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THE AMUNDSEN SEA AND THE ANTARCTIC ICE SHEET

BY STAN JACOBS, ADRIAN JENKINS, HARTMUT HELLMER, CLAUDIA GIULIVI, FRANK NITSCHKE, BRUCE HUBER, AND RAUL GUERRERO

ABSTRACT. A few decades ago, Antarctic ice sheets were expected to grow as the atmosphere warmed and increasing poleward moisture transport added snowfall to regions that would remain below freezing year-round. Concerns about their sensitivity to climate change were centered on air temperature and on glacially paced ice dynamics. Southern Ocean roles were relegated to iceberg transport, a mix of melting and freezing under ice shelves buffered by the frigid shelf waters generated by sea ice production, and slow sea level rise by other forcing. At that time, observations were lacking in the remote Amundsen Sea, where difficult ice conditions have vexed explorers for more than 200 years. Mapping of its ocean structure and circulation began in 1994, revealing that “warm” Circumpolar Deep Water has access to its continental shelf. Glacially scoured troughs in the seafloor provide conduits for that seawater to melt regional ice shelves far more rapidly near their deep grounding lines. Coincident satellite data showed the ice shelves were thinning, in turn leading to accelerated glacier flow and loss of grounded ice to the sea. Repeated measurements and modeling suggest ocean changes that could impact the stability of the marine-based West Antarctic Ice Sheet.

BACKGROUND

The first ship to reach the continental shelf of the Amundsen Sea was HMS *Resolution* on January 31, 1774 (Figure 1). At 106°54'W, 71°10'S, that was the highest latitude attained by Captain James Cook during several attempts to find land around the South Pole (Beaglehole, 1974). William Walker valiantly tried to better that record on the US Exploring Expedition brig

Flying Fish, but was also turned back by heavy ice and foul weather near 100°16'W, 70°S in March 1839 (Stanton, 1975). The beset, westward-drifting *Belgica* unintentionally reached the northeast corner of the sea, later named after its second mate before breaking free near 70°45'S, 102°15'W on March 14, 1889 (Cook, 1909). *Discovery II* made it as far as 70°20.3'S, 102°48.2'W in early March 1938 (Discovery Reports,

1947). A persistent sea ice cover, thickened by heavy snowfall (Maksym et al., 2012, in this issue) and constrained by easterly winds and icebergs, also kept the USNS *Eltanin* at bay in the 1960s (Gordon, 2012, in this issue). Although the US Coast Guard icebreakers *Burton Island*, *Glacier*, and *Polar Sea* penetrated deep onto the shelf for exploratory, geological, and sea ice work (Seelig, 1975; Kellogg and Kellogg, 1987; Dalziel et al., 1992; Veazey et al., 1994), the region remained the largest unmapped area below 65°S in the Southern Ocean, well beyond the charting of Johnson et al. (1980).

The apparent lack of bottom water formation in the Amundsen sector (Orsi et al., 1995) may have damped historical oceanographic interest, although Circumpolar Deep Water (CDW) or its modified forms were known to reach the coastline to its west and east (Countryman and Gsell, 1966; Potter and Paren, 1985). This intriguing data gap led to use of the new US Research Vessel Icebreaker *Nathaniel B. Palmer* to survey the Antarctic coastal sector in the

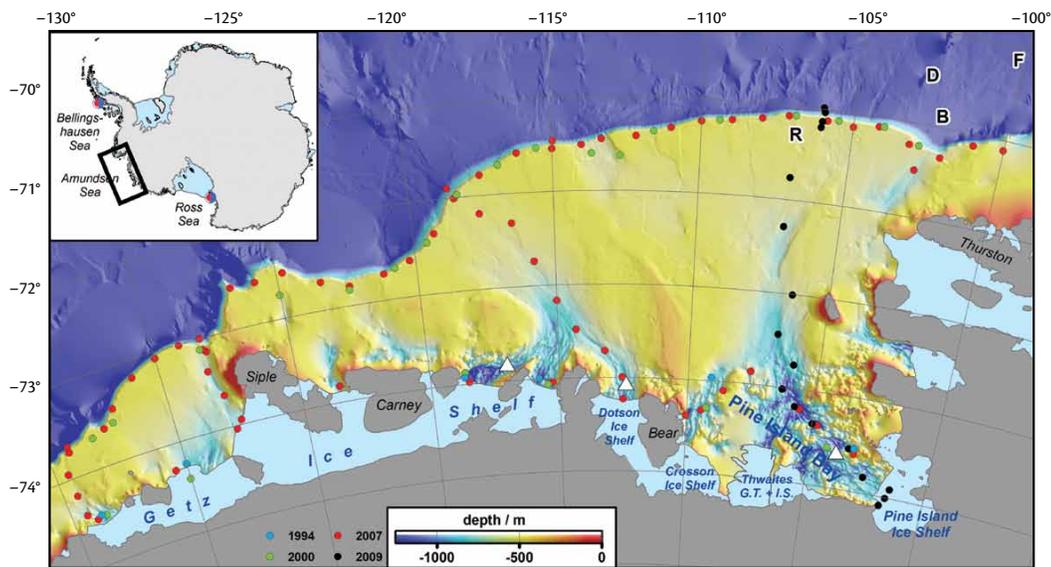


Figure 1. The Amundsen Sea continental shelf spans $\sim 315,000 \text{ km}^2$, larger than Arizona, and extends beneath several floating ice shelves along the Southeast Pacific coastline of the West Antarctic Ice Sheet. Compiled from Nathaniel B. Palmer, JC Ross, Polarstern, Oden, and Glacier measurements, its bathymetry is characterized by deep, glacially cut troughs that merge and shoal toward the continental shelf break. R, F, B, and D show the southernmost positions attained by HMS Resolution in 1774, Flying Fish in 1839, Belgica in 1889, and Discovery II in 1938. Colored dots locate the Nathaniel B. Palmer conductivity-temperature-depth (CTD) profiles used in Figures 2–6. White triangles indicate Figure 7 mooring sites seaward of the Getz and Dotson Ice Shelves, and in Pine Island Bay (PIB).

Southeast Pacific in early 1994. Living up to its forbidding reputation, a long, fast ice tongue anchored by hundreds of grounded icebergs barred the *Palmer's* eastward progress, and coincidental *Polarstern* entry, to the eastern continental shelf. The Pine Island Ice Shelf (PIIS) could only be accessed along the rugged eastern coastal route plied by the *Burton Island* and *Glacier* (Figure 2 in Hellmer et al., 1998). Conductivity-temperature-depth (CTD) profiling showed warm CDW reaching the PIIS and eroding it from below a couple orders of magnitude faster than cold shelf waters melt the large Ross and Filchner-Ronne Ice Shelves. This observation had important glaciological and oceanographic implications (Jenkins et al., 1994) in the context of risks to the West Antarctic Ice Sheet (WAIS) posed by ice shelf melting (Thomas, 1979) and ice dynamics in the absence of a large ice shelf buttressing ice stream flow (Hughes, 1981). Our limited 1994 measurements were cautiously interpreted, discounting probable upwelling of PIIS cavity outflow and assuming mass balance for the ice-ocean

system. Areal average basal melt rates were thus estimated to be $10\text{--}12.5 \text{ m yr}^{-1}$ (Jacobs et al., 1996; Jenkins et al., 1997; Hellmer et al., 1998).

Vaughan et al. (2001) reported no overall measurable imbalance in the Pine Island Glacier drainage basin, but contemporary satellite data were beginning to make equilibrium assumptions untenable. Thinning and faster movement of floating and grounded ice was coupled with rising estimates of in situ melting, particularly near grounding lines where seaward-bound ice streams begin to float (Rignot, 1998; Rignot and Jacobs, 2002; Shepherd et al., 2002, 2004). Rates of ice sheet loss were highest in the Southeast

Amundsen, focusing attention on the rapidly melting PIIS and its accelerating glacier, for which there are now frequent updates on thickness, velocity, and structure (Rignot, 2008; Wingham et al., 2009; Joughin et al., 2010; Bindshadler et al., 2011; Mankoff et al., 2012). A common inference has been that the ice shelves are being thinned where warming CDW intrudes nearly unmodified onto the continental shelf. Thinning shelf ice reduces sidewall and seafloor friction and, therefore, the back stress on grounded inflowing glaciers, which then can move more rapidly. Possible temporal and spatial variability of CDW properties and access to ice shelf cavities have

Stan Jacobs (sjacobs@ldeo.columbia.edu) is Lamont Research Professor, Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY, USA. **Adrian Jenkins** is Science Programme Coordinator, British Antarctic Survey, Cambridge, UK. **Hartmut Hellmer** is Physical Oceanographer, Alfred-Wegener-Institute, Bremerhaven, Germany. **Claudia Giulivi** is Senior Staff Associate, Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY, USA. **Frank Nitsche** is Research Scientist, Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY, USA. **Bruce Huber** is Senior Staff Associate, Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY, USA. **Raul Guerrero** is Head, Physical Oceanography Program, Instituto Nacional Investigacion Desarrollo Pesquero, Mar del Plata, Argentina.

thus become central to concerns about glacier thinning and acceleration, overall stability of the WAIS, and sea level rise (Joughin and Alley, 2011).

It took seven cruises on four ships over 15 years before initial observations near the PIIS could be repeated, and then augmented by autonomous underwater vehicle (AUV) mapping of its sub-ice cavity (Jenkins et al., 2012, in this issue). The growing gap above a transverse submarine ridge discovered beneath the thinning ice shelf was permitting increased ocean heat transport into deeper, more vulnerable grounding line regions (Jenkins et al., 2010). Inverse methods devised to estimate the

meltwater content of CTD/O profiles (Jenkins, 1999) revealed ~ 50% higher production in 2009 than in 1994 (Jacobs et al., 2011). That change appears to have resulted more from a large increase in the cavity circulation strength than a small rise in CDW temperature, consistent with the dependence of melting on seawater temperature and velocity, amplified by cavity geometry and freshwater crossing the grounding line (MacAyeal, 1984; Jenkins, 2011). The volume of CDW in Pine Island Bay (PIB; Figure 1) was also larger in 2009 than in 1994, raising its overlying thermocline to ridge-gap depths. That warm deep water becomes more modified westward along

the continental shelf, as can be shown by describing thermohaline structure and variability along the southern coastline, across the continental shelf, and near the continental shelf break. Limited space precludes more than a brief discussion of CDW access to and ocean modeling of the continental shelf regime, but more detailed information appears in the cited literature. The *Nathaniel B. Palmer* CTD data and related reports are available from <http://www.ideo.columbia.edu/~claudiag/ASEP>, <http://www.marine-geo.org>, and the National Oceanographic Data Center. Bathymetric measurements have been used in the “RTOPO” data set, <http://doi.pangaea.de/10.1594/PANGAEA.741917>.

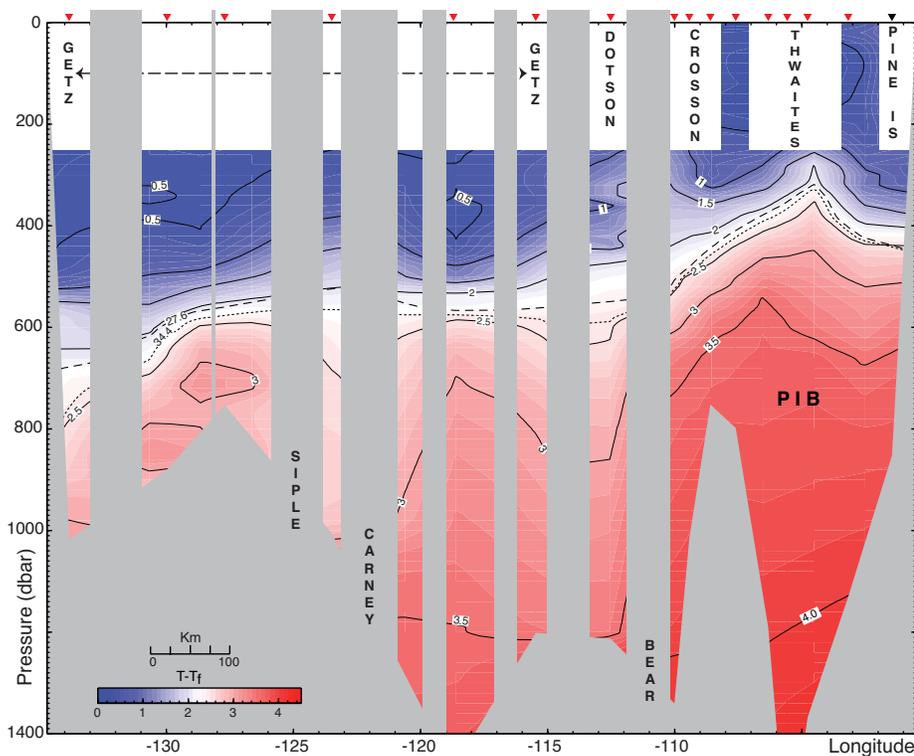


Figure 2. Temperature above the pressure freezing point, derived from a 2° longitude x 20 dbar gridding of late summer 2007 (2009) CTD profiles (▼) from the western Getz to the Pine Island Ice Shelf (PIIS). Casts west of Bear Island were within 500 m of ice fronts along the southern Amundsen coastline; to the east, they swing north of 74°S and follow Pine Island Bay (PIB) trough to the PIIS (Figure 1). The schematic ice shelves with nominal 250 m ice front drafts thicken toward their grounding lines, and are pinned by islands and other landforms that partially constrain the sub-ice ocean circulation. The near coincidence of isotherms, isohalines, and isopycnals, illustrated by selected salinity and density anomaly contours, allows much of the subsurface ocean structure to be tracked by temperature alone.

THERMOHALINE VARIABILITY

Temperatures in Figures 2–4 are plotted relative to the in situ seawater freezing point, which declines with increasing pressure and salinity. That is also the pressure melting point of ice in seawater, and the figures are color-coded to depict its melting potential with depth and location. Beneath the surface layers, temperature correlates positively with salinity, which dominates the density field (examples in Figure 2) and controls vertical stability in polar seas. The highest temperatures are in the eastern sector, separated from the central Amundsen by a shallow bank extending and deepening north and northwest of Bear Island (Figure 1). A maximum $T-T_f$ (temperature above freezing) of 4.34°C was recorded near 1,625 m seaward of the Thwaites Glacier Tongue in 2007, but it is not yet known whether seawater that warm has access to a deep Smith Glacier/Crosson Ice Shelf grounding line. Cooler deep water appears near the gap between Siple and Carney Islands, and near

the western end of the Getz Ice Shelf. Contours generally deepen west of PIB, with the CDW becoming less voluminous beneath a well-developed thermocline that grades upward into the colder and fresher Antarctic Surface Waters.

North-south sections between the continental slope and ice shelf fronts are separated by shallower regions that will constrain east-west flow at depth (Figure 1; Figure 3a vs. 3b). To a lesser degree, the submarine bank seaward of Bear Island will also inhibit and modify shallower flow by grounding icebergs along its eastern side. Those bergs anchor the frequent Thwaites fast ice extension noted above, which in turn facilitates the persistent Amundsen Polynya north of Dotson Ice Shelf. Although the largest volume of warmest CDW occurs on the wider eastern shelf, Figure 3c shows relatively warm deep water reaching the Getz Ice Shelf west of Siple Island. Isotherms that deepen southward across the shelf and at the northern ends of most sections imply westward flow and stronger currents along the continental shelf break (Figure 3b–d). Slightly lower deep temperatures over the outer shelf suggest stronger vertical mixing in that area and narrow or episodic inflows. A cyclonic gyre or large eddy can sometimes be identified at the sea surface in PIB by the relative warmth of buoyant, melt-laden CDW from beneath the PIIS (Mankoff et al., 2012), centered along the inner PIB trough (Figure 2, Figure 3a). Strong ocean property gradients along the 27.6 isopycnal have marked the southern boundary of the Antarctic Circumpolar Current (ACC) north of the continental shelf on most circumpolar meridians (Orsi et al., 1995). That definition falters in the Amundsen and Bellingshausen Seas, where temperatures

above 1°C and dissolved oxygen less than 5 ml L⁻¹ can extend hundreds of kilometers to the coastline (Figure 3).

Sea ice prevented the *Palmer* from reaching the PIIS in 2000 and 2007, but most of the continental shelf break region was defined for the first time during those cruises (Nitsche et al., 2007). CTD casts show the cold and fresh upper water column deepening westward along

the outer shelf, as along the coastline in Figure 2, coincident with a decreasing presence of CDW (Figure 4). The thick, cold surface layers shield the persistent pack and fast ice fields from the warmth of the underlying deep water. The CDW was warmer in 2007, with its temperature maximum often well above outer shelf depths in both years, and strongest above the eastern sides of seafloor

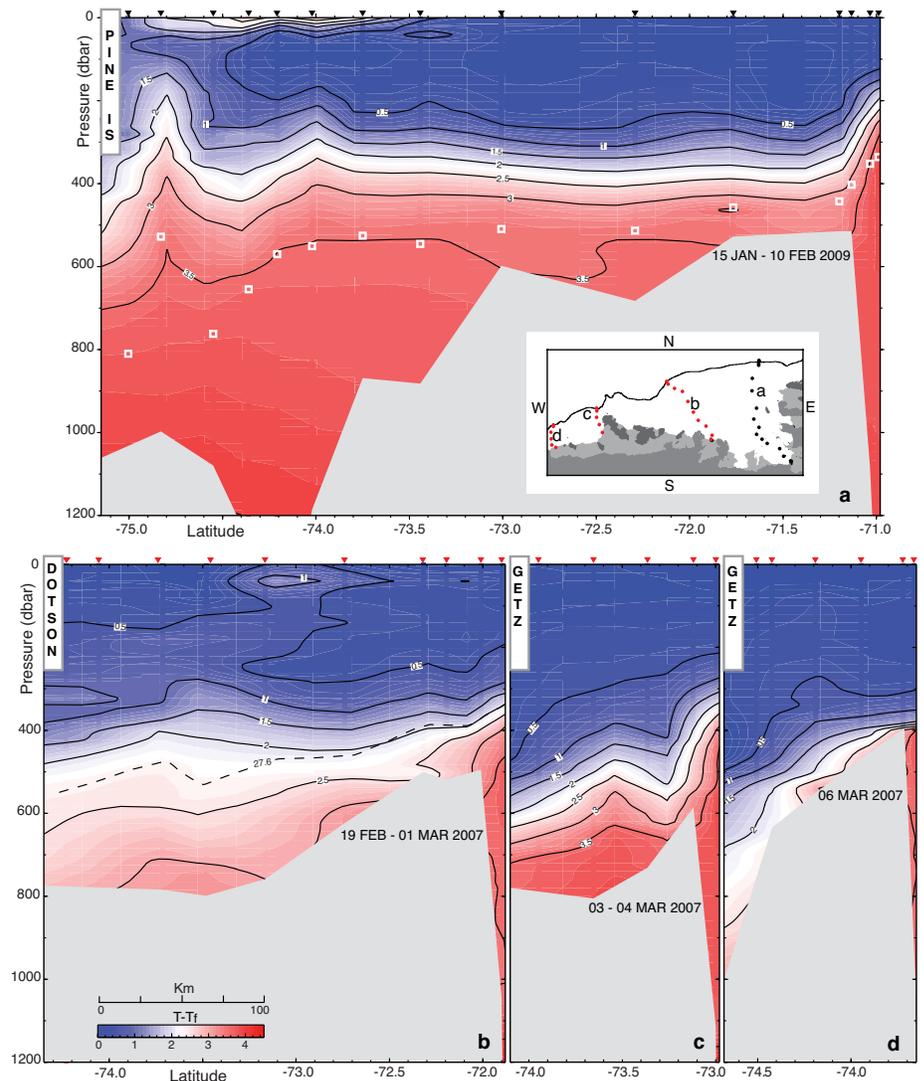


Figure 3. Temperature above freezing along north-south transects across the Amundsen continental shelf in early 2009 (a) and 2007 (b–d), from gridded CTD profiles as in Figure 2. These sections roughly follow troughs that deepen southward from the outer/mid-shelf toward ice shelf fronts at the coastline (Figure 1). The dissolved oxygen minimum (white boxes) at each CTD in section (a) is characteristic of upper Circumpolar Deep Water (CDW), varies little from 4.29 ml L⁻¹ across the shelf, and parallels a mid-depth CDW temperature maximum (Jacobs et al., 2011), suggesting isopycnal inflows well above the seafloor. The center of a cyclonic gyre near the Pine Island Ice Shelf (a) appears more to the northeast in Pine Island Bay trough in 2007 (Figure 2).

depressions where inflows may be concentrated. Some of those greater depths may have been glacially scoured, but the bathymetry typically shoals southward before reaching the deeper and rougher inner shelf troughs (Figure 1; Evans et al., 2006). The modeling work described below indicates enhanced CDW inflow associated with greater shelf break depths, which may also influence the scales of frontal zone eddies, as suggested by the thermocline structure near the center of Figure 4a.

Well before supporting data had been acquired from the Amundsen Sea, it was presciently inferred that modified warm deep waters dominated the Antarctic continental shelf in the Pacific sector (Carmack, 1977). We can now see that the deepwater intrusions are spatially variable in summer along the southern coastline (Figure 2), and temporally variable from a representative subset of 2007 and nearby 1994 and 2000 profiles (Figure 5). Eastern Amundsen

water columns are homogeneous below ~ 700 dbar, but thermoclines reach the seafloor at shallower stations in the western sector. Generally warmest in 2007 and coldest in 2000, the degree of CDW modification increases westward and its overlying thermoclines and haloclines are shallower when the deep water is warmer. The coastal Amundsen profiles are also compared in Figure 5 with examples from upstream on the Antarctic Peninsula and downstream in the Ross Sea. In Marguerite Bay ($\sim 70^\circ\text{W}$ near the northern George VI Ice Shelf front), deep thermohaline properties resemble those in the eastern Amundsen, but extend much higher in the water column beneath a less-variable thermocline and fresher surface water. Near the Ross Ice Shelf ($\sim 168^\circ\text{E}$ off Ross Island), a thick layer of saline shelf water at temperatures near the sea surface freezing point lies beneath highly modified CDW and surface waters. That wide range of properties along the

Pacific-Antarctic coastline is atypical for the Antarctic continental shelf, most of which is similar to the Ross Sea.

The thermohaline field near southern Amundsen Sea ice shelves is largely defined by mixing between surface and deep waters, the summer properties of which have varied considerably over the brief time they have been observed (Figure 6). CDW and its modified forms, mostly saltier than 34.5 and warmer than 0.7°C , have differed by more than 1.0°C near the seafloor. Antarctic Surface Water, mostly fresher than 34.2 and colder than -1.2°C , includes temperature minima set at the winter sea surface and beneath the shelf ice, but rarely reaching the surface freezing line. The upper three-quarters of the thermocline can encompass the entire water column and falls into the Antarctic Surface Water category of Whitworth et al. (1998). And, despite high rates of ice shelf melting in this sector, the ocean circulation is strong enough that most of the

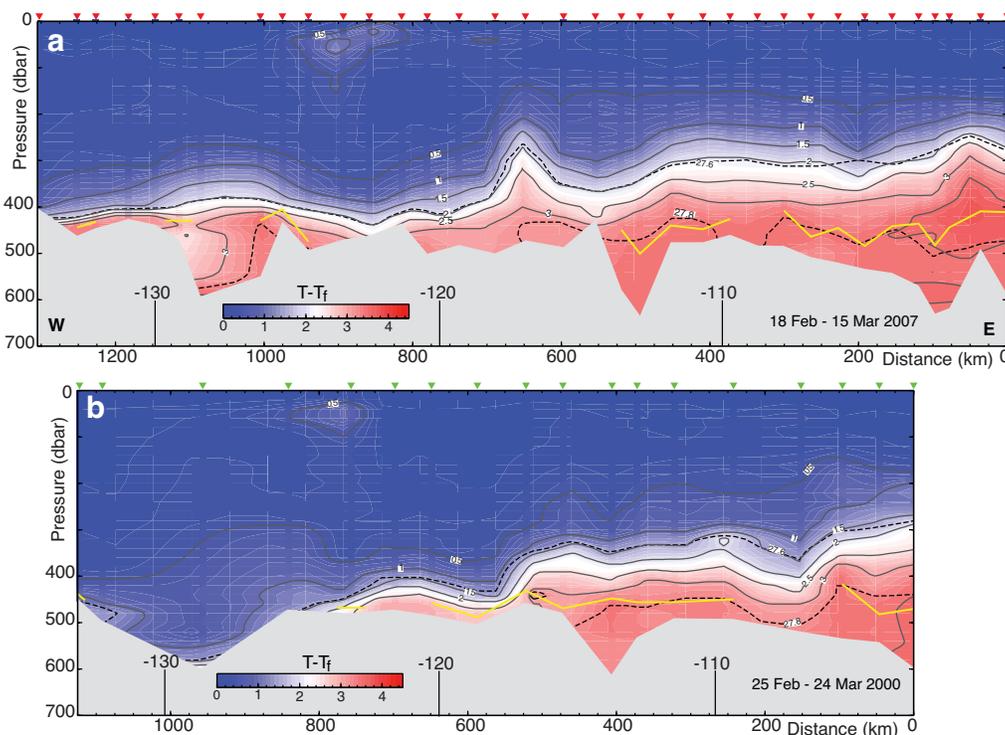


Figure 4. Temperature above freezing and representative isopycnals along the outer continental shelf in 2007 (a) and 2000 (b), derived from $50 \text{ km} \times 20 \text{ dbar}$ gridding of CTD profiles (\blacktriangledown , \blacktriangledown , and Figure 1). The 2007 transect extends ~ 30 (80) km farther west (east) than in 2000, with yellow lines showing potential temperature maxima above the seafloor. Shifting local winds were accompanied by a 50–100 m shallower thermocline and a 0.3°C higher temperature maximum on profiles < 5 days after the 2007 CTD near 200 km, providing a short-term example of the high variability of CDW inflow across a weak Antarctic Slope Front.

subsurface meltwater content is low, for example, averaging $0.7 \pm 0.3\%$ from 100–800 dbar at the central 2007 profiles in Figure 5. CDW inflow, melt-driven upwelling, and the outflow of surface and thermocline waters characterize large-scale vertical circulation on the Southeast Pacific–Antarctic continental shelf. In contrast, surface and intermediate waters are imported and sea ice and shelf water exported on the downwelling Ross Sea shelf, forced by stronger sea ice production.

From March 2006 through February 2007, CDW temperatures near the bottom in PIB were $\sim 0.7^\circ\text{C}$ higher and less variable than at similar depths near the Dotson and eastern Getz Ice Shelves (Figure 7). Greater variability in the central Amundsen may result in part from changing CDW access to the continental shelf west of $\sim 115^\circ\text{W}$, as its overlying pycnocline fluctuates around the outer shelf depths (Figure 4). Surprising differences between DOT 851 and GTZ 781, both linked to the same mid-shelf trough (Figure 1; Larter et al., 2009), suggest that factors beyond bathymetric control influence deep temperatures near those ice shelves. On the outer continental shelf west of the Antarctic Peninsula, CDW temperatures were $\sim 0.3^\circ$ higher than in PIB over the same period of time. At both sites, bottom temperatures were more than 3° higher than the near-freezing shelf water off the Ross Ice Shelf, as recorded in 1984–1985 and little changed in 1994 and 2007 profiles (Figure 6).

CDW ACCESS TO THE CONTINENTAL SHELF

Recent Amundsen Sea cruises have focused on the continental shelf region, although several long CTD sections

have been occupied to its north from the *Palmer* and other vessels (e.g., Hellmer et al., 1998; Swift and Orsi, 2012, in this issue). Continental shelf waters are close relatives of their off-shelf counterparts, separated only by a weak Antarctic Slope Front (Jacobs, 1991), typically manifested as a landward deepening of the pycnocline separating CDW from surface waters over the continental slope (Figure 3) and a cold, fresh, westward shelf-break current. Unlike the full-depth expression of the slope front in the eastern Weddell (Whitworth et al., 1998; Heywood et al., 1998), the eastern Amundsen pycnocline is generally found above the level of the 450–600 m shelf break (Figure 4). CDW and portions of the overlying water column are believed to intrude onto the shelf more strongly through outer shelf depressions that are linked to the deep inner shelf troughs (Walker et al., 2007; Wähler et al., 2010). As these CDW inflows provide most of the ocean heat that melts the ice shelves

farther south, understanding the processes that regulate their strength and variability is critical to assessing ocean forcing on ice sheet evolution.

While CDW properties are similar on the continental shelves of the Amundsen Sea and West Antarctic Peninsula (Figure 5), subtly different processes deliver CDW to these shelf regions. The peninsula protrudes into the region of climatological westerlies, the ACC hugs its steep continental slope, and a slope front is lacking (Orsi et al., 1995; Whitworth et al., 1998). Instabilities and eddy activity associated with interactions between the eastward-flowing ACC and shelf edge topography appear responsible for on-shelf transport of CDW (Dinniman and Klinck, 2004; Moffat et al., 2009). In contrast, the Amundsen outer shelf and continental slope lie in a region of climatological easterlies and the ACC tracks north of the slope until reaching the area of Thurston Island (Orsi et al., 1995). The iceberg-infested

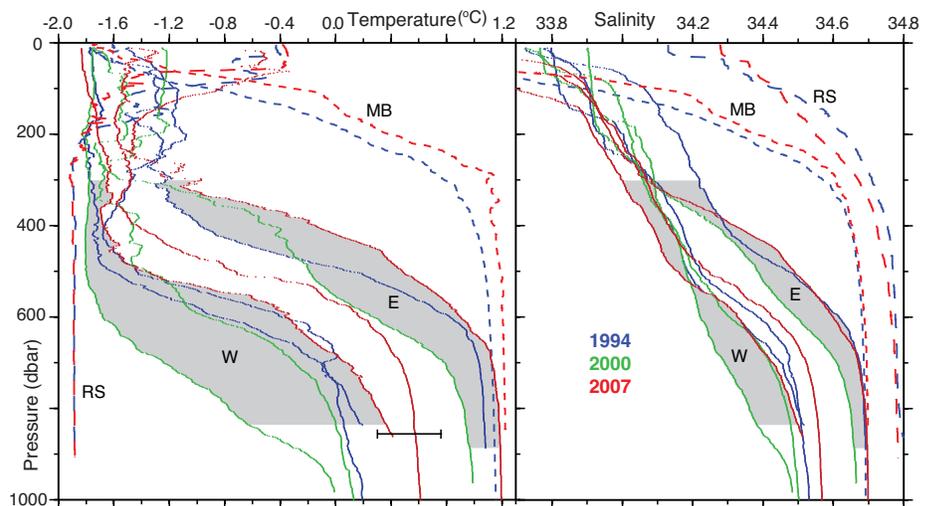


Figure 5. A subset of the Figure 2 profiles, with in situ temperature and salinity averages showing that the volume and thermohaline properties of deep water near ice shelves in the Amundsen Sea were relatively low in 2000 and high in 2007. Each of the shaded Amundsen profiles are averaged from two CTDs east of Bear and west of Siple Islands, between which the central region 2007 average is overlain by the 2006–2007 temperature range at 851 m from Figure 7. For comparison with other sectors of the Pacific–Antarctic continental shelf (Figure 1 inset), the dashed profiles are from single, closely spaced CTDs in Marguerite Bay (MB) and near Ross Island (RS).

region between the ACC and the shelf break may be characterized by weak and variable currents, and occupied by intermittent extensions of the Ross Gyre (Assmann and Timmermann, 2005).

Prevailing southeast winds force a westward drift circulation over much of the Antarctic continental shelf and slope, accompanied by downwelling that can enhance the slope front near the continental shelf edge and where the edge of the ice sheet is steep (Nøst et al., 2011). Steeply sloping isopycnal surfaces with high vertical shear correspond to a weakening of the westward flow with depth and a current reversal on the continental slope below the level where the deepening pycnocline intersects the rising topography (Heywood et al., 1998; Chavanne et al., 2010). In the Amundsen Sea, the ice sheet edge is less steep and mostly farther south on the continental shelf, with the pycnocline

only intersecting shallower parts of the seabed (Figure 4). Analyses of CTD sections across that break have also indicated westward surface flow yielding to an eastward undercurrent beneath the pycnocline (Walker et al., 2007, 2012), but located above the upper continental slope and on the edge of the continental shelf. The flow weakens as it encounters a deeper shelf break, consistent with a southward turn carrying CDW onto the shelf. In a domain without the then poorly mapped deeper eastern shelf break trough (Figure 1), inflow near 114°W turned east and joined the main trough leading into PIB as described in Walker et al. (2007) and Thoma et al. (2008). In the latter study, variability in wind forcing on seasonal and interannual timescales drove fluctuations in the depth-mean flow at the shelf edge, modifying the strength of the undercurrent and associated on-shelf transport. Both

were enhanced during winter and spring when the wind stress had a stronger and more persistent westerly component.

Atmospheric forcing near the shelf edge of the eastern Amundsen Sea thus seems intermediate in character between regions influenced by persistent circumpolar westerlies or coastal easterlies. With a weak slope front and distant ACC, zonally variable wind stress may be the main influence on CDW inflow fluctuations. Regional atmospheric variability is primarily linked to changes in position and strength of the Amundsen Sea Low, in part internally generated or locally forced by changes in stratospheric ozone (Turner et al., 2009). Remote forcing from the central tropical Pacific may also play a role, as anomalously high sea level pressure in the central Amundsen is associated with tropical Pacific warm phases with increased atmospheric convection generating tropospheric height anomalies during the Southern Hemisphere winter (Ding et al., 2011). That pattern favors a westward shift of the Amundsen Low to the Ross Sea, westerly wind stress over the continental shelf edge, and stronger CDW inflow (Thoma et al., 2008). West Antarctic Ice Sheet changes may thus be sensitive to trends and extremes in sea surface temperature in the central tropical Pacific (Steig et al., 2012).

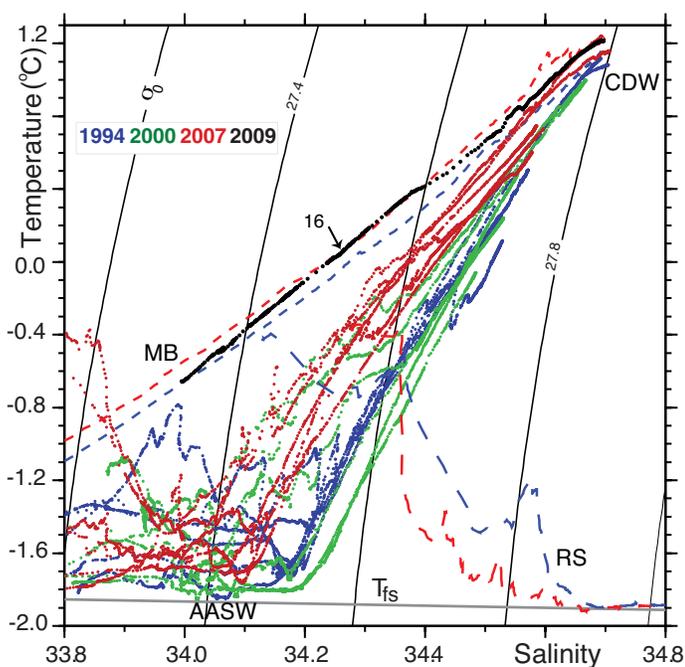


Figure 6. Temperature/salinity/density diagram of the individual CTD profiles from Figure 5 illustrate mixing between the two dominant coastal Amundsen Sea water masses, and their variability in 1994, 2000, and 2007. Subsurface thermohaline properties are mainly set by mixing between Circumpolar Deep Water (CDW) and the “winter water” base of the Antarctic Surface Water (AASW). Values bow slightly toward an ice shelf melt/CDW mixing line—for example, from 2009 CTD 16

at the site of the primary outflow from beneath the Pine Island Ice Shelf—indicative of meltwater being fed into thermocline and surface waters. The Marguerite Bay (MB) profiles in 1994 and 2007 were also in a strong outflow and upwelling region near George VI Ice Shelf (Jenkins and Jacobs, 2008). Water column properties in the Ross Sea (RS) are dominated by ice formation at the sea surface freezing temperature (T_{fs}).

NUMERICAL MODELING AND OCEAN-ICE INTERACTIONS

Initial modeling in the Amundsen Sea focused on the PIIS and considered flow and ocean-ice interactions along a vertical section in the poorly defined sub-ice cavity (Hellmer et al., 1998).

A mean (maximum) basal melt rate of 12 (30) m yr^{-1} exceeded the prior highest values of ~ 2.1 (3.6) m yr^{-1} under George VI Ice Shelf (Potter and Parren,

1985), results that were soon eclipsed by higher estimates from satellite and airborne observations (Rignot, 1998; Corr et al. 2001; Shepherd et al., 2004). A reduced gravity dynamics plume model tuned for a mean melt rate of $\sim 26 \text{ m yr}^{-1}$ indicated melting over 100 m yr^{-1} for thick ice near the grounding line (Payne et al., 2007). Mixing and melting processes did not utilize all of the available CDW heat in those models, consistent with buoyant, above-freezing outflows that correlate with small polynyas observed at the ice front (Payne et al., 2007; Bindschadler et al., 2011; Mankoff et al., 2012). Idealized models of melt rate sensitivity to ocean temperature, velocity, tides, and ice shelf basal slope suggest an overall quadratic response to warming, but less-efficient use of ocean heat as ocean temperature rises (Holland, 2008; Holland et al., 2008; Little et al., 2009). Using an adjoint ocean model capturing sub-ice processes, Heimbach and Losch (2012) also demonstrated circulation sensitivity to seafloor topography, and connections between the PIIS calving front and cavity interior on timescales of 30–60 days.

The Thoma et al. (2008) model related heat content variability in PIB to wind-driven fluctuations in CDW inflow modified during its trough-guided southward propagation. A higher-resolution coupled sea ice-ocean model (Schodlok et al., 2012) with updated bathymetry showed the main CDW on-shelf intrusions occurring through the deeper and larger eastern trough near 102.5°W in Figure 1. They also found interannual variability of area-averaged PIIS basal melting of $\sim 22\text{--}32 \text{ m yr}^{-1}$ from 1979–2010, mainly due to fluctuating strength of the PIB gyre. Anomalously low temperatures and a low melt rate for the PIIS appear

to result from the atmospheric forcing and sea ice representation in PIB with FESOM, a global coupled ice-ocean finite element model with eddy permitting resolution (5–10 km) near Antarctica (Timmermann et al., 2012). That model gives a melt rate of 5.4 m yr^{-1} for the larger Getz Ice Shelf, where cavity inflows and outflows are spread among several openings (Figure 2).

CONCLUDING REMARKS

During past ice ages, the ice streams of a larger WAIS carved deep troughs into the seafloor of the continental periphery, potentially setting the stage for their eventual demise. During the last two decades, we have found that warm deep water now floods those troughs on the Southeast Pacific continental

shelf, thinning Amundsen ice shelves and accelerating its glaciers, with possible implications for ice sheet stability and sea level rise. Much remains to be learned about the roles of cyclical to tropical atmospheric forcing on CDW access to the Antarctic continental shelf, the influence of ocean temperature and circulation variability on ice shelf mass balance, and controls that the seafloor and shelf ice exert on ice stream flow. Records are still short and sporadic for the Amundsen Sea, as its nearly perennial sea ice cover often foils expeditionary plans, but fieldwork is essential and should include representative time-series measurements for assimilation into high-resolution “ocean reanalyses” of the continental slope, shelf, and ice shelf cavity circulations. Existing satellite records

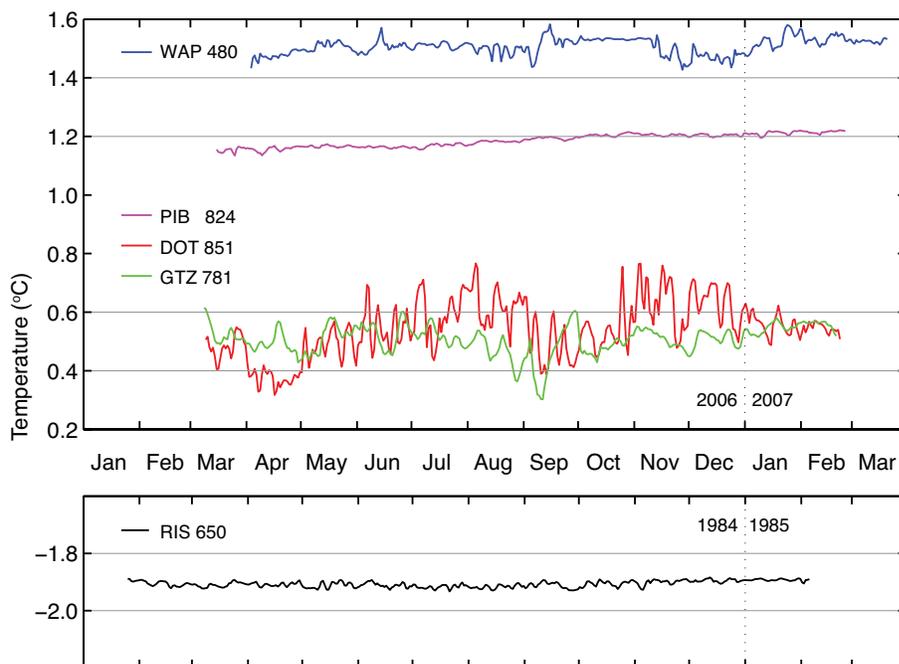


Figure 7. Daily average temperatures from Sea-Bird instruments moored near the bottom at the Amundsen sites in Figure 1 (center), at 67.677°S , 73.328°W and 480 m west of the Antarctic Peninsula (top), and at 77.882°S , 178.533°W and 650 m near the Ross Ice Shelf (RIS; bottom). While fortnightly tidal cycles are prominent in the Amundsen records, seasonality is limited to a slight rise in temperature during winter at 824 m in Pine Island Bay (PIB), lower temperatures in April and September at 851 m off the Dotson Ice Shelf (DOT), and at 781 m in September off the eastern Getz Ice Shelf (GTZ). The $\sim 0.45^\circ$ annual temperature range at DOT 851 was recorded at a depth where the mean of 2007 summer profiles averaged $\sim 0.6^\circ$ (Figure 5), in turn corresponding to $\sim 3^\circ$ above the in situ freezing point in Figure 2.

must be extended to provide the spatial coverage and continuity of sea and shelf ice observations that ships cannot. The upper deep water is warming (Levitus et al. 2009), the modeled continental shelf circulation is sensitive to sea ice production (Hellmer et al., 2012), and the response of deep-draft ice to ocean temperature may not be linear (Rignot and Jacobs, 2002; Holland et al., 2008). From the oceanography perspective, it is also not just about what the sea is doing to the ice, as increased melting adds to freshening and stratification, altering the properties of shelf and bottom waters and quite likely the rates of deep ocean convection (Jenkins, 1999; Hellmer, 2004; Jacobs and Giulivi, 2010).

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