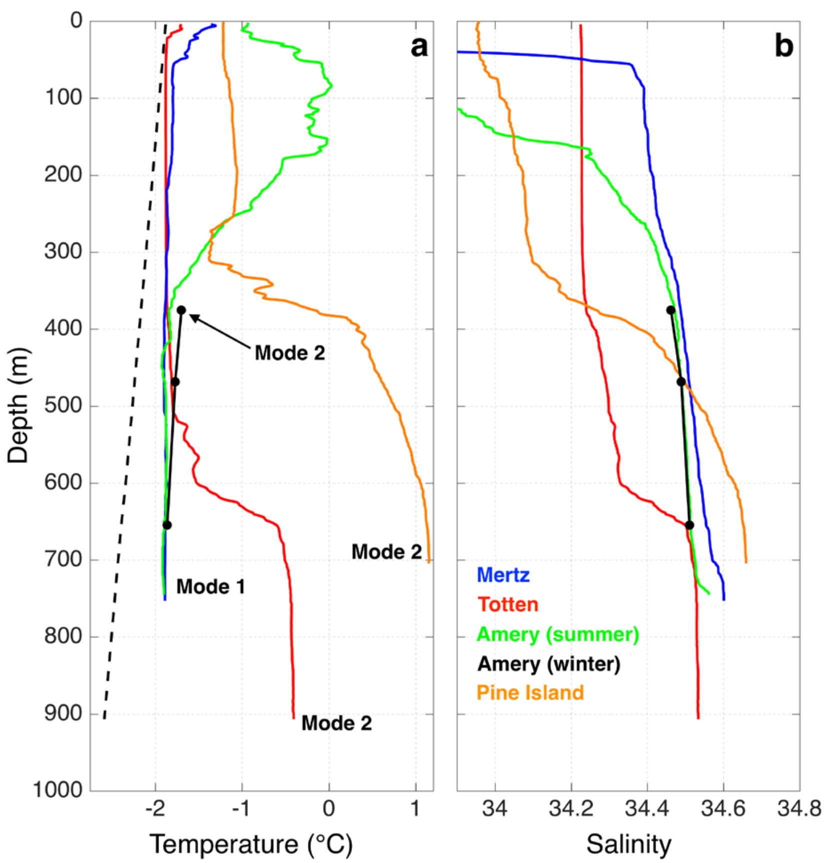


FIGURE 7. Water properties of the strongest inflow to ice-shelf cavities in East and West Antarctica. (a) In situ temperature (°C) versus depth (m) of individual profiles from Figures 3, 4, and 5 representative of summer ocean stratification near the calving fronts of Totten Glacier (red), Mertz Glacier (blue), and Amery Ice Shelf (green). Mean winter temperature in 2001 measured by a mooring at the Amery ice front (PBM2; see Figure 5a) is shown by the black line (instruments at 370, 465, and 660 m depths). For comparison, a summer profile collected in 2009 near Pine Island Glacier is included in orange (Jacobs et al., 2011). The black dotted line is the local freezing temperature, which decreases with depth. (b) As in (a) for salinity.

CONCLUSIONS

Growing evidence from studies of past sea level, oceanographic and glaciological observations, and simulations of ice sheet responses to high emissions of greenhouse gases suggest the East Antarctic Ice Sheet is more dynamic than once thought. In West Antarctica, there is evidence that ocean-driven basal melting is the primary control on mass loss from the floating ice shelves and the grounded ice sheets they buttress (Pritchard et al., 2012). The lack of oceanographic observations in East Antarctica has prevented an assessment of the extent to which ocean heat flux

TABLE 1. Basal melt rates at selected West and East Antarctic ice shelves. Total basal melt (Gt yr^{-1}) and area-averaged basal melt (m yr^{-1}) are shown for each ice shelf. Data from Rignot et al. (2013).

West Antarctica		
	Total basal melt (Gt yr^{-1})	Area-averaged basal melt (m yr^{-1})
Pine Island	101.2 ± 8	16.2 ± 1
Thwaites	97.5 ± 7	17.7 ± 1
Crosson	38.5 ± 4	11.9 ± 1
Dotson	45.2 ± 4	7.8 ± 0.6
Getz	144.9 ± 14	4.3 ± 0.4

East Antarctica		
	Total basal melt (Gt yr^{-1})	Area-averaged basal melt (m yr^{-1})
Amery	35.5 ± 23	0.6 ± 0.4
West	27.2 ± 10	1.7 ± 0.7
Shackleton	72.6 ± 15	2.8 ± 0.6
Totten	63.2 ± 4	10.5 ± 0.7
Moscow Univ.	27.4 ± 4	4.7 ± 0.8
Mertz	7.9 ± 3	1.4 ± 0.6

influences the East Antarctic ice shelves. New measurements show that warm water enters the Totten Ice Shelf cavity (Rintoul et al., 2016) and drives rapid basal melting (Depoorter et al., 2013; Rignot et al., 2013; Liu et al., 2015). In this sense, the Totten system behaves in a similar way to the rapidly melting West Antarctic ice shelves that have received much more attention. Simulations of future change in the Antarctic Ice Sheet show that all three of the East Antarctic marine-based ice sheets discussed here lose mass and contribute to sea level rise on time scales of a few centuries (Golledge et al., 2015; DeConto and Pollard, 2016). Estimates of future sea level rise need to take the potential contribution from a dynamic EAIS into account.

Further work is needed to improve our understanding of the vulnerability of the East Antarctic Ice Sheet. We show here that warm water approaches the margin of Antarctica along much of the East Antarctic coastline, but that delivery of ocean heat to ice shelf cavities varies strongly from region to region, even for ice shelves that are relatively close together. This highlights the need for future work to more fully understand the physical processes regulating the transport of heat from the open ocean to an ice shelf cavity, and how these processes vary from region to region. In particular, very few oceanographic measurements have been made near Ninnis and Cook Glaciers, the primary ice streams draining the large marine-based Wilkes Basin ice sheet, or near Denman Glacier, which drains a portion of Aurora Basin and supplies the Shackleton Ice Shelf. How ocean heat flux to ice-shelf cavities varies in time is even less well understood. Very few time-series measurements have been made near East Antarctic ice shelves (the mooring and borehole work at the Amery Ice Shelf is a notable exception). Glaciological measurements indicate the grounded portion of Totten Glacier has thinned (Harig and Simons, 2015) and retreated (Li et al., 2015) in recent decades and that the ice shelf volume is either decreasing (Liu et al., 2015) or highly variable in time, with no significant trend (Paolo et al., 2015). No oceanographic data exist to test the hypothesis that changes in ocean heat transport have driven the recent changes observed in the floating and grounded portions of Totten Glacier. Observing systems capable of tracking variations in ocean temperature and circulation over time near Antarctic ice shelves are urgently needed to improve our understanding of the susceptibility of the ice sheet to ocean change.

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