From circumpolar deep water to the glacial meltwater plume on the eastern Amundsen Shelf

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ABSTRACT

The melting of Pine Island Ice Shelf (PIIS) has increased since the 1990s, which may have a large impact on ice sheet dynamics, sea-level rise, and changes in water mass properties of surrounding oceans. The reason for the PIIS melting is the generally warmer (~1.2°C) Circumpolar Deep Water (CDW) that penetrates into the PIIS cavity through submarine glacial troughs located on the Amundsen Sea continental shelf. In this study, we mainly analyze the hydrographic data obtained during ANTXXVI/3 in 2010 with the focus on pathways of the intruding CDW, PIIS melt rates, and the fate of glacial meltwater. We analyze the data by dividing CTD profiles into 6 groups according to intruding CDW properties and meltwater content. From this analysis, it is seen that CDW warmer than 1.23°C (colder than 1.23°C) intrudes via the eastern (central) trough. The temperature is controlled by the thickness of the intruding CDW layer. The eastern trough supports a denser CDW layer than the water mass in Pine Island Trough (PIT). The eastern intrusion is modified on the way into PIT through mixing with the lighter and colder CDW from the central trough. Using ocean transport and tracer transport calculations from the ice shelf front CTD section, the estimated melt rate in 2010 is ~30 m yr⁻¹, which is comparable to published values. From spatial distributions of meltwater content, meltwater flows along the bathymetry towards the west. When compared with earlier (2000) observations, a warmer and thicker CDW layer is observed in Pine Island Trough for the period 2007–2010, indicating a recent thickening of the CDW intrusion.

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1. Introduction

Pine Island Glacier (PIG) and Thwaites Glacier (TG) are the major ice streams of the West Antarctic Ice Sheet (WAIS), draining around 393,000 km², about 25% of the total area, and driving the largest ice loss of the Antarctic ice sheet (e.g. Shepherd et al., 2001; Rignot et al., 2002, 2008, 2011; Wingham et al., 2009; Joughin and Alley, 2011; Shepherd et al., 2012). The observed ice velocity of the PIG has increased since the 1990s even far inland (exceeding 3.5 km yr⁻¹ close to the grounding line), leading to the thinning and further acceleration of the PIG (Shepherd et al., 2001; Vaughan et al., 2006; Rignot et al., 2011).

The reason for the large ice loss is increased basal melting due to the interaction of deep-draft ice shelves with Circumpolar Deep Water (CDW) which is about 3°C warmer than the surface freezing point (e.g. Jacobs et al., 2011; Pritchard et al., 2012). This relatively warm water flows onto the continental shelf through submarine glacial troughs (Jacobs et al., 2011). Possible pathways of the intrusion have been studied. Walker et al. (2007) reported onshore flow of CDW through the central trough (C, Fig. 1). However, measurements were only conducted at the entrance of trough C due to heavy sea-ice conditions further south, and it is still unknown whether CDW penetrates all the way south to the Pine Island Ice Shelf (PIIS) cavity from trough C. Numerical model results with the most realistic topography available (Nitsche et al., 2007) show that CDW enters the continental shelf mainly through the eastern trough (E, Fig. 1) as a result of an onshore baroclinic eddy transport of warm water and reaches the PIIS cavity (Schodlok et al., 2012). However, there has not been any observation determining the pathway of CDW intrusions.

When this warm water reaches the grounding line, it melts the glacier and forms a buoyant plume of CDW and meltwater. This plume flows out from the ice shelf cavity and drives the continuous intrusion of CDW into the PIIS cavity where it follows an almost clockwise circulation (Payne et al., 2007). The meltwater outflow along the PIIS front is observed at the depth of 100–500 m (Jacobs et al., 1996; Hellmer and Jacobs, 1998; Jenkins et al., 2010; Jacobs et al., 2011). Most recent results show that the averaged melt rate in 2009 is estimated to be about 30 m yr⁻¹ (Jacobs et al., 2011) and melting close to the grounding line can be as high as 80–100 m yr⁻¹ (Payne et al., 2007; Heimbach and Losch, 2012).

The melting of PIIS and other ice shelves of the WAIS can have large impacts on the global ocean. First, the WAIS has the potential
to raise global sea level by 3.3 m (Bamber et al., 2009), and 10% of the observed sea level rise has been attributed to the thinning of the WAIS (Rignot et al., 2008). Second, it may cause the freshening of the shelf water locally in the Amundsen Sea as well as remotely in the Ross Sea (Jacobs et al., 2002). This may lead to a change in the characteristics of the Antarctic Bottom Water (AABW) formed in the Ross Sea (Jacobs et al., 2002; Rintoul, 2007) and thus may influence the global thermohaline circulation. Therefore, investigations related to the PIIS melting and its impact on the ocean is crucial for understanding climate change in the Southern Ocean.

Due to an extremely low sea-ice coverage on the whole Amundsen Sea continental shelf in austral summer 2010, we were able to conduct CTD measurements during ANTXXVI/3 in its whole eastern region, including the two submarine glacial troughs that may channel CDW beneath the PIIS. Such data is essential to determine the routes of CDW into the PIIS cavity and to detect freshwater input due to ice shelf basal melting. In this study, we analyze hydrographic data focusing on the pathway of CDW into the PIIS cavity, the PIIS melt rate in 2010, and the spreading of the meltwater plume on the Amundsen Sea continental shelf.

2. Data and methods

Sampling was carried out during ANTXXVI/3 from the research ice breaker Polarstern (Gohl, 2010). In total 62 CTD profiles shown in Fig. 1 were collected on this cruise including one Heli-CTD measurement.

The measurements use a Seabird 911+ CTD (SN 561) connected to a caroussel (SBE 32, SN 202) with 24-(12-l) water bottles. This instrument system contains two sensor pairs of conductivity (SBE 4, SN 3607, SN 3590) and temperature (SBE 3, SN 1373, SN 2629), a high precision pressure sensor Digiquartz 410K-105 (SN 68997), one oxygen sensor (SBE 43, SN 743 until station no. 143 and SN 880 from station no. 147 on), one oxygen sensor (Rinko SN 10, optode), a fluorometer (Wetlab ECO-AFL/FL, SN 1365), and a Benthos altimeter Model PSA 916 (SN 1228).

For the identification of meltwater, a few helium and neon samples were taken from the water samples of the CTD casts at the PIIS front section (Fig. 1). They were all sealed in copper tubes and analyzed by mass spectrometer at the IUP Bremen (Sültenfuß et al., 2009). The accuracy of measurements is 0.01 nmol kg$^{-1}$ or 1% for He and 0.04 nmol kg$^{-1}$ or 1% Ne, determined from the large number of observations in the western Weddell Sea (Huhn et al., 2008).

In the latter part of the analysis, Nathaniel B. Palmer CTD data from 2000, 2007, and 2009 is used to discuss the temporal variabilities. All data sets we use are taken in austral summer between January and...
March, because no observations have been conducted in winter time in the Amundsen Sea up to now.

3. Results

3.1. Observational results

The potential temperature, salinity, and dissolved oxygen for Sections 1 and 2 (Figs. 2 and 3) mainly show Winter Water (WW) with the potential temperature minimum at 100–300 m and warm CDW from 300 to 400 m to the bottom. The north–south section (Fig. 4), which connects the deepest stations of each east–west section (Fig. 1), shows that CDW clearly dominates the lower part of the water column, and that CDW is continuously supplied to the PIIS cavity. A similar result is presented in Jacobs et al. (2011) from their observations in 2009. Along the surface-referenced isopycnal of 27.7, CDW with the potential temperature of $1.1^\circ$C, salinity of 34.5, and dissolved oxygen of 4.2 ml kg$^{-1}$ flows onto the continental shelf and into the PIIS cavity. The maximum isopycnal that reaches the PIIS is $\sim 27.79$. The 27.8-isopycnal advances to the center of a bathymetric trough, which extends to the PIIS and is colored in yellow and green in Fig. 1. Following Jakobsson et al. (2011), we name this trough Pine Island Trough (PIT). The deep water column at the PIIS front is filled with CDW, overlain by the outflowing mixture of meltwater and CDW (Fig. 5).

3.2. Typical $\theta$–$S$ profiles observed on the eastern Amundsen Sea continental shelf

To analyze the intrusion of CDW and spreading of meltwater, we have color-coded the CTD stations on the Amundsen Sea continental shelf. We use maximum potential temperature ($T_{\text{max}}$) and specific meltwater fraction (0.6%), calculated using potential temperature and salinity, to define the color-codes (Table 1). For the calculation of the meltwater fraction (see Appendix A), water mass characteristics are defined for CDW, WW, and meltwater following Jenkins (1999), Jenkins and Jacobs (2008), and Jacobs et al. (2011), summarized in Table 2.

Since the $\theta$–$S$ profiles of color-coded stations (Fig. 6) represent water columns from 100 m to the bottom, the shallower parts represent WW properties while the deeper parts show CDW properties. The profiles with meltwater fraction higher than 0.6% are color-coded green and the ice front profiles are color-coded cyan (Table 1). For all the non-green and non-cyan profiles, two different regimes are observed; $\theta$–$S$ profiles north of 73.34$^\circ$S show large spatial variability which is much smaller for $\theta$–$S$ profiles south of 73.34$^\circ$S. Thus, we define all the non-green and non-cyan profiles south of 73.34$^\circ$S (dashed blue line in Fig. 1) as the blue group, representing the typical profiles of the PIT (Fig. 7a). Other northern profiles are color-coded red, pink, or orange according to the $T_{\text{max}}$ (Table 1 and Fig. 7b–d). From the definition of the color-code (Table 1), the green and cyan profiles have higher meltwater fraction at shallower depth than the blue profiles, but their CDW properties are similar (Fig. 7e and f). The blue profiles have the maximum potential temperature of $-1.13$–$1.23^\circ$C (Fig. 6). The red (orange) group has warmer (colder)
CDW properties (Fig. 7b and c) than the blue and pink groups, but their WW properties are similar to those of the blue group.

Note that we only focus on the profiles from the eastern part of the continental shelf. The three stations in the western submarine glacial trough are not included in this analysis (gray dots in Fig. 8). The Heli-CTD station is also not used because it has a thick WW layer from ~100 m almost to the bottom, and thus is different from all other profiles (Fig. 3).

For comparison, we also color-coded the profiles of 2007 and 2009 (Fig. 8b). Since the color-code is defined by the potential temperature and the meltwater fraction, the location of each group demonstrates the pathway of the intrusion of CDW and the spreading of meltwater on the eastern Amundsen Sea continental shelf (Fig. 8a). The pattern of the color-code in 2010 and 2007–2010 is similar (Fig. 8a and b), suggesting that the feature of the intrusion of CDW and the spreading of meltwater have not changed significantly during this period.

4. Discussion

4.1. The reason for the warmer and colder CDW intrusion

Two submarine glacial troughs (marked E and C in Fig. 1) exist which may preferentially guide CDW onto the shelf towards the PIIS cavity (Walker et al., 2007; Thoma et al., 2008; Schodlok et al., 2012).

From the observations in 2010, the presence of CDW with potential density higher than 27.7 is observed on the eastern and western sides of Section 1 (Fig. 2). In Section 2, CDW is also observed but only on the eastern side (Fig. 3). From the locations of the color-coded stations (Fig. 8), warmer CDW (red, $1.23 \degree C < T_{\text{max}}$) penetrates via trough E, and colder CDW (pink and orange, $T_{\text{max}} < 1.23 \degree C$) flows via trough C onto the continental shelf. In this subsection, we focus on the profiles on the continental shelf north of 73.34°S, especially the red and orange.

### Table 1

**Definition of the color code used for the eastern Amundsen Sea continental shelf.**

<table>
<thead>
<tr>
<th>Color</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cyan</td>
<td>Profiles observed at PIIS front</td>
</tr>
<tr>
<td>Green</td>
<td>Meltwater fraction exceeding 0.6% in the salinity range of 34.0–34.4</td>
</tr>
<tr>
<td>Blue</td>
<td>Non-green and non-cyan profiles south of 73.34°S</td>
</tr>
<tr>
<td>Red</td>
<td>Non-green profiles north of 73.34°S and $1.23 \degree C &lt; T_{\text{max}}$</td>
</tr>
<tr>
<td>Pink</td>
<td>Non-green profiles north of 73.34°S and $1.13 \degree C &lt; T_{\text{max}} &lt; 1.23 \degree C$</td>
</tr>
<tr>
<td>Orange</td>
<td>Non-green profiles north of 73.34°S and $T_{\text{max}} &lt; 1.13 \degree C$</td>
</tr>
</tbody>
</table>

### Table 2

**The water mass characteristics of CDW, WW, and meltwater.** The water mass characteristics of CDW and WW in potential temperature, salinity and dissolved oxygen are chosen such that the CDW/WW mixing line is the same as in Jacobs et al. (2011).

<table>
<thead>
<tr>
<th>Characteristics</th>
<th>CDW</th>
<th>WW</th>
<th>MW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Potential temperature (°C)</td>
<td>1.18</td>
<td>~1.8</td>
<td>~90.8</td>
</tr>
<tr>
<td>Salinity (psu)</td>
<td>34.7</td>
<td>34.0</td>
<td>0.0</td>
</tr>
<tr>
<td>Dissolved Oxygen (ml kg⁻¹)</td>
<td>4.18</td>
<td>6.83</td>
<td>28.5</td>
</tr>
<tr>
<td>Helium concentration (nmol kg⁻¹)</td>
<td>1.85</td>
<td>1.86</td>
<td>25.7</td>
</tr>
<tr>
<td>Neon concentration (nmol kg⁻¹)</td>
<td>8.03</td>
<td>8.22</td>
<td>89.2</td>
</tr>
</tbody>
</table>
groups, and investigate why different potential temperatures are observed for the CDW intrusions from troughs E and C.

The vertical profiles of potential temperature and potential density of the red, orange, and blue groups (Fig. 9) show fairly uniform CDW properties below a CDW/WW thermocline (where potential temperature increases rapidly with depth). Since the $T_{\text{max}}$ is observed just below the WW/CDW thermocline, we define the intruding CDW layer as the depth range between the $T_{\text{max}}$ and the bottom. A thick CDW layer is observed for the red group (100–300 m) in contrast to a thin CDW layer at the bottom for the orange group (Fig. 9). The offshore profiles show that CDW temperature decreases with increasing potential density within the range of 27.76 to the bottom. Thus, the difference in $T_{\text{max}}$ can be explained by the different thicknesses of the CDW intrusion; the thin intrusion from trough E contains light and warm (red, 1.23°C < $T_{\text{max}}$) CDW as well as dense and cold (pink and orange, $T_{\text{max}}$ < 1.23°C) CDW, while the thin intrusion from trough C contains only dense and cold CDW (Fig. 10). We define the potential density range of the intruding CDW as the density range between the $T_{\text{max}}$ and the bottom. Although the variability of the red and orange groups is high, the potential density of the intruding CDW at $T_{\text{max}}$ is 27.77 and 27.82 ($T_{\text{max}}$ is observed at the bottom) for the red and orange groups, respectively, based on the isopycnally averaged (increment of 0.01) $\theta$–S profiles from the data of 2007, 2009, and 2010 (Fig. 11). This indicates that the intruding CDW of the red group is ~0.35°C warmer than that of the orange group. Thus, the red, pink, and orange groups can also indicate the thickness of the CDW intrusion. However, it is important to note that the $T_{\text{max}}$ can get modified as a result of mixing with surrounding water away from the continental shelf break. For example, pink stations in the middle of the continental shelf are considered to be formed by the mixing of the red and orange groups. In addition, the potential density at the bottom of most orange profiles does not reach 27.82 (Fig. 9). This means that all orange profiles should originally have formed due to the intrusion of CDW at ~27.82, but CDW already gets modified (colder and less saline) due to vertical mixing as it progresses southward (Fig. 12).

The different thicknesses of both CDW intrusions may be related to Ekman transport induced by along-slope currents over the continental slope and shelf break (Wåhlin et al., 2012). The CDW observed at the bottom near the shelf break has a potential density of ~27.81–27.82, i.e. this CDW is withdrawn from a depth of ~700–900 m (consistent with Wåhlin et al., 2012). The thickness and transport of the Ekman layer depends on the topography and depth of the shelf break, which may be the reason for the different characteristics of the red and orange groups (MacCready and Rhines, 1993; Wåhlin et al., 2012). The occurrence of warmer (colder) CDW in trough E (C) may also be related to the cyclonic circulation in the PIT, which cools the CDW entering from trough C, as proposed in Schodlok et al. (2012).

4.2. The pathway of CDW intrusion on the continental shelf

Dense CDW, intruding onto the continental shelf and cascading down to the PIT, descends deep and flows in the clockwise direction as a result of the Coriolis force. Thus, possible pathways of the CDW intrusion should follow the bathymetry as indicated by the two gray arrows in Fig. 8a. Schodlok et al. (2012) show, based on numerical modeling, that most of the PIT is dominated by a cyclonic gyre that extends to the north of Section 1 (consistent with possible pathways in Fig. 8a).

Since dense CDW cascades into the PIT, the intruding CDW should be denser than the water mass at the bottom of PIT (blue), where the potential density is ~27.81 (gray line in Fig. 9). The 27.81-contour (Fig. 12) indicates that CDW denser than 27.81 originates from the eastern shelf break, trough E (102°W–108°W), but not from trough C. Thus, the eastern trough plays the most important role for dense CDW flowing into the PIT. Since the blue group has the same $T_{\text{max}}$ as the pink group, the formation of ‘blue’ characteristics could be easily explained if the pink group would dominantly exist on the continental shelf. However, when we focus on the intrusion via trough E, the CDW is mostly in the red group and CDW properties at the density of 27.75–27.81 are largely modified. This modification is obvious as the $T_{\text{max}}$ changes from ~1.4°C to ~1.2°C (blue) on the way from trough E to PIT (lower left panel of Fig. 12). Since mainly red and orange groups exist on the continental shelf (except for a few pink stations which are considered to be formed by mixing of red and orange groups) and the 27.78-contour (Fig. 8b) shows the intrusion of CDW from both troughs, it is considered that the orange group (from trough C) mixes with the red group (from trough E) modifying and forming the blue group. Several red profiles on the western side of trough E (e.g. two stations surrounded by black squares in lower left box of Fig. 12) show the mixing between the red and orange groups (Fig. 11, red dashed lines). To analyze the contribution of the red and orange groups for forming the blue group, we assume that the averaged red, orange and blue groups are representative and the contribution of each group can be calculated from the distance of each $\theta$–S profile to the blue $\theta$–S profile. The ratio of red to orange contributions to form the blue is calculated to be roughly 1:1.5 below the 27.78-isopycnal (Fig. 11).

Because all the CTD measurements were conducted in austral summer, it cannot be excluded that the intrusion of CDW has a seasonal cycle. Thoma et al. (2008) show that the inflow via trough C is weakest in austral summer. However, the distribution of CDW-layer thickness in their model always shows a thicker CDW layer in trough C than in trough E, which is not consistent with the observations between 2007 and 2010. This may be due to old topographic data, which does not include the eastern submarine channel. Hence, the eastern trough plays the most important role for dense CDW flowing into the PIT. Since the blue group has the same $T_{\text{max}}$ as the pink group, the formation of ‘blue’ characteristics could be easily explained if the pink group would dominantly exist on the continental shelf. However, when we focus on the intrusion via trough E, the CDW is mostly in the red group and CDW properties at the density of 27.75–27.81 are largely modified. This modification is obvious as the $T_{\text{max}}$ changes from ~1.4°C to ~1.2°C (blue) on the way from trough E to PIT (lower left panel of Fig. 12). Since mainly red and orange groups exist on the continental shelf (except for a few pink stations which are considered to be formed by mixing of red and orange groups) and the 27.78-contour (Fig. 8b) shows the intrusion of CDW from both troughs, it is considered that the orange group (from trough C) mixes with the red group (from trough E) modifying and forming the blue group. Several red profiles on the western side of trough E (e.g. two stations surrounded by black squares in lower left box of Fig. 12) show the mixing between the red and orange groups (Fig. 11, red dashed lines). To analyze the contribution of the red and orange groups for forming the blue group, we assume that the averaged red, orange and blue groups are representative and the contribution of each group can be calculated from the distance of each $\theta$–S profile to the blue $\theta$–S profile. The ratio of red to orange contributions to form the blue is calculated to be roughly 1:1.5 below the 27.78-isopycnal (Fig. 11).
intrusion from trough E at its peak level, but the seasonality of the intrusion requires further investigation.

4.3. Melting of PIIS

4.3.1. Meltwater fraction at the ice shelf front

We calculate the meltwater fraction (see Appendix A) for the ice front section. From the θ–S diagrams of the green and cyan groups (Fig. 7e and f), the zero meltwater fraction contour connects the characterized WW and CDW properties (Table 2). The deviations from this contour, especially for the cyan group, indicate the inclusion of glacial melt (Fig. 7e and f).

The vertical profiles of the meltwater fraction in θ–S, O₂–S, and O₂–θ spaces from the ice shelf front section (Fig. 13) show strong meltwater signals at the depth of 100–500 m with a somewhat larger meltwater fraction (~1.5%) at the depth of 200–300 m. The meltwater fraction is also calculated from the concentration of dissolved noble gases (neon and helium), which are supersaturated in seawater when air bubbles are released from the ice by basal melting. Due to their low solubility, these supersaturated noble gas signals show the clearest signature of meltwater from meteoric ice (Schlosser, 1986; Schlosser and Lehmann, 1987).

![Color-coded θ–S diagrams with the same profiles shown in Fig. 6. The blue, red, orange, pink, green, and cyan groups are shown separately in (a)-(f), respectively. For all figures, 2010-data are plotted with dark color, and 2007-data and 2009-data are plotted with light color. For (b)-(f), averaged profiles of blue are always shown as reference. In (e) and (f), the thick black lines are meltwater fractions (contour of 0%, 0.6% (only in d), and 2%) and the black dotted contour is the theoretical upper bound of the meltwater fraction. The dashed line indicates the CDW/meltwater mixing trend.](image-url)
Hohmann et al., 2002). We define representative helium and neon concentrations in CDW, WW, and meltwater for CDW following Hohmann et al. (2002), for meltwater following Hohmann et al. (2002) and Huhn et al. (2008), and Huhn et al. (2008) for WW (Table 2). Helium and neon samples are only available at some specific depths because of the limited amount of sample devices having been available on the cruise. The vertical profiles of meltwater fractions (Fig. 13) show that the fractions in $\theta$-S, $O_2$-S, and $O_2$-$\theta$ spaces are mostly consistent with the meltwater fractions in He-S and Ne-S spaces. The causes for some small disagreements are considered to be due to nonuniform WW properties. It is not possible to fit all the meltwater fractions in $\theta$-S, $O_2$-S, and $O_2$-$\theta$ spaces to the meltwater fractions obtained from helium and neon, even though the defined water mass characteristics are adjusted (not shown). It is

Fig. 8. Locations of the color-coded CTD stations in (a) 2010 and (b) 2007, 2009, and 2010 on the eastern Amundsen Sea continental shelf (all stations on the shelf east of the subsurface ridge extending out from Bear Peninsula are shown). Pink stations are surrounded by black circles. Stations not used for the analysis are shown in gray. The enclosed region indicates the location referred to in Fig. 15. Two stations in 2009 are not used because of a thick WW layer reaching almost to the bottom. The two arrows in (a) show the possible pathway of the CDW intrusion.
important to note that the meltwater fractions derived in He–S and Ne–S spaces are as low as 0.3% at 800 m (station 153), which will be discussed below. Since the detection limits of the meltwater fraction for helium and neon are in the order of 0.1%, these signals are clearly above the detection limit.

The calculation of the meltwater fraction is invalid for the surface mixed layer, because equilibration with the atmosphere raises the dissolved oxygen concentration toward saturation, and other sources of freshwater reduce the salinity (Jenkins, 1999; Jenkins and Jacobs, 2008). Thus, the meltwater fractions derived from $\theta$–S, $O_2$–S, and $O_2$–$\theta$ spaces show larger disagreement at depths shallower than $\sim$150 m (Fig. 13). In the following analysis, the average of these three meltwater fractions is calculated, while the meltwater fractions for depth $\sim$150 m, where the standard deviation of these three meltwater fractions is larger than half of the theoretical upper bound (Appendix A) is excluded from the analysis.

4.3.2. Basal melt rate of the PIIS

We calculate the transport of meltwater into and out of the cavity beneath the PIIS by analyzing the CTD ice front section (Fig. 5). Since direct observations of ocean currents do not exist, we estimate the velocity assuming hydrostatic and geostrophic balance. In estimating the full-depth ocean current, however, it is difficult to define the barotropic reference velocity (Jenkins and Jacobs, 2008). Thus, we first provide a simple estimate of the melt rate using geostrophic balance and conservation of salt between inflow and outflow.

Fig. 9. Vertical profiles of potential temperature (a–c) and potential density (d–f) of the red, orange, and blue groups from 2010–(dark color), 2007–, and 2009-data (light color). In a, b, d, and e the blue lines are the depth-averaged profile of the blue group. In (d–f), the gray lines mark the 27.81-isopycnal and the insets magnify the bottom part of the profiles.

Fig. 10. Vertical sections of potential temperature of (a) Section 1 (but from stations 195–224) and (b) Section Appendix A from 2009 shown in Fig. 8. Green lines show $T_{\text{max}}$, and shading is drawn for only the part deeper than $T_{\text{max}}$, which corresponds to the intruding CDW layer. Shading colors are defined according to the red, pink, and orange groups. The vertical black lines mark the location of CTD profiles. For Section 1, color-coded station numbers are shown. For Section Appendix A, color-coded circles are shown. Bottom topography is drawn based on the bottom depth of each profile. For (a), south of station 224 is shaded blue to indicate that there stations are defined as the blue group. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Our first estimate of the melt rate is based on the assumption that the level of no motion is at the depth of 500 m following Jacobs et al. (2011) and CDW flows into the cavity below that level. The mean geostrophic velocity below 500 m is 1.89 cm s$^{-1}$, resulting in an inflow of 228 mSv (1 mSv = $10^3$ m$^3$ s$^{-1}$). In addition, salt should be conserved and $(V_{in} + V_{melt})S_{out} = V_{in}S_{in}$ should be satisfied, where $V_{in}$ and $V_{melt}$ are the volume transport of inflow and meltwater, respectively and $S_{in}$ and $S_{out}$ are the salinity of the inflow and outflow, respectively. The average salinity for the inflow and outflow are calculated to be 34.59 and 34.28 (average outflow salinity is calculated between 200 and 500 m depth). Thus, $V_{melt}$ is calculated to be 2.05 mSv, which is equivalent to an ice loss of 72.3 km$^3$ yr$^{-1}$, or a mean melt rate of 30 m yr$^{-1}$, assuming a cavity size of 70 $\times$ 35 km$^2$. (Note that the cavity size might have increased from 2009 to 2010 to $\sim$74 $\times$ 35 km$^2$, causing our result to be overestimated by $\sim$5%.)

Although a somewhat realistic melt rate was obtained with this simple approach, we employ the more sophisticated method by Wunsch (1978), Jenkins and Jacobs (2008), and Jacobs et al. (2011) (see Appendix B). Following Jacobs et al. (2011), we assume that the level of no motion is set to the depth where the meltwater fraction is 0.5% (yellow dashed line in Fig. 14) and the surface layer, where the meltwater fractions show large disagreements, is excluded from the calculation (region surrounded by black dashed line in Fig. 14). In this calculation, half of the theoretical upper bound (Appendix A) is used to investigate the accuracy of the meltwater fraction, while a quarter of theoretical upper bound was used in Jacobs et al. (2011) to obtain a more realistic circulation.

The adjusted geostrophic velocity profile along the PIIS front (Fig. 14) shows inflow mostly in the north and outflow mostly in the south, indicating a clockwise sub-ice shelf circulation. Our estimated inflow, outflow, and difference are 602 mSv, 604 mSv and 1.98 mSv, respectively, which is equivalent to a basal melting of 70 km$^3$ yr$^{-1}$ of ice. Jacobs et al. (2011) calculated for inflow, outflow, and difference 400 mSv, 402 mSv, and 2.26 mSv (80 km$^3$ yr$^{-1}$ of ice) for 2009 and 219 mSv, 220 mSv, and 1.49 mSv (53 km$^3$ yr$^{-1}$ of ice) for 1994. If all the meltwater transported out of the cavity originated from the central region of the ice shelf, whose area is 70 $\times$ 35 km$^2$, the basal melt rate would be 29 m yr$^{-1}$ in 2010, 33 m yr$^{-1}$ in 2009, and 22 m yr$^{-1}$ in 1994 (Jacobs et al., 2011).
The melt rate in 2010 is similar to that of 2009 (~30 m yr\(^{-1}\)), but both are significantly higher than the rate of 1994. However, transport in and out of the PIIS cavity increased from 1994 to 2010 with a 50% rise from 2009 to 2010. From 1994 to 2009, large changes in the shape of the cavity might have occurred causing the significant increase in the strength of sub-ice shelf circulation and melt rate (Jacobs et al., 2011). However, since ice shelf geometry changes with a time scale of decades (Jenkins et al., 2010), it is unlikely that an altered cavity shape increased the sub-ice shelf circulation from 2009 to 2010. The enhanced sub-ice shelf circulation in 2010 may be the result of capturing the recirculation of CDW, caused by short term variability of the clockwise eddy located in front of PIIS (Mankoff et al., 2012; Jacobs et al., 2012). Because average inflow and outflow potential temperature are 0.5 °C and 0.0 °C, respectively, in 2009 (calculated from Jacobs et al., 2012) and 0.5 °C and 0.1 °C in 2010, more of the CDW flowed in and out without melting the ice shelf. However, it might be erroneous to assume that our CTD stations effectively closed off the cavity. Because the southernmost station is located too far south, our measurements could not capture the primary meltwater outflow, which should be to the south of Section 4. To test the sensitivity of our result, we substitute temperature, salinity, and oxygen of station 157 by those of station 16 from 2009 (Jacobs et al., 2011, Figs. 1 and 2) but keeping the geostrophically calculated velocity. This lead to a similar melt rate with an inflow and outflow closer to the result of Jacobs et al. (2012), also indicating that the strong inflow and outflow in 2010 may be an artifact of sparse CTD observations.

4.4. The pathway of meltwater

The location of the green group shows that the meltwater flows along the bathymetry towards the west (Fig. 8). This flow may be considered as a current supplied by the constant outflow of buoyant meltwater from the PIIS and other cavities. On the isopycnal of 27.5–27.7, meltwater fractions of about 0.4% are observed in the blue group (Fig. 7a, e, and f). Since the meltwater fractions are fairly small, we investigate whether these values are meaningful. In general, when two water masses mix, the \( \theta-S \) diagram shows a straight line between these two water masses. The blue \( \theta-S \) profiles, however, curve toward a higher meltwater fraction at the potential density of 27.5–27.7, which cannot be formed as a mixing of CDW and WW only. In addition, the helium and neon concentrations at the PIIS front at the depth of 800 m also show a meltwater signal (~0.3%, clearly above the detection limit), indicating the addition of meltwater to the CDW layer. Thus, the meltwater signal observed at the isopycnal of 27.7 should be meaningful, possibly caused by the recirculation of meltwater in PIT.

4.5. Comparison with previous observations

At the locations of the color-coded CTD stations (Fig. 8b), the characteristics of the intruding CDW remain similar during the years 2007, 2009, and 2010. We thus compare the blue profiles (PIT) of the three years with five CTD profiles from 2000 in PIT (black dots within black box in Fig. 15). Although Jacobs et al. (2012) stated that the temporal variability of CDW intrusion is large, the temporal variability of the \( \theta-S \) profiles in PIT (blue) is extremely small compared to the red and orange group on the outer shelf. Thus, the \( \theta-S \) profile in PIT can be used for detecting changes in the CDW intrusion.

Compared to the 2000-data, the \( \theta-S \) profiles of 2007–2010 (Fig. 15) are warmer and saltier. The main difference, however, is
the thicker CDW layer, while the thickness of the WW layer (potential temperature ~1.8°C) remained similar. The potential temperature of the CDW increased by ~0.15°C and the gradient of the WW/CDW thermocline became steeper in 2009–2010, but the density of the CDW remained almost the same. Thus, a thicker CDW layer with higher temperature was able to penetrate onto the continental shelf in 2007–2010 compared to 2000.

5. Summary

After the discovery in 1994 that PIIS is rapidly melting due to the presence of CDW on the Amundsen Sea continental shelf (Jacobs et al., 1996), several observations were conducted. To understand the melting of PIIS, we studied the CDW flow onto the Amundsen Sea continental shelf, basal melting of PIIS, and spreading of the resulting meltwater. In this study, we mainly analyzed the data from observations in 2010 done by the German expedition ANTXXVI/3.

CDW intrudes onto the eastern Amundsen Sea continental shelf through the eastern and central submarine glacial troughs (Fig. 1). The potential temperature of the CDW intrusion is controlled by the thickness of the CDW layer. The thick intrusion via trough E contains light and warm CDW (warmer than 1.23°C) as well as dense and cold CDW (colder than 1.23°C), while the thin intrusion via trough C only contains dense and cold CDW (Figs. 8 and 10).

To fill PIT with CDW, the potential density of the intruding CDW has to be denser than the water mass at the bottom of PIT (blue). The course of the 27.81-isopycnal of bottom potential density shows that the CDW intrusion denser than 27.81 reaches PIT only from trough E (red). However, this CDW is largely modified and the Tmax changes from ~1.4°C to ~1.2°C on the way from trough E to PIT (lower left panel of Fig. 12). This modification occurs as a result of mixing with the orange group (CDW typically seen in the vicinity of trough C). The mixing ratio of the red and orange groups is calculated to be about 1:1.5 below the 27.78-isopycnal (Fig. 11).

The vertical profiles of meltwater fractions at the PIIS front show the outflow of meltwater at the depth of 100–500 m (Fig. 13), consistent with the meltwater fractions obtained in He–S and Ne–S spaces. The estimated melt rate in 2010 ranges from 29 to 30 m yr⁻¹ depending on the method applied. This is significantly larger than the melt rate in 1994 (~22 m yr⁻¹) and similar to the melt rate in 2009 (Jacobs et al., 2011). However, the calculated inflow and outflow shows a rather suspicious 50% increase from 2009 to 2010. This is either caused by capturing short term variability in Pine Island Bay or the assumption that our CTD section effectively closes off the cavity.

The meltwater spreads along the bathymetry towards the west (green group in Fig. 8). From the meltwater fraction on the isopycnals 27.5–27.7 and the helium and neon concentrations at the ice shelf front at the depth of 800 m, the meltwater signal of the intruding CDW is calculated to be 0.3–0.4%, indicating that some of the meltwater mixes with CDW.

The characteristics of the intruding CDW remain similar from 2007 to 2010 (Fig. 8b). The main difference is that the CDW layer became thicker and warmer, containing warm and light CDW as well as cold and dense CDW (Fig. 15). Such potential temperature increase in the CDW of PIT is not considered to be the cause for increased melting of PIIS, which seems to be the consequence of a change in cavity shape as reported by Jenkins et al. (2010) and Jacobs et al. (2011).

Despite numerous observations and modeling studies on PIIS, we still do not fully understand how PIIS melting is controlled and how it will respond to climate change. Understanding PIIS melting is important, because CDW is also melting some other glaciers in West Antarctica in a similar way. Numerical modeling also shows that intrusion of CDW can occur on the southern Weddell Sea continental shelf, resulting in rapid basal melting in the near future (Hellmer et al., 2012). Thus, a detailed understanding of CDW intrusion onto the continental shelf and ice sheet-ocean interaction is crucial for predicting the consequences of climate change in the Southern Ocean.

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Appendix A. Meltwater fraction calculations

The meltwater fraction is calculated simply as the mixture of three water masses (Jenkins, 1999; Jenkins and Jacobs, 2008; Jacobs et al., 2011) as shown in the following equations:

\[ \chi_{\text{mix}} = A \chi_{CDW} + B \chi_{MW} + C \chi_{WW}. \]  

\[ \chi_{\text{mix}} = A \chi_{EDW} + B \chi_{CDW} + C \chi_{WW}. \]  

\[ A + B + C = 1, \]  

where \( A, B, \) and \( C \) are fractions of CDW, meltwater, and WW, respectively, and \( \chi \) is a tracer that is conserved during mixing. Two tracers \( \chi_{1} \) and \( \chi_{2} \) are chosen from the available parameter sets of potential temperature, salinity, dissolved oxygen, and concentrations of helium and neon.

The subscripts of \( \chi \) indicate defined water mass characteristics for CDW, WW, and meltwater following Jenkins (1999), Jenkins and Jacobs (2008), and Jacobs et al. (2011), summarized in Table 2.

By solving for \( B \), the meltwater fraction can be obtained as:

\[ B = \frac{(\chi_{\text{mix}} - \chi_{CDW})(\chi_{\text{mix}} - \chi_{CDW})(\chi_{\text{mix}} - \chi_{CDW})}{(\chi_{\text{melt}} - \chi_{CDW})(\chi_{\text{melt}} - \chi_{CDW})(\chi_{\text{melt}} - \chi_{CDW})}. \]  

The theoretical upper bound is the maximum meltwater fraction \( (C=0 \text{ in Eq. (A.1))} \) which occurs when the potential temperature of CDW \( (\chi_{CDW}) \) is reduced to the freezing point and no further melting can occur (the intersection point of freezing point and CDW/meltwater mixing line in Fig. 7, Jenkins and Jacobs, 2008).

Appendix B. Transport calculations

We use the same techniques used near George VI Ice Shelf (Jenkins and Jacobs, 2008) to estimate the basal melt rate, which is based on the classical inverse method (Wunsch, 1978). We assume that our CTD sections effectively close off the cavity, and the total mass transport must satisfy:

\[ \sum_{j=1}^{4} (M_{\text{out}} - M_{\text{in}}) = M_{\text{melt}}. \]  

where \( M_{\text{in}}, M_{\text{out}} \) and \( M_{\text{melt}} \) are the mass transport of inflow, outflow and meltwater, respectively, and \( n \) is the number of vertical velocity profiles between \( n+1 \) CTD stations. Assuming that cavity properties are in steady state, a similar equation must be satisfied for the transport of each tracer \( \chi \) across the section shown as

\[ \sum_{j=1}^{4} (M_{\text{in}}^{\chi} - M_{\text{out}}^{\chi}) = M_{\text{melt}}^{\chi} \text{MWV}. \]  

Combining the conservations of total mass transport and tracer transport, Eqs. (B.1) and (B.2) can be reduced to

\[ \sum_{j=1}^{4} \int_{p_{\text{in}}}^{p_{\text{out}}} (v_{j} + \nabla(p))\nabla(p) \nabla(p) \Delta(p) \, dp = 0, \]  

where overbars indicate averages over the width \( \Delta x \) between stations, \( g \) is the acceleration due to gravity, \( p \) is the pressure, \( v_{j}^{\chi} \) is the barotropic reference velocity and \( \nabla(p) \) is the velocity obtained from the geostrophic balance. In our calculation, \( n \) equals to four with four unknown barotropic reference velocities, while we have only three equations for potential temperature, salinity, and dissolved oxygen. Since, this problem has many possible solutions, we choose the one that minimizes the sum of the squares of the barotropic reference velocities.

References


