Decadal Ocean Forcing and Antarctic Ice Sheet Response

LESSONS FROM THE AMUNDSEN SEA

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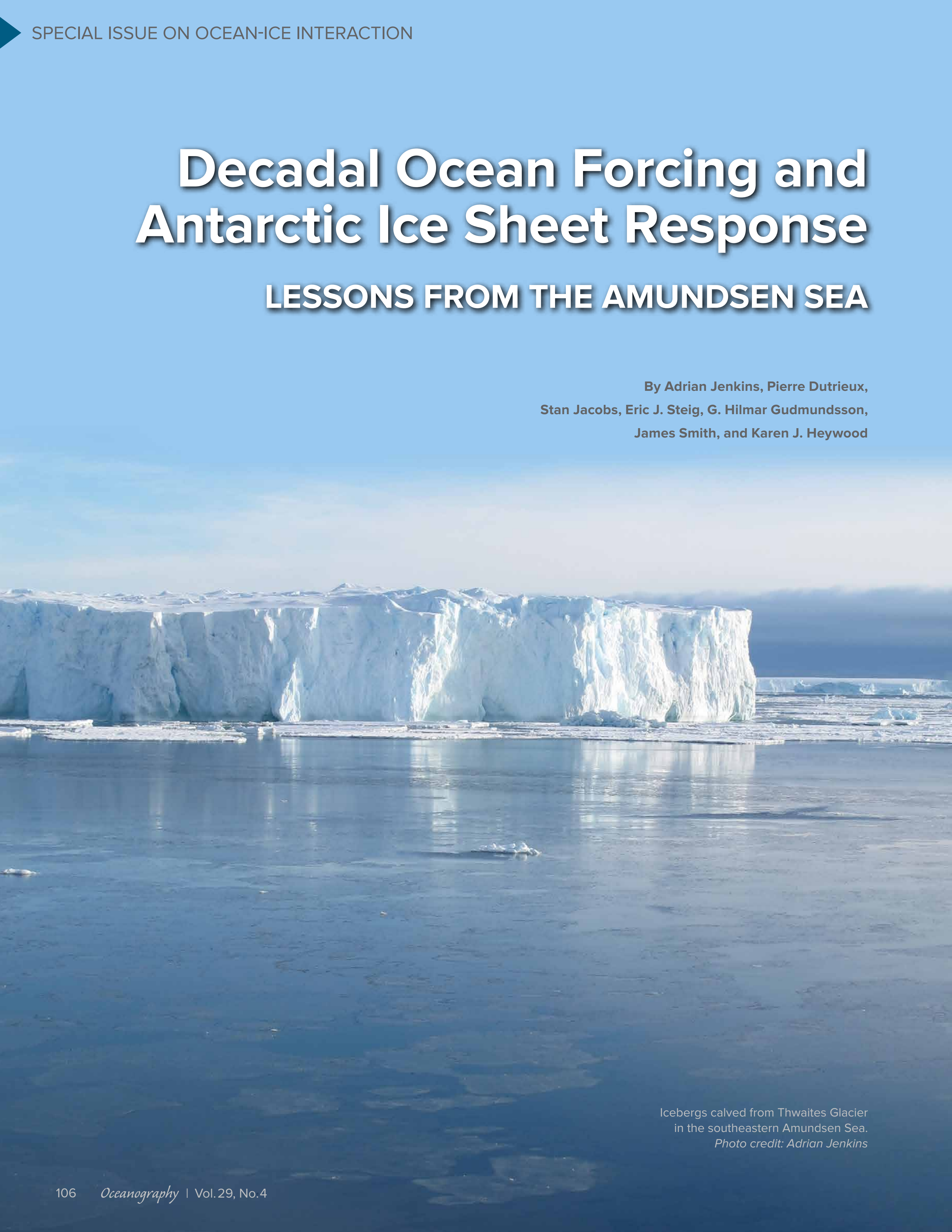

Icebergs calved from Thwaites Glacier in the southeastern Amundsen Sea. *Photo credit: Adrian Jenkins*

ABSTRACT. Mass loss from the Antarctic Ice Sheet is driven by changes at the marine margins. In the Amundsen Sea, thinning of the ice shelves has allowed the outlet glaciers to accelerate and thin, resulting in inland migration of their grounding lines. The ultimate driver is often assumed to be ocean warming, but the recent record of ocean temperature is dominated by decadal variability rather than a trend. The distribution of water masses on the Amundsen Sea continental shelf is particularly sensitive to atmospheric forcing, while the regional atmospheric circulation is highly variable, at least in part because of the impact of tropical variability. Changes in atmospheric circulation force changes in ice shelf melting, which drive step-wise movement of the grounding line between localized high points on the bed. When the grounding line is located on a high point, outlet glacier flow is sensitive to atmosphere-ocean variability, but once retreat or advance to the next high point has been triggered, ocean circulation and melt rate changes associated with the evolution in geometry of the sub-ice-shelf cavity dominate, and the sensitivity to atmospheric forcing is greatly reduced.

INTRODUCTION

Despite featuring prominently in debate over the potential instability of the marine-based West Antarctic Ice Sheet (Thomas et al., 1979; Hughes, 1981), the Amundsen Sea sector of Antarctica (Figure 1) received only limited attention until 1994, when Jacobs et al. (1996) made the first observations of ocean conditions on the continental shelf in the eastern Amundsen Sea. They revealed the ubiquitous presence of slightly modified Circumpolar Deep Water (CDW) on the deeper parts of the shelf beneath a colder and fresher layer of Antarctic Surface Water (AASW). Maximum temperatures below the main thermocline were just above 1°C in the east, but declined to the west as the overlying layer of AASW thickened (Jacobs et al., 2012). The combination of warm ocean waters on the shelf and the deep draft of the glaciers draining from the heart of West Antarctica into the eastern Amundsen Sea was shown to give rise to exceptionally high melt rates beneath the floating ice shelves that form at the glacier termini (Jacobs et al., 1996; Rignot, 1998), raising the prospect of marine ice sheet retreat in response to loss of buttressing from thinning ice shelves. Nevertheless, at the end of the twentieth century, no clear evidence had been found for imbalance between snowfall in the interior of the ice sheet and drainage through the glaciers (Vaughan et al., 2001), although some retreat of the grounding line of Pine Island Glacier had been documented in the mid-1990s (Rignot, 1998).

The picture of approximate glaciological balance changed with the advent of satellite measurements of surface elevation that extended over the interior of the ice sheet, starting in the early 1990s. After a decade of data gathering, clear signals of imbalance emerged, particularly along the marine margins of the ice sheet. The most extensive and rapid thinning was seen over the glaciers draining into the eastern Amundsen Sea (Shepherd et al., 2002), and expansion and acceleration of the thinning could be discerned as the record lengthened (Wingham et al., 2009). Data over the ice shelves revealed even more rapid thinning (Shepherd et al., 2004) at rates that could only be explained by increased basal melting. Synthetic aperture radar instruments flown on the same satellites allowed ice flow speeds to be measured and grounding lines to be mapped, completing a coherent picture of the process of change. Ocean-driven thinning of the ice shelves led to acceleration of the inland ice flow (Rignot, 2008), and as grounded ice thinned in response, regions floated free of the bed and the grounding line retreated inland (Rignot, 1998; Joughin et al., 2010). Further acceleration followed in a positive feedback that mirrored theoretical descriptions of the demise of marine ice sheets (Weertman, 1974; Thomas et al., 1979; Schoof, 2007).

The changes described above are now well documented, with recent summaries (Rignot et al., 2014; Mouginot et al., 2014) providing quite detailed pictures of ice sheet change in the Amundsen Sea sector of West Antarctica over the past few decades. The growing interest in the region and the urgency behind answering key questions about how far and how fast the current changes are likely to continue (Joughin et al., 2014) has stimulated further oceanographic observations on the continental shelf. There now exists a spatially and temporally patchy record of shelf water properties extending over a period of 20 years. Despite sparse sampling, that time series provides invaluable insight into the ocean forcing that accompanied glaciological change.

Here, we present and discuss elements of those ocean observations and the implications they have for our understanding of how ocean forcing of ice shelves can drive changes in outlet glaciers that drain the interior of a marine ice sheet. We begin with a brief summary of the observations of recent glaciological change and the evidence for its longer-term context. There next follows a discussion of the processes that determine the mean state of the Amundsen Sea continental shelf waters and impose variability on that mean state. Understanding those processes is the key to making sense of the discontinuous observational record and its longer-term context. Finally, we draw the oceanographic and glaciological evidence together to show how ocean and ice sheet behavior might be linked over time scales up to many decades.

GLACIOLOGICAL CHANGE IN THE AMUNDSEN SEA SECTOR

At the Last Glacial Maximum, grounded ice extended to, or near to, the continental shelf edge in the eastern Amundsen Sea, and well-dated marine geological data indicate that post-glacial retreat began around 20 kyr ago (Smith et al., 2014), taking the grounding line to within ~100 km of its current position by ~10 kyr ago (Hillenbrand et al., 2013). The average rate of retreat over the course of the

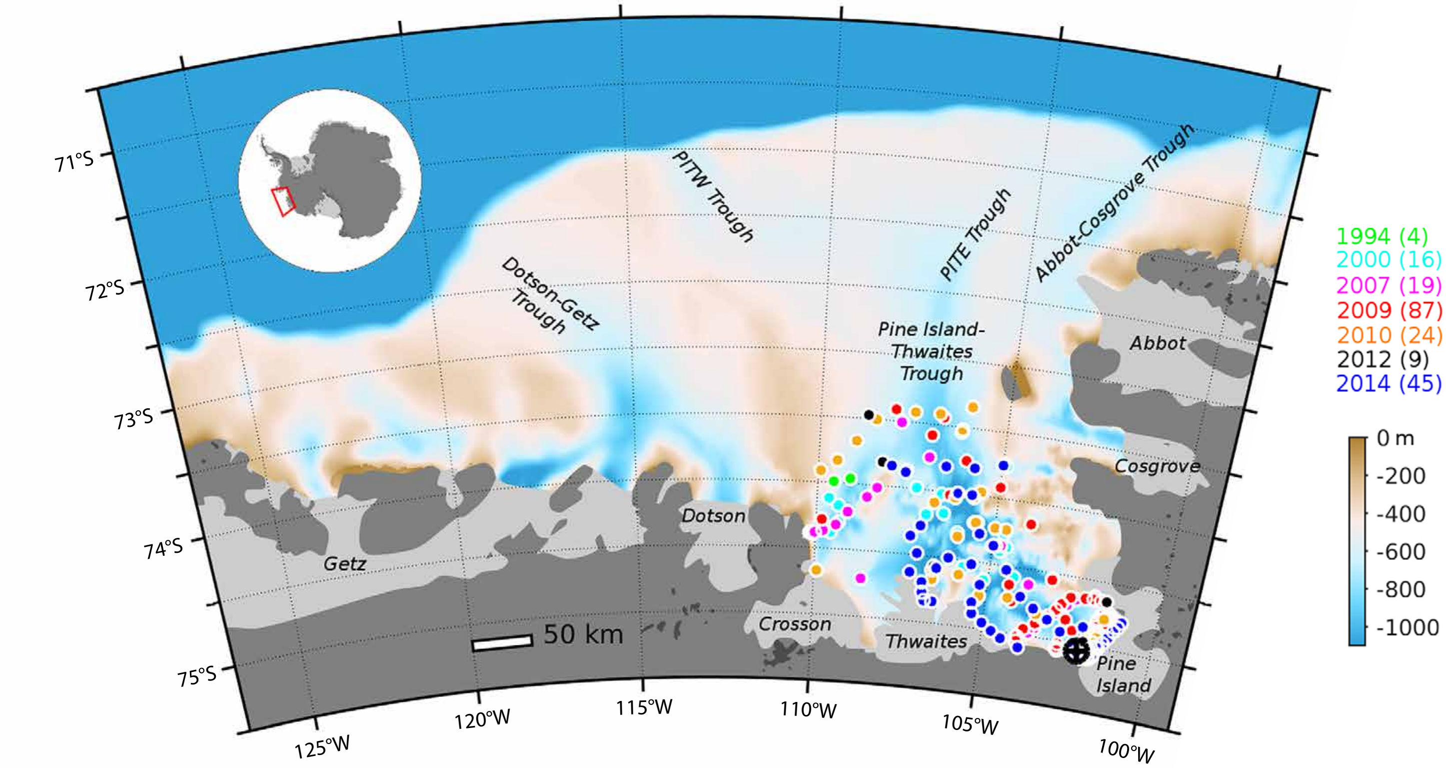

FIGURE 1. Bathymetry of the eastern and central Amundsen Sea continental shelf (color scale), with ice sheet (dark gray) and ice shelves (light gray, named) overlaid. Colored dots indicate locations and years of summer CTD stations used to derive mean thermocline depth in Figure 5 (key to right with year and number of stations), and the circled black cross shows the location of mooring records also used in Figure 5. Names of seabed troughs, including Pine Island-Thwaites East (PITE) and Pine Island-Thwaites West (PITW) follow Smith et al. (2014), while bathymetry comes from the RTopo-1 gridded data set (Timmermann et al., 2010).

deglaciation was thus at least an order of magnitude slower than the rates recently observed (Rignot, 1998; Park et al., 2013; Rignot et al., 2014), but the seabed sedimentary record indicates a stepped deglaciation, with phases of relative stasis separating periods of more rapid change (Smith et al., 2014). Interior thinning of the ice sheet in response to the marginal retreat was also variable and included a period around 7–8 kyr ago when thinning rates comparable to today's were sustained over decades to centuries, and were perhaps triggered by the breakup of an extensive ice shelf that had survived over the inner continental shelf (Johnson et al., 2014). That picture of episodic rapid change accords with current theory of marine ice sheet dynamics (Schoof, 2007), which predicts rapid and irreversible retreat of the grounding line over sections of the bed that slope down toward the interior of the ice sheet, unless the buttressing provided by a confined ice shelf can provide stability (Goldberg et al., 2009; Gudmundsson, 2013).

Observations of the seabed beneath Pine Island Ice Shelf revealed a 300 m high submarine ridge (Jenkins et al., 2010) 20–30 km downstream of the 1992 grounding line (Rignot, 1998). The ridge crest is a strong candidate for the location of the grounding line immediately prior to the current phase of retreat. If the rates recorded since 1992 are representative of those prior to that time, they suggest a time scale as short as a few decades for the initial retreat from the ridge to the 1992 location. However, the earliest satellite imagery from 1973 suggests that much of the initial retreat had already occurred, although the resulting ice shelf was probably still in contact with the highest point of the ridge at that time (Jenkins et al., 2010). Furthermore, the grounding line promontory mapped in the 1990s (Rignot, 1998) was shown to be a lightly grounded "ice plain" formed by a seabed rise (Corr et al., 2001). Thus, the most likely scenario is that retreat of the grounding line from the ridge crest started some decades before the 1970s, but change was either relatively slow while the ice maintained contact with the higher parts of the ridge, or halted for a while on the seaward side of the ice plain. Subsequent retreat across the ice plain has been well documented (Rignot, 1998; Park et al., 2013; Rignot et al., 2014), proceeding at an average rate close to 1 km yr^{-1}, and being accompanied by a 60% increase in the ice flux across the grounding line (Mouginot et al., 2014). The retreat has slowed since the grounding line reached the landward side of the ice plain (Park et al., 2013), and the ice flux has stabilized since 2010, even showing a slight decrease since 2012 (Mouginot et al., 2014), although inland thinning has continued to spread and accelerate (McMillan et al., 2014). A broadly similar picture has emerged for the other glaciers that drain into the eastern Amundsen Sea (Thwaites, Haynes, Smith, Pope, and Kohler Glaciers), with all showing periods of rapid grounding-line retreat and flow acceleration since 1992 (Rignot et al., 2014; Mouginot et al., 2014). The near-synchronous response of so many individual drainage basins points to an oceanic trigger.

OCEANOGRAPHIC CONDITIONS OVER THE AMUNDSEN SEA CONTINENTAL SHELF

Figure 2 shows Circum-Antarctic water properties at the depths of greatest relevance for the ice shelf cavities. Potential temperatures over the continental shelves (Figure 2a) range from just below −2°C to just over 1.5°C. Most of the sector extending from the northern tip of the Antarctic Peninsula near 55°W, around East Antarctica to the eastern side of the Ross Sea continental shelf near 155°W, is dominated by waters at the low end of this range, while the remaining sector, the Amundsen and Bellingshausen seas and the western Antarctic Peninsula, is mostly dominated by waters at the upper end. This observation has led to the frequent assumption that there are two distinct continental shelf regimes, each characterized by its own processes and variability. However, the salinity distribution (Figure 2b) tells a different story. The salinity itself has little direct impact on melting, but the differences reflect different water masses on the continental shelf.

Whitworth et al. (1998) suggested that because there is little distinction between on- and off-shelf waters near the surface, the lighter classes of on-shelf water can be regarded as forms of AASW. Beneath the AASW, they identified either modified forms of CDW or a water mass they termed simply Shelf Water (SW), defined as near-freezing water that is denser than the local variants of CDW. AASW is formed when CDW that upwells south of the Polar Front is cooled and freshened by interaction with the atmosphere and the ice. Over the continental shelf, the salinity increase associated with net sea ice production can raise the density of the AASW above that of the CDW, creating SW. However, Figure 2b indicates that the regions where such high salinities are generated are atypical, and that around much of the coast the cold shelf waters remain less dense than the warmer off-shelf CDW (Figure 2c).

Although the large-scale atmospheric circulation around Antarctica is dominated by westerly flow associated with the high-latitude jet, the surface winds at the edge of the continent are predominantly easterly as a result of air masses sinking over the pole, spreading equatorward, and coming under the influence of Earth's rotation. The easterlies are enhanced, particularly in winter, because cooling of the air over the ice sheet creates a cold, dense layer that drains from the polar plateau as gravity-driven katabatic winds. Ekman transport is driven by the easterly winds toward the continent, transporting AASW onto the continental shelf and driving downwelling at the coast where southward Ekman transport is blocked. Wind-driven coastal downwelling

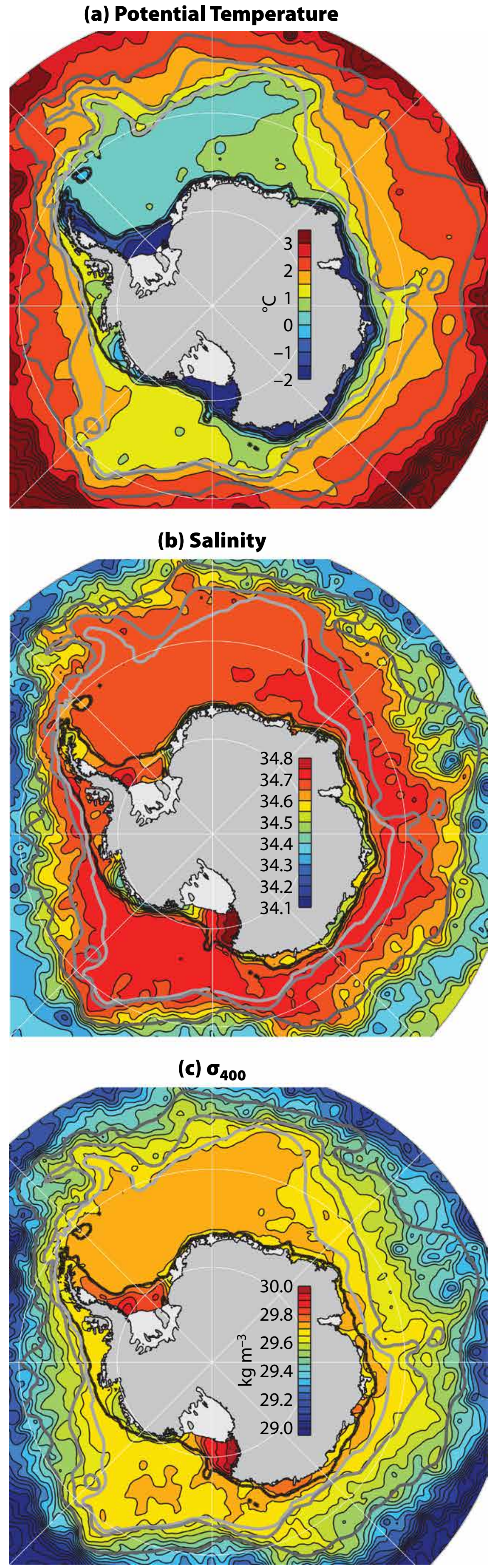

FIGURE 2. Potential temperature (a), salinity (b), and density (c) south of 45°S averaged from 400 m depth to either 1,200 m depth or the seabed, whichever is shallower. Density (c) is that associated with the averaged properties (a, b) at a pressure of 400 dbar. Gray shading indicates the Antarctic Ice Sheet (dark) and ice shelves (light). The bold black line indicates the 1,000 m isobath, while the bold gray lines show the southern boundary of the Antarctic Circumpolar Current (ACC; light), the Southern ACC Front (medium) and the Polar Front (dark). Data are taken from the World Ocean Circulation Experiment Southern Ocean Atlas Database (Orsi and Whitworth, 2005).

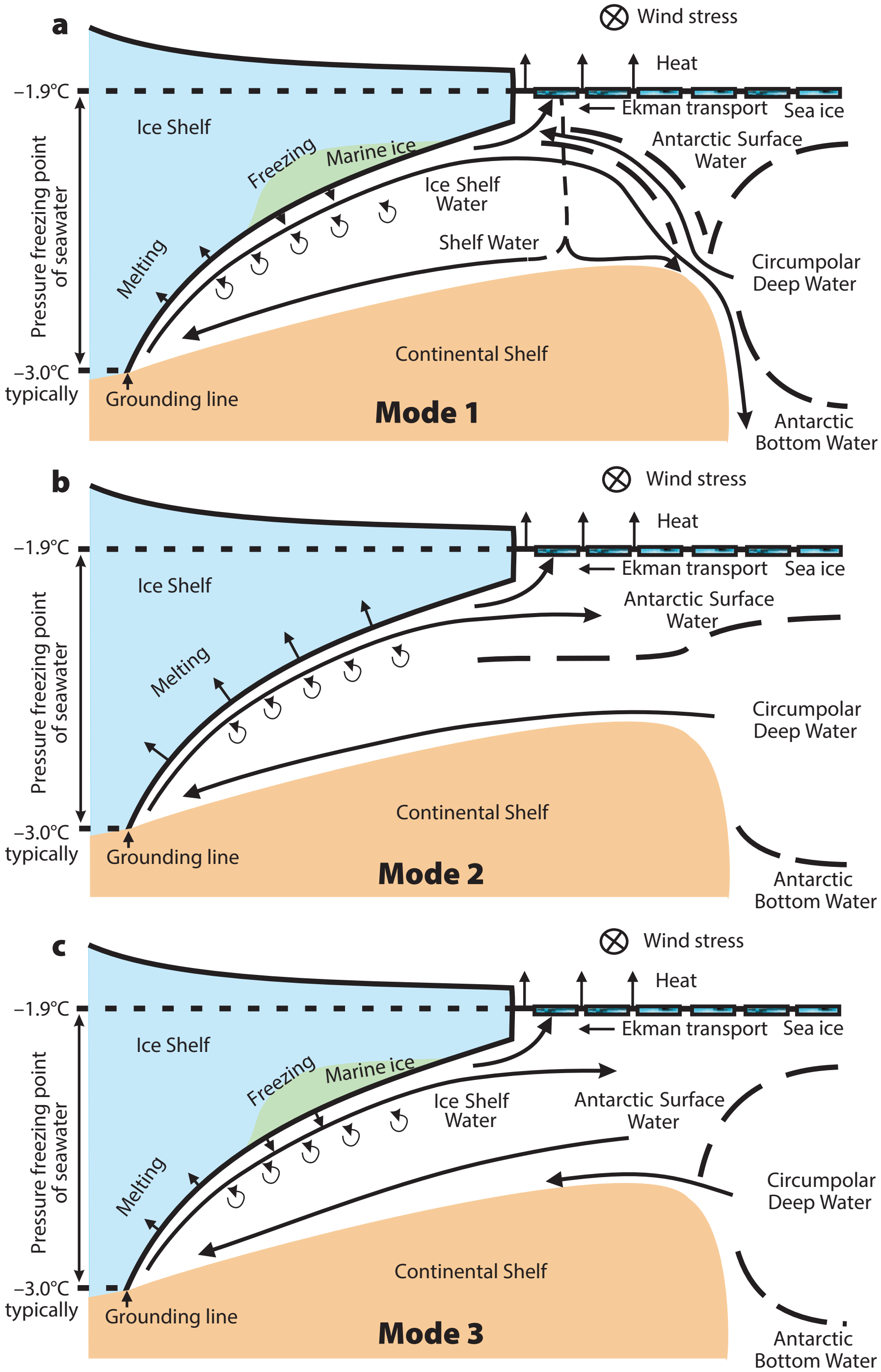

FIGURE 3. Three modes of sub-ice-shelf circulation and associated stratification on the continental shelf, after Jacobs et al. (1992). (a) In Mode 1, dense Shelf Water, formed by brine rejection beneath growing sea ice, dominates the sub-ice cavity. Shelf Water has a temperature at or close to the surface freezing point, and can melt ice at depth only because of the pressure dependence of the freezing point. Some refreezing occurs in the cavity because the water produced by melting (Ice Shelf Water) becomes supercooled as it rises along the shoaling ice shelf base. (b) Mode 2 dominates if Shelf Water is absent and Circumpolar Deep Water is the densest water on the shelf. Circumpolar Deep Water temperatures are typically around 3°C above the surface freezing point so melting is rapid, no Ice Shelf Water forms, and there is no refreezing. (c) Mode 3 dominates where both Shelf Water and Circumpolar Deep Water are absent, leaving Antarctic Surface Water as the densest water on the shelf. Only the upper layer of Antarctic Surface Water is seasonally warmer than the surface freezing point, so melt rates are low and Ice Shelf Water formation and refreezing can result. Although the Circumpolar Deep Water is denser, its access to the shelf is limited by the deepening of the Antarctic Surface Water layer at the coast, where the southward Ekman transport driven by the easterly wind is blocked. Note that in (a), Modes 2 and 3 may influence the outer cavity because Antarctic Surface Water and modified Circumpolar Deep Water are present in the upper water column, while in (b), Mode 3 melting may occur above the permanent thermocline separating Antarctic Surface Water and Circumpolar Deep Water.

of AASW generates the Antarctic Slope Front that separates fresher and colder on-shelf AASW from warmer and saltier off-shelf CDW, limiting the on-shelf intrusion of CDW and keeping the shelf cold along sections of the East Antarctic coastline where SW is absent (Nøst et al., 2011). We can thus understand the on-shelf presence of CDW in the sector from 55°W to 155°W as a result of surface buoyancy forcing that is too weak to produce SW (Talbot, 1988; Petty et al., 2013) and coastal downwelling that is too weak to deepen the layer of AASW to the seabed. The resulting large input of meltwater helps to maintain the on-shelf stratification by limiting the buoyancy- and wind-forced deepening of the AASW.

Overall, the subsurface waters on the Antarctic continental shelf occupy a relatively narrow thermohaline range (Figure 2) that can be classified as cold and salty (SW), warm and salty (CDW), or cold and fresh (AASW), and each can be related to a distinct mode of sub-ice-shelf circulation and melting (Figure 3). Because cooling and brine rejection beneath growing sea ice create both SW and the cold core of AASW (often termed Winter Water, WW), Mode 1 and most Mode 3 melting is driven by water with a temperature close to the surface freezing point. However, Mode 3 melting can be enhanced where wind-forced downwelling is sufficient to drive the seasonally warmer upper layer of AASW beneath the ice shelf (Hattermann et al., 2012), while weaker downwelling can allow incursions of modified CDW along the seabed (Nøst et al., 2011) that result in higher Mode 2 melting at depth. In contrast, a weakening of the buoyancy forcing that creates SW formation is necessary to reduce Mode 1 melting, while the wind regime will then determine whether the Mode 2 or Mode 3 cell grows in response. The Mode 2 regime (Figure 3b) is likely to be most sensitive to atmospheric variability, because any associated vertical displacements of the thermocline separating AASW and CDW will alter the extents of the high-melt Mode 2 cell and the overlying, low-melt Mode 3 cell.

Finally, we note that although Mode 2 melting dominates at depth throughout the sector from 55°W to 155°W, the processes that drive variability might differ through the region. The Antarctic Peninsula extends into the zone of the circumpolar westerlies, so there is no wind-forced coastal downwelling, and the Antarctic Circumpolar Current (ACC) flows along the continental slope. Eddies shed by the ACC are therefore one of the main processes driving renewal of, and variability in, the on-shelf CDW (Martinson and McKee, 2012). However, further to the west, easterly winds over the shelf become increasingly important. In the Amundsen Sea, there is a well-developed westward circulation over the continental shelf and slope, and a weak but discernible Antarctic Slope Front at the continental shelf break, while the ACC lies further to the north (Walker et al., 2013), so it is likely to be a less influential source of on-shelf variability.

PROCESSES DRIVING OCEAN VARIABILITY ON THE AMUNDSEN SEA CONTINENTAL SHELF

Despite the growing interest in the Amundsen Sea sector, little was known a decade after the initial oceanographic investigations beyond the basics of the bathymetry and the summer oceanic state (Jacobs et al., 2012). However, a modeling study by Thoma et al. (2008) identified significant variability in the AASW/CDW thermocline depth over the inner continental shelf on all time scales, from seasonal to decadal, that was related to the strength of the CDW inflow at the shelf edge. The inflow was focused in a seabed trough where significant on-shelf transport of CDW had previously been inferred from observations (Walker et al., 2007), and changes in the modeled inflow were linked with the strength of the zonal winds over the shelf edge. Steig et al. (2012) showed that the shelf edge winds are influenced by sea surface temperature variability in the central tropical Pacific Ocean. Surface warming there enhances atmospheric convection, creating tropospheric height anomalies that propagate to the Amundsen Sea as a standing Rossby wave train (Ding et al., 2011). Resulting sea level pressure anomalies in the Amundsen Sea weaken the prevailing easterly winds over the Amundsen Sea shelf and can bring westerly air flow over the continental slope and shelf edge. Thoma et al. (2008) argued that such a pattern of wind forcing led to enhanced inflow of CDW at the shelf edge.

The suggested link with the tropical Pacific received direct observational support in 2012 when pronounced cooling of the eastern Amundsen Sea shelf followed a period of low temperature in the central tropical Pacific and anomalously strong easterly winds over the Amundsen Sea continental shelf and slope (Dutrieux et al., 2014). However, other observational and modeling studies (Schodlok et al., 2012; Nakayama et al., 2013; Wåhlin et al., 2013; St-Laurent et al., 2015) have raised questions about the location and seasonality of the main CDW inflow described by Thoma et al. (2008) and suggested surface buoyancy forcing as the primary control on shelf water properties. We address those issues now through a re-examination of the Thoma et al. (2008) results, which were obtained with an isopycnic coordinate model (Figure 4).

The Thoma et al. (2008) analysis focused on the model's deepest layers (7 and 8), which were assumed to be the waters reaching the deep ice shelf grounding lines, and which came on-shelf mainly through the Pine Island-Thwaites West (PITW) Trough (Figure 4c). The subsequent discovery of a seabed ridge beneath Pine Island Ice Shelf (Jenkins et al., 2010) has shown mid-water column variability to be more important. Refocusing on the 400–700 m depth range, it is clear from Figure 4b that isopycnic layer 6 also carries significant heat to the sub-ice cavities. The distribution of layer 6 over the shelf (Figure 4c) shows that it comes on-shelf primarily through the Abbot-Cosgrove Trough, where observations typically reveal the presence of warmer and lighter classes of CDW derived from the Upper CDW core off-shelf (Jacobs et al., 2012). The picture portrayed in Figure 4c of inflows from both PITW and Abbot-Cosgrove troughs merging and flowing south in Pine Island-Thwaites Trough is consistent with other model results (Assmann et al., 2013; Nakayama et al., 2014) and observations (Nakayama et al., 2013). Had the analysis of Thoma et al. (2008) been based on the thicknesses of layers 6, 7, and 8, rather than on those of just the latter two, the conclusions would have differed little, because depth changes of the layer 5/6 interface closely track those of the layer 6/7 interface (Figure 4b).

A recent modeling study (St-Laurent et al., 2015) focused on the contribution of surface forcing to mid-water column variability, arguing that mid-depth cooling is driven by convective deepening of the AASW layer in polynyas that form along the eastern coast of the Amundsen Sea. From Figure 4 it is clear that winter deepening of layer 1 in the Thoma et al. (2008) simulation played an analogous role in seasonal cooling of the upper water column. Nevertheless, observations on the landward side of the seabed ridge beneath Pine Island Ice Shelf (Jenkins et al., 2010) show that the high melt rates near the grounding line are fueled by waters found at depths that just clear the ridge crest (around 700 m). Variability at this depth is linked with the deeper layers rather than with the surface in both models, except possibly in extreme years. The results of St-Laurent et al. (2015) show seasonal warming at 700 m, much like that described by Thoma et al. (2008). The warming in both models peaks in late winter, so appears to be unrelated to local surface cooling, and that observation is what led Thoma et al. (2008) to seek a remote source for the signal. From Figure 4b it is clear that the processes responsible for deepening of the AASW layer are enhanced when the input of CDW at the shelf edge is weaker. The reason is that the strong easterly winds over the shelf that Thoma et al. (2008) associated with weak inflow also promote polynya formation along the eastern coast (St-Laurent et al., 2015) and downwelling at the southern coast (Kim et al., 2016).

Given the coarseness of the model grid and forcing, Thoma et al. (2008) remained equivocal about the origins of the link between shelf-edge winds and CDW inflow. However, a strong candidate mechanism was revealed by observations of shelf-edge currents in and around the PITW Trough (Walker et al., 2013). Beneath a westward surface current, an eastward undercurrent, created by the isopycnal slope associated with the Antarctic Slope Front, was shown to transport CDW along the shelf break. Walker et al. (2013) showed that the undercurrent was strongest immediately to the west of the PITW Trough, then weakened in the trough and partially reformed to the east, consistent with the bulk of the current turning on-shelf within the trough and supplying the southward flow that had previously been identified (Walker et al., 2007). The structure of the shelf-edge currents within the Thoma et al. (2008) model was shown to be consistent with these observations, albeit with weaker baroclinicity as a result of the poorly resolved Antarctic Slope Front (Walker et al., 2013). The implication is that the undercurrent represents a relatively steady baroclinic flow of CDW onto the shelf, upon which the shelf-edge winds impose barotropic variability. A consequence of the weak baroclinicity in the Thoma et al. (2008) model is a weak time-averaged flow and an over-estimate of wind-forced variability. That resulted in an over-estimate of the seasonality of the inflow, but serendipitously highlighted a key underlying process.

Subsequent modeling studies (Assmann et al., 2013; Nakayama et al., 2014) resolve the shelf edge undercurrents much better and show that they are responsible for the on-shelf flow of CDW in all the major eastern Amundsen troughs, although a systematic investigation of the link between undercurrent strength and wind forcing is lacking. An analogous but weaker undercurrent has been identified deeper on the continental slope in the eastern Weddell Sea (Heywood et al., 1998), where stronger and more persistent easterly winds produce a stronger Antarctic Slope Front that intersects the seabed seaward of the shelf break (Nøst et al., 2011). Models suggest a link between undercurrent strength and wind forcing (Smedsrud et al., 2006).

Observations made near the mouth of the Dotson-Getz trough (Wåhlin et al., 2010) show an on-shelf flow of CDW of comparable magnitude to that identified by Walker et al. (2007). Subsequent time-series observations revealed a relatively steady baroclinic flow transporting warm water on-shelf along the eastern flank of the trough, with a much more variable barotropic component that correlated with the wind forcing at short time scales (Wåhlin et al., 2013), consistent with the above description of the currents associated with the Antarctic Slope Front.

Further work has shown that the observed inflow in Dotson-Getz Trough

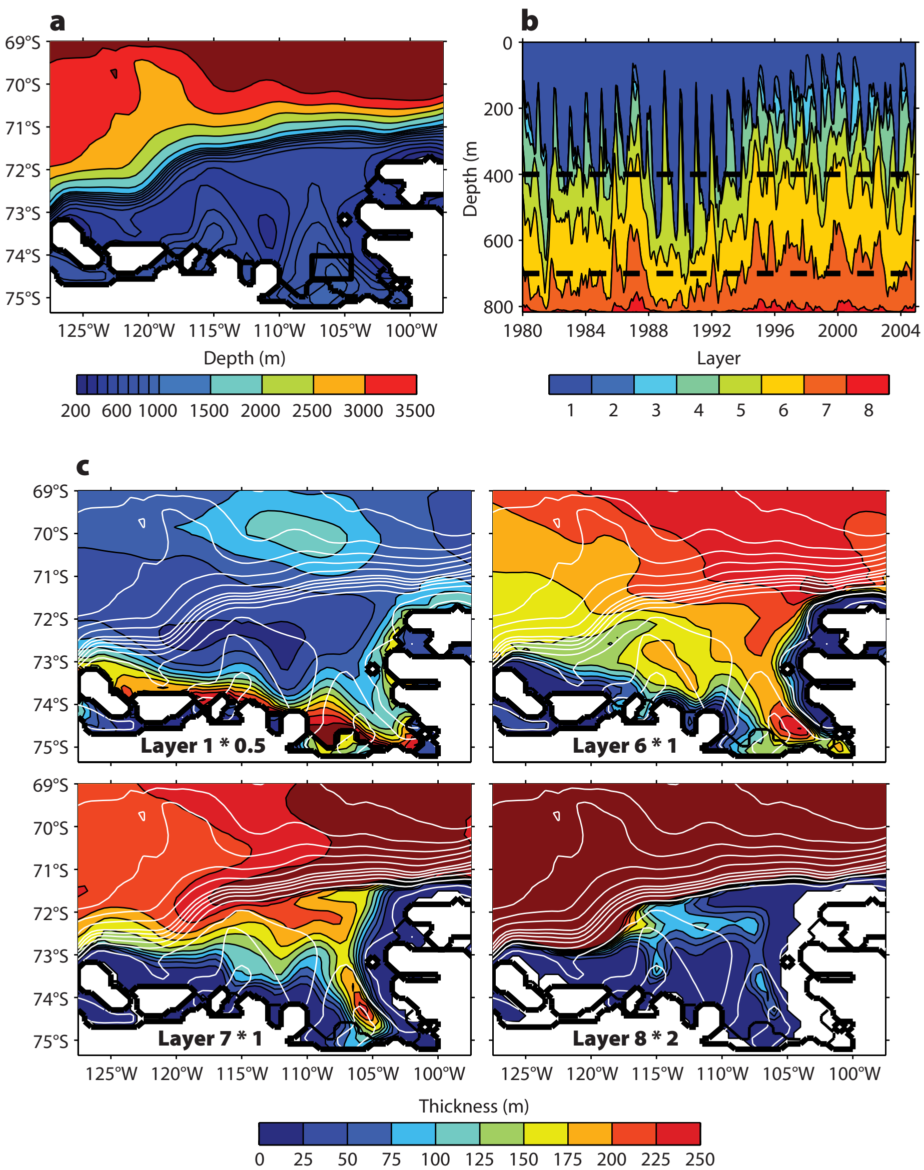

FIGURE 4. Results from the model of Thoma et al. (2008). (a) Model bathymetry and coastline. (b) Time series of isopycnal layer thicknesses averaged over the box near 74°S, 105°W in (a). Horizontal dashed lines indicate the approximate ice front draft and ridge height beneath Pine Island Glacier. (c) Mean thickness distribution of isopycnal layers 1, 6, 7, and 8 in September-October-November over the period shown in (b). Note the scale factors applied to the plotted thicknesses.

is part of a gyre that transports warm water south on the eastern side of the trough and cooler and fresher water north on the western side (Ha et al., 2014; Kalén et al., 2016). Model results (Kalén et al., 2016) show decadal-scale evolution in the strength of the gyre, with times of strong circulation being characterized by greater short-term (predominantly seasonal) variability and more rapid melting of the ice shelves. Schodlok et al. (2012) obtained similar results, pointing out that the gyre strength was the main determinant of heat transport toward the ice shelves. While winds will play a role, particularly at shorter time scales, it is possible that long-term changes in gyre strength are a response to ice shelf melting. Melting drives upwelling of CDW on the shelf because of the associated input of buoyancy at depth, allowing renewal of the water that would otherwise be topographically trapped and injecting cyclonic vorticity into the circulation. Simulations that lack the input of buoyancy at depth produce a much weaker, shelf-wide gyre, leaving the deep basins stagnant and unventilated (Pierre Mathiot, Met Office, *pers. comm.*, March 21, 2016). While the ice shelves were included in the domain of Thoma et al. (2008), they were poorly resolved, lacking the thickest ice near the grounding lines, so simulated melt rates were significantly lower than observed, reducing the on-shelf forcing of the gyre circulation.

In summary, some aspects of the Thoma et al. (2008) results are suspect, primarily due to coarse grid resolution, but there are good reasons to believe the suggested link between zonal wind forcing and thermocline depth. The zonal winds determine buoyancy forcing along the eastern coast (St-Laurent et al., 2015) and downwelling along the southern coast (Kim et al., 2016), as well as imposing time-varying barotropic forcing on the shelf (Wåhlin et al., 2013) and shelf-edge circulation, while observations and models all point to the importance of the shelf edge undercurrents in delivering warm water to the shelf via the shelf edge troughs. Despite these arguments in favor of the Thoma et al. (2008) conclusions, the time series in Figure 4b is a poor match with observations, discussed in the following section, and more recent modeling efforts. However, given the smoothed topography, the weak slope front, the weak on-shelf gyres, and the known biases of the coarse resolution wind forcing applied (Bracegirdle, 2013), that should come as no surprise. Indeed, if the physical processes are correctly represented, a model should not match observation exactly if the forcing is inexact, but the model's response to the applied forcing can still inform us about the physical processes that control the real-world response.

SHELF WATER VARIABILITY INFERRED FROM OBSERVATION

Figure 5 shows a Pine Island Bay record of thermocline depth inferred from recent mooring data and historical summer cruise data. We use the depth of the 0.8°C isotherm as a convenient proxy for thermocline depth for three reasons: the isotherm is always present in the lower thermocline, even in the coldest years observed; its depth provides a good indication of conditions around the critical 700 m depth level that supplies water to the inner cavity beneath Pine Island Ice Shelf; and it is deep enough that the seasonal signal is weak, so the bias in years when we have only summer observations should be limited. Mean temperature-depth profiles for individual summers were computed by averaging along isopycnal surfaces and remapping onto mean density-depth profiles. As spatial sampling of the cruises is uneven (Figure 1), we have accounted for spatial variations in thermocline depth by removing a mean offset between near- and far-field observations estimated from years that include profiles close to the Pine Island Ice Front. For those years, the ice front data allow estimates of meltwater transport away from the ice shelf (Dutrieux et al., 2014), and subject to calculation uncertainties, they demonstrate that the ice shelf melt rate has changed by a factor of at least two in response to the observed changes in thermocline depth (Figure 5).

Three important points emerge from Figure 5. First, the recent cooling that peaked in 2013 was not unique in the record but rather was a repeat of an event observed some 10–15 years earlier (Jacobs et al., 2012). Second, while Schmidtko et al. (2014) inferred a warming trend from seabed temperatures recorded up to 2010, our extended time series shows mid-water-column conditions to be dominated by decadal variability. Given the amplitude and apparent dominant periodicity of our record of thermocline depth, a much longer and better sampled record would be required to infer a significant trend. Third, although the sampling in the early part of our record could have missed higher frequency variations, the periodicity and phasing of the apparent dominant mode matches with the output of numerical models forced by recent atmospheric reanalyses that all show a cool period at the end of the 1990s and another in recent years (Schodlok et al., 2012; Kalén et al., 2016).

The preceding section of this paper discussed the physical processes responsible for variability on the eastern Amundsen Sea shelf (Figure 6) and the links with zonal wind forcing suggested by the study of Thoma et al. (2008). Given the limitations of that model, discussed above, we would expect to see a slightly different response in the data, with wind-forced variability of the shelf-edge currents making a smaller contribution to stronger background flows that evolve over longer time scales associated with changes in the density structure of the Antarctic Slope Front. Additionally, if we estimate the strength of the background inflow of CDW either from the on-shelf transport estimate of Walker et al. (2007) or the melt rates of Rignot et al. (2013) and an assumption that outflows comprise 1% meltwater and 99% CDW, we arrive at a residence time of two to eight years for CDW on the eastern Amundsen shelf. We should thus view the zonal wind changes as forcing variable input to a relatively large reservoir of CDW, and the volume of that reservoir should be related to the time-integral of the inflow (Figure 5). The phasing and periodicity of the dominant mode of variability forced by the winds matches well with that in the observational record, even if some details are not captured (e.g., the strength of the cooling in 2000). To illustrate the link with sea surface temperatures in the central tropical Pacific (Steig et al., 2012), we also plot the time-integrated temperature anomaly in the NINO4 region (5°N to 5°S and 160°E to 150°W) in Figure 5. While other factors must also influence zonal winds, which are furthermore not the only agents of change on the shelf, Figure 5 suggests that the tropical Pacific, through its impact on the regional atmospheric circulation, plays a key role in setting the period and phase of the dominant mode of variability.

ICE SHEET RESPONSE TO OBSERVED AND INFERRED FORCING

The clear implication of Figure 5 is that ocean forcing of the ice shelves in the eastern Amundsen Sea is characterized by strong decadal-scale variability, masking any trend that might exist. The near-monotonic trend in ice discharge (cyan line in Figure 5) thus cannot be explained by a trend in ocean forcing. The most rapid phase of acceleration on Pine Island Glacier in the late 2000s did coincide with some of warmest conditions observed in Pine Island Bay, and a slight reduction in discharge occurred during the most recent cool phase, but acceleration stopped at the height of the warm phase, and the earlier cool period observed in 2000 appears to have had almost no impact. A recent slowing of the grounding line retreat (Park et al., 2013) probably stabilized the discharge and put the glacier in a configuration that heightened its sensitivity to the most recent ocean cooling (Christianson et al., 2016). While the grounding line retreat was in rapid progress during the 2000 cool phase, the ongoing ice sheet reconfiguration was relatively unaffected by the change in forcing.

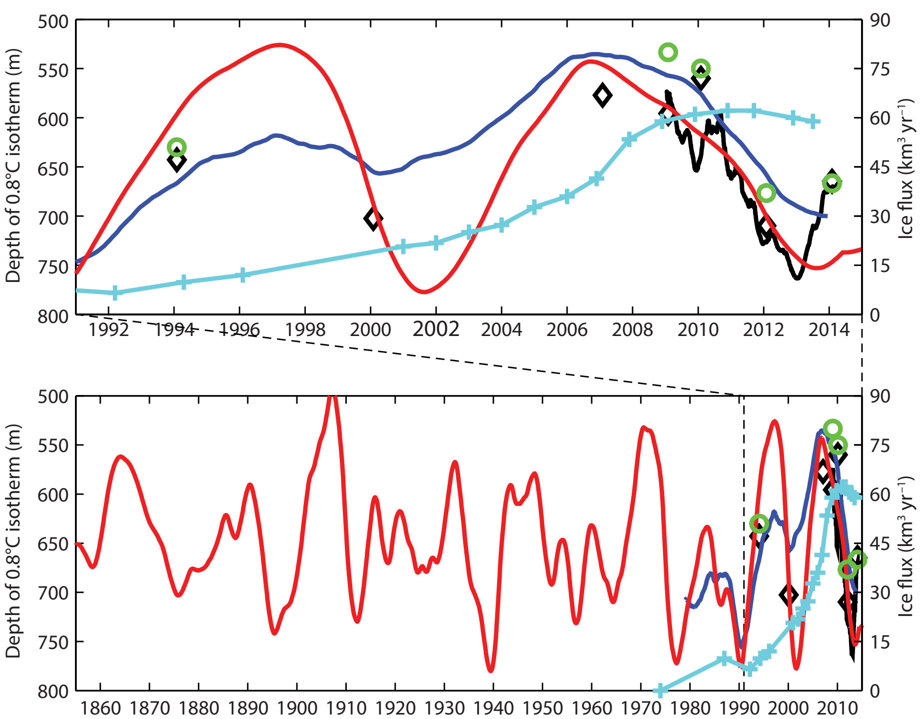

FIGURE 5. Proxies for thermocline depth on the inner shelf of the eastern Amundsen Sea over the last 160 years (the upper panel is an expanded version over the observational period of the lower panel). The depth of the 0.8°C isotherm (left-hand axis) is extracted from mooring data (black line; see Figure 1 for mooring location) and averages of summer CTD stations (black diamonds; see Figure 1 for station distributions). Less direct proxies come from the cumulative zonal wind anomaly (dark blue line) and cumulative central tropical Pacific sea surface temperature anomaly (red line). The former is an appropriately scaled time-integral of the zonal wind anomaly (relative to the mean) from ERA data averaged over 67°S to 71.5°S and 100°W to 114°W. The latter is an appropriately scaled time-integral of the sea surface temperature anomaly (relative to a 20-year running mean) from ERSSTv3b data averaged over the NINO4 region (5°N to 5°S and 160°E to 150°W). The use of a 20-year running mean removes a long-term trend in tropical Pacific surface temperature that is not directly relevant to the generation of the Rossby wave train that influences the Amundsen Sea. Green circles indicate the net melt rate (right-hand axis) of Pine Island Ice Shelf inferred from oceanographic observations (Dutrieux et al., 2014; Heywood et al., 2016, in this issue) while the cyan line and crosses show the excess flux of ice across the grounding line relative to a baseline of 85 $km^3 yr^{-1}$ observed in 1974 (Mouginot et al., 2014).

Such a response is analogous to the more rapid reaction of tidewater glaciers to seasonal and interannual forcing, in that a seasonal cycle of advance and retreat is seen when the glacier terminus is located on a seabed sill, but the process of retreat, acceleration, and thinning is continuous once the terminus has detached from the sill (e.g., Morlighem et al., 2016). Similarly, if the Pine Island Glacier grounding line is held on a seabed sill, the glacier can respond to variable ocean forcing, but when those fluctuations are sufficient to force the grounding line off the sill, subsequent acceleration, thinning, and grounding line retreat will continue largely unaffected by ocean forcing, until halted on the next inland rise in the bed. If an ice shelf remains intact at the glacier terminus, the reconfiguration is a coupled process, as movement of the grounding line and thinning of the floating ice will be accompanied by a change in sub-ice-shelf circulation and melting (Jacobs et al., 2011) that will be instrumental in driving the retreat (De Rydt and Gudmundsson, 2016), but this is largely independent of the far-field forcing (De Rydt et al., 2014).

The foregoing discussion suggests that the recent flow acceleration, ice thinning, and grounding line retreat of Pine Island Glacier (Rignot et al., 2014) are part of an ongoing response to the initial ungrounding from the crest of the seabed ridge that lies beneath the present ice shelf (Jenkins et al., 2010). While the grounding line was established there, it may have migrated back and forth on the crest in response to decadal-scale variability in ocean forcing. Some decades prior to the 1970s, a particularly strong or sustained warm phase must have pushed the grounding line too far back and triggered the reconfiguration we are presently observing. The process may have slowed when the grounding line reached the seaward side of an ice plain, and the deceleration observed between the late 1980s and early 1990s (Mouginot et al., 2014) might have been a response to an inferred (Figure 5) cooling while the glacier was in that configuration, as suggested by Thoma et al. (2008). A subsequent warm phase, sampled in the mid-1990s, triggered renewed retreat across the ice plain that was observed to continue through the next cycle of forcing. Most of the neighboring glaciers to the west of Pine Island showed a clearer response to the observed 2000 cool period, with greatly reduced acceleration at that time, and a dramatic speed-up during the subsequent warming (Mouginot et al., 2014). However, in all cases, the grounding lines showed little movement immediately prior to 2000 (Rignot et al., 2014), while the post-2000 warming appears to have triggered retreat.

If the dominant mode of ocean variability on the Amundsen Sea continental shelf is linked with tropical Pacific sea surface temperature, we can use the longer record of conditions in the Pacific to infer when the warming that initiated the recent changes might have occurred. A prominent inferred warming in the 1970s may have driven the thinning that resulted in the ice shelf losing contact with the ridge crest (Jenkins et al., 2010), while the preceding period of extended warmth that might have initiated the current retreat is inferred to be in the 1940s. Although there are other candidates in Figure 5, the 1940s also stand out in the paleoclimate record, as inferred from the stable isotopic composition of precipitation in ice cores, as the most anomalous climate in West Antarctica in the twentieth century, with the possible exception of the 1990s (Schneider and Steig, 2008; Steig et al., 2013). There is no comparable event in the West Antarctic ice core record any more recent than that of the 1830s (Steig et al., 2013). Moreover, sediment cores obtained from beneath Pine Island Ice Shelf indicate that waters from the open ocean first appeared landward of the ridge crest in the 1940s (Smith et al., 2016).

SUMMARY

The relatively rich record of glaciological change in the Amundsen Sea sector of West Antarctica collected over the past two and a half decades is complemented by a lengthening record of oceanic change on the continental shelf. Despite recent improvements, the sampling of ocean conditions remains too sparse to independently infer a trend or even the dominant mode of variability. Models suggest that the depth of the permanent thermocline over the continental shelf is sensitive to wind forcing through the impacts it has on surface buoyancy fluxes, coastal downwelling, and CDW input to the shelf (Figure 6). The thermocline responds on a range of time scales to these various forcings, and feedback resulting from the impact of melting on the continental shelf circulation may further complicate the response. However, changes in wind forcing provide a plausible explanation for some of the observed variability and suggest a dominant decadal periodicity associated with atmospheric circulation anomalies forced from the central tropical Pacific.

Recent observations of Pine Island Glacier show a muted response to multiyear cooling (Christianson et al., 2016), and our proxy records of shelf water conditions suggest that an inferred cooling

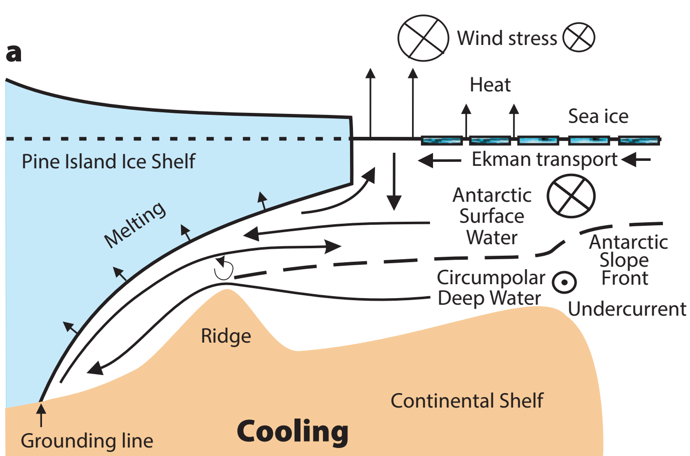

FIGURE 6. Schematic of processes that lead to (a) cooling and (b) warming of the eastern Amundsen Sea continental shelf. Cooling is promoted by strong easterly winds over the shelf that enhance polynya activity and coastal downwelling on the inner shelf while suppressing the Antarctic Slope Front undercurrent that brings Circumpolar Deep Water onto the shelf. A lower thermocline means that less Circumpolar Deep Water can access the inner cavity beneath Pine Island Ice Shelf and that it is more susceptible to mixing over the ridge crest with overlying waters. Warming is promoted by weak easterly winds that result in reduced polynya extent and weaker downwelling, while the shelf edge winds, which can switch to weak westerlies, enhance the inflow of Circumpolar Deep Water. A higher thermocline allows a thicker, warmer layer of Circumpolar Deep Water over the ridge.

in the early 1990s may have been responsible for a similar deceleration (Thoma et al., 2008). However, the lack of a clear response to an observed cooling in 2000 suggests that the glacier's sensitivity to ocean forcing is dependent on the stability of the grounding line position. In 2000, the grounding line was undergoing sustained retreat across an ice plain, and the acceleration of flow and thinning of the ice proceeded almost independently of the ocean state. Ocean-driven melting would still have played a key role in the process of retreat. As sections of the glacier floated free from the bed, basal ice, newly exposed to the ocean, would have experienced a rapid increase in melt rate from near zero to tens of meters per year. The accompanying changes in ocean circulation would have influenced melt rates throughout the growing subice cavity, and the geometrically driven changes in melting apparently dominated those associated with thermocline depth changes beyond the ice front (Jacobs et al., 2011). Thermocline depth changes driven by wind and buoyancy forcing would have reasserted their control on the glacier once the grounding line stabilized and the changes in cavity geometry slowed. The coupled ice stream/ice shelf/ocean system thus seems to function in a manner analogous to the well-known tidewater glacier cycle, but operating on a longer time scale.

Finally, we should note that although the physical processes outlined above should operate at the marine margin of any ice sheet, the Amundsen Sea sector of the West Antarctic Ice Sheet is perhaps uniquely sensitive to wind-forced oceanic variability. The Amundsen Sea is subject to some of the largest interannual atmospheric circulation variability in the Southern Hemisphere, at least in part because of the impact from the tropical Pacific. Furthermore, the mean state of the continental shelf is such that the heat content of the waters is particularly sensitive to changes in atmospheric forcing through the impact of wind and buoyancy forcing on thermocline depth. In the Amundsen Sea, decadal-scale variability linked to the tropical Pacific may thus dominate other processes that drive circumpolar ocean changes on the Antarctic continental shelf.

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