Abstract. We present the first year-long current meter records ever obtained near the floating Filchner-Ronne Ice Shelf in the Weddell Sea. The currents are steered along the ice front, but in the lower layer where the bottom topography is descending toward the west the current has a component toward the ice front of about 3 cm s\(^{-1}\). During winter the temperature stayed near the surface freezing point, while the salinity increased, indicating that ice was formed and brine released. The seasonal variation in salinity was 0.15 ± 0.05 psu, corresponding to the formation of 1–2 m of ice on a shelf depth of 400 m. The transport of High-Salinity Shelf Water (HSSW) into the ice shelf cavity was found to be of the order 0.5 \(\times\) \(10^6\) m\(^3\) s\(^{-1}\). The production of this water due to oscillating tides and off shelf winds was found to be of the same order of magnitude. In contact with glacial ice at great depths, and because of the depression of the freezing point, the HSSW is transformed to Ice Shelf Water (ISW) by cooling and melting processes. The melting rate was estimated to 1 \(\times\) \(10^{11}\) ton yr\(^{-1}\). This corresponds to the melting of 0.2 m ice per year if the melting is evenly distributed over the Filchner-Ronne Ice Shelf. If the melting is concentrated along a path from the Berkner Shelf around the Berkner Island to the Filchner Depression, then melting rates up to 7 m yr\(^{-2}\) must be expected. A comparison of HSSW characteristics in the Ronne Depression, our winter observations on the Berkner Shelf, and the ISW flowing out of the Filchner Depression indicates that very little water passes through the cavity from the Ronne to the Filchner Depression. It appears that most of the ISW originating from processes on the Berkner Shelf escapes the cavity in the Filchner Depression. This leaves the Berkner Shelf as the important source of ISW and subsequently of the Weddell Sea Bottom Water formed from ISW.

1. Introduction

The Antarctic Ice Sheet is believed to be a key factor controlling the climate of the planet. The impact of the ice sheet on the climate has two main components: (1) the cooling of the atmosphere through its impact on the radiation balance, and (2) the cooling of the world ocean through the formation of extremely cold and dense water, ventilating the world ocean abyss.

The snow deposited over the Antarctic Ice Sheet is returned to the ocean at the base of the melting ice shelves and icebergs. Thus it is of vital interest to quantify and understand the mass balance of the ice sheet. For example, increased melting will enhance the cooling of the ocean, and if it results in a negative mass balance, the global sea level will increase.

The most important drainage of the grounded Antarctic Ice Sheet is via the Filchner-Ronne Ice Shelf in the southern Weddell Sea [Giovineotto and Bentley, 1985]. The melting under the floating ice shelves and the calving of icebergs are believed to be of the same order of magnitude, [Jacobs et al., 1992]. The melting processes have traditionally been studied by summer observations of water masses at the northern rim of the floating ice shelves. During the 1990s a few direct measurements have been made underneath the floating ice shelves by hot water drilling through the ice; for a review, see Nicholls and Makinson [1998]. In the present paper further insight into the melting process is obtained by analysis of the first year-long current meter records obtained near the Ronne Ice Front.

Early in the century [Brennecke, 1921; Deacon, 1937], the Weddell Sea was pointed out as the main source region for Antarctic Bottom Water (AABW). High-Salinity Shelf Water (HSSW) (earlier termed Western Shelf Water), which is a dense water mass formed on the shelf through the formation of sea ice and brine release, is known to be one of the “parent” water masses of AABW [Foster and Carmack, 1976; Foldvik et al., 1985b].

The Norwegian Antarctic Research Expedition 1976–1977 located a source of the extremely cold Ice Shelf Water (ISW) overflowing the sill (at ~600 m depth) of the Filchner Depression [Foldvik et al., 1985a]. Subsequent current meter observations at the sill showed that this source represented a substantial contribution to the overflow with a yearly average volume
flux of \((0.8 \pm 0.2) \times 10^6 \text{ m}^3 \text{ s}^{-1}\) at an average temperature of \(-2.0^\circ\text{C}\) [Foldvik et al., 1985b; Nygaard, 1994]. Observations on the continental slope showed that the overflowing plumes kept their characteristic features past the 2000 m level. The ambient water temperature is here below \(-0.3^\circ\text{C}\), and a simple mixing model indicated a resulting production of Weddell Sea Bottom Water (WSBW) (defined at \(-0.8^\circ\text{C}\)) of about \(4-5 \times 10^6 \text{ m}^3 \text{ s}^{-1}\) [Foldvik and Gammelsrød, 1988]. The Filchner overflow is not the only source of WSBW. On the basis of tracer analysis, Anderson et al. [1991] postulated a separate source farther to the west, later confirmed by conductivity-temperature-depth (CTD) measurements by Gordon et al. [1993] and by current meter observations by Fahrbach et al. [1995].

CTD observations obtained during several summer expeditions indicate a persistent southward flow of Modified Warm Deep Water (MWDW) into the ice shelf cavity on the western slope of the Berkner Shelf [Gammelsrød and Slotsvik, 1981; Rohardt, 1984; Foldvik et al., 1985c; Gammelsrød et al., 1994; Nøst and Østerhus, 1998]. The southbound MWDW flow also clearly emerges in model experiments using realistic topography [Grosfeld and Gerdes, 1998]. A currentmeter rig with two currentmeters was positioned on the slope to monitor this flow into the cavity, and the results of these measurements provide the basis of this paper.

Subject to the limitations of having only one currentmeter rig, the measurements allow us to make the order of magnitude estimates of the formation of dense HSSW near the ice front and the transport of this water into the ice shelf cavity. The tidal regime and the seasonal variations are discussed. Melting rates and the formation of ISW are estimated. On the basis of water mass observations along the ice front and model experiments the sources of ISW as well as circulation patterns below the ice are discussed.

2. Data, Instruments, and Methods

2.1. Navigation

The main objective of the Norwegian Antarctic Research Expedition 1992–1993 was to study water masses along the Filchner-Ronne Ice Shelf. This is extremely difficult territory to navigate. The vessel approached the area from the northeast, going westward along the ice front of the Filchner Ice Shelf. The approximate cruise track may be deduced from the CTD stations shown in Figure 1. Satellite pictures received onboard (see Figure 2) showed a narrow \(\sim 1\) and \(\sim 50 \text{ nm}\) long lead between the icebergs and the fast ice. The ship used this narrow lead to reach the Ronne Ice Shelf. This lead closed a few weeks later, and the expedition had to make a detour around the grounded icebergs to get back to the Filchner Ice Shelf.
2.2. Moorings

The current meter moorings were equipped with Aanderaa RCM7 instruments provided with temperature and conductivity sensors. The accuracy of the velocity measurements is $\pm 1 \text{ cm s}^{-1}$ for the speed and $\pm 5^\circ$ for the direction. The temperature and conductivity sensors were calibrated prior to the cruise and checked in situ against CTD measurements. The accuracy of these measurements is $\pm 0.05^\circ$C for the temperature and $\pm 0.02 \text{ mmho cm}^{-1}$ for the conductivity. One of the conductivity sensors at R2 (at 400 m depth) was obviously contaminated after about 13 months, giving unrealistically low salinities. The last 2 months of this time series are therefore not used.

Mooring R2 (Figure 1) was deployed in 419 m water depth on February 2, 1993, carrying two Aanderaa current meters at 245 and 400 m depth. The distance from the ice front was 4.3 km at deployment and 0.5 km at retrieval on February 13, 1995, indicating a minimum translation speed of the ice front of $1.9 \text{ km yr}^{-1}$. The exact positions, duration of measurements, and average values are given in Tables 1a and 1b.

A short-term mooring RT (see Figure 1) was deployed on February 4, 1993, and recovered after 10 days. It carried one

<table>
<thead>
<tr>
<th>Rig</th>
<th>Position</th>
<th>Instrument Type</th>
<th>Instrument</th>
<th>Time Out</th>
<th>Period, days</th>
<th>Observation Depth</th>
<th>$T_c$ °C</th>
<th>$S$, psu</th>
</tr>
</thead>
<tbody>
<tr>
<td>R2</td>
<td>76°28.85'S, 53°00.03'W</td>
<td>RCM-7</td>
<td>10907</td>
<td>February 5, 1993</td>
<td>435</td>
<td>245</td>
<td>-1.69</td>
<td>34.40</td>
</tr>
<tr>
<td>R2</td>
<td>bottom 419 m</td>
<td>RCM-7</td>
<td>10909</td>
<td>February 5, 1993</td>
<td>435</td>
<td>400</td>
<td>-1.87</td>
<td>34.60</td>
</tr>
<tr>
<td>RT</td>
<td>76°57.73'S, 49°04.61'W</td>
<td>RCM-7</td>
<td>9707</td>
<td>February 4, 1993</td>
<td>10</td>
<td>235</td>
<td>-1.90</td>
<td>34.60</td>
</tr>
<tr>
<td>RT</td>
<td>bottom 270 m</td>
<td>WLR-8</td>
<td>1473</td>
<td>February 4, 1993</td>
<td>10</td>
<td>270</td>
<td>-1.94</td>
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</table>
Table 1b. Results of the Current Measurements From the Moorings, Mean Speed, Average Speed in Direction Parallel (114°) and Normal (24°) to the Ice Front, Average Current (Speed and Direction), and the Current Stability Factor

<table>
<thead>
<tr>
<th>Rig</th>
<th>Depth (m)</th>
<th>Mean Speed (cm s⁻¹)</th>
<th>Mean u (114°)</th>
<th>Mean v (24°)</th>
<th>Average Current</th>
<th>Stability Factor</th>
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</thead>
<tbody>
<tr>
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<td>0.05</td>
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</tr>
<tr>
<td>R2</td>
<td>400</td>
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<td>9.07</td>
<td>275</td>
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<tr>
<td>RT</td>
<td>235</td>
<td>14.2</td>
<td>-4.85</td>
<td>-0.11</td>
<td>5.35</td>
<td>245</td>
</tr>
</tbody>
</table>

current meter at 235 m depth and one Aanderaa WLR-8 pressure sensor at the bottom (270 m). The accuracy of the pressure record is ±6 × 10⁻³ m. Its position and additional information are given in Tables 1a and 1b.

3. Results

3.1. Currents

The R2 progressive vector diagrams at 245 and 400 m depth are shown in Figure 3 where also the general orientation of the ice front is indicated. The average speed was about 8 and 9 cm s⁻¹, respectively (Tables 1a and 1b). The upper current was obviously steered by the ice front, while at the lower meter the current showed a component toward the ice front.

The long-term variations (15 days low-passed Butterworth filter) of the current components parallel with (114°) and perpendicular to (24°) the ice front are shown in Figure 4. The component parallel to the ice front (Figure 4a) was rather barotropic. Perpendicular to the ice front (Figure 4b), the currents at the two levels show larger differences since the flow at the upper level is more restricted by the ice front than the flow at the lower level. Thus at the upper meter the velocities are usually below 2 cm s⁻¹ in the direction of the ice front, whereas the lower meter almost always shows a current component toward the ice front with an average speed of about 3 cm s⁻¹ (Table 1b). There is no obvious seasonal signal in the current measurements.

3.2. Power Spectra

The power spectra for the current components at rigs R2 and RT are shown in Figures 5 and 6. The diurnal and semidiurnal tidal components clearly emerge in the spectra, with most energy on the component normal to the ice front. The semidiurnal components are the most energetic tidal components. We also note that the subharmonics of the diurnal and semidiurnal periods show up with ~8, 6, and 4 hour periods. These periods are probably forced by the additional boundary conditions imposed by the nearby ice front. There is also some mesoscale energy at 7–20 days for the flow along the ice front at R2.

3.3. Temperature and Salinity Records

The temperature and salinity records (101 hours low-pass filtered) are presented in Figure 7. For both records a seasonal signal is clearly seen. During late winter and spring (September to November) the temperatures at both levels were close to the surface freezing point (Figure 7a), and occasionally, the salinities exceeded 34.7 psu (Figure 7b). In this period we note a low stability (Figure 7c), probably indicating active convection due to cooling and freezing at the surface. The annual salinity amplitude was within 0.15 ± 0.05 psu at both current meters corresponding to the formation of 1–2 m of ice on a shelf depth of 400 m.

Note that the temperatures at both levels were higher in March 1994 than in March 1993, indicating substantial year to year variations. In fact, at 400 m the temperature remained near the surface freezing point until April 1993, whereas in 1994 the warming started already in January.

The relative warm and fresh (T > -1.7°C, S < 34.6) water mass observed in summer is usually referred to as MWDW [Foster and Carmack, 1976]. MWDW has been observed near the ice front on the western slope of the Berkner Shelf during several (summer) expeditions [Gammelsrød and Slotsvik, 1981; Rohardt, 1984; Foldvik et al., 1985c; Gammelsrød et al., 1994; Nøst and Østerhus, 1998].

HSSW has not been observed close to the ice front on the Berkner Shelf during the above mentioned summer expeditions. Usually, HSSW is only observed in the Ronne Depression during summer.

Nicholls [1997] observed the fingerprints of the seasonal signal produced at the ice front as far as 500 km away under the ice shelf. However, the seasonal signal was not detectable in the ISW flowing out of the Filchner Depression around 1500 km away from the source [Foldvik et al., 1985d].
4. Discussion

4.1. The Tides

4.1.1. Tidal currents and water level. The current and pressure records have been analyzed for tidal constituents using the algorithms developed by Foreman [1977, 1978], and the most important components are given in Tables 2a–2d. The tidal ellipses for the main semidiurnal (M2) and diurnal (K1) periods are shown in Figure 8. Note that the major axis is oriented almost perpendicular to the ice front. Also note that for M2 at rig R2 the direction of rotation changes from counterclockwise at 245 m depth to clockwise at 400 m; see the discussion below.

Makinson and Nicholls [1999] and Robertson et al. [1998] have modeled the tides in the area using depth-integrated, numerical models. Their results compare well with our observations of the dominant semidiurnal tidal constituents both in amplitude and wave direction. The diurnal constituents are underestimated in the models. The discrepancies are probably due to the poor representation of bathymetry in the models and to the baroclinic effects, in particular because of the floating ice shelf. The tidal constituents for RT given in Tables 2c and 2d should be interpreted with some caution since the time series is short (10 days).

The 10 day long bottom pressure record at RT and the current component in the direction 24° (normal to the ice front) are shown in Figure 9. We note that the tidal sea level amplitude was about 1 m. The full moon occurred on February 7. The tidal pressure signal was mixed during spring and neap, while in the transition from spring to neap it was more purely semidiurnal. An inspection of the unfiltered current measurements at R2 (not shown) also reveals this change in periodicity between transitions from spring to neap and from neap to spring.

From Figure 9 we observe that maximum tidal current out from the cavity coincides with low water level. Assuming that the M2 tide behaves like a progressive wave, its amplitude and speed will be in phase. The relationship between wave amplitude $A$ and speed $U$ is then given by the general relationship for long progressive waves:

$$U = A (g/H)^{1/2}$$

(1)

From Tables 2c and 2d we find the observed values for M2:

$$U = 0.17 \text{ m s}^{-1} \quad A = 0.68 \text{ m}.$$

The computed value of $U$ using (1) with water depth $H = 270$ m is 0.13 m s$^{-1}$. If, instead, we use the water depth below the floating ice shelf (150 m), we derive the observed speed.

4.1.2. Critical latitude effects. The M2 current changes from clockwise rotation at 400 m depth to counterclockwise at 245 m (see Figure 8). This may be explained by the strange behavior of tides near the so-called critical latitude. This is the latitude where the semidiurnal frequency equals the inertial frequency, which is at S74°28’18”, about 220 km north of the mooring site. The tidal ellipse for a tidal constituent of frequency $\omega$ may be looked upon as the sum of the two circularly rotating components $-A_e e^{i \omega t}$ and $-A_e e^{-i \omega t}$. The radii of the circular components are related to the corresponding tidal ellipse as

$$A_e = (a + b)/2 \quad A_c = (a - b)/2,$$

where $a$ and $b$ are the major and minor axes of the ellipse. The direction of rotation of the tidal ellipse is then clockwise if $A_c > A_e$ and counterclockwise if $A_c < A_e$. For barotropic tides where $A_c$ and $A_e$ are nearly constant within the water column the direction of rotation of the tidal ellipse will not change with depth. However, near the critical latitude for the tidal constituent in question the amplitude of the anticyclonic rotating component ($A_e$ in the Southern Hemisphere) is not constant but increases from bottom upward [Foldvik et al., 1990]. The cyclonic component may still be depth-independent. The result may be a counterclockwise rotating ellipse in the upper water column and a clockwise rotating ellipse in the lower water column (see Figure 10).

4.2. Water Mass Transformation on the Berkner Shelf

4.2.1. Volume transports. The production of ISW under the floating ice shelves depends on the volume flux $V_{in}$ of HSSW actually flowing into the cavity along the western slope of Berkner Island (see Figure 11). Keeping in mind the limitations of having only one currentmeter rig, we may obtain a
rough estimate of this inflow of HSSW from the current meter data and from CTD sections. The volume flux $V_{\text{in}}$ is given by

$$V_{\text{in}} = l_v h_v \bar{v},$$

where $l_v$ and $h_v$ denote some representative length and depth of the HSSW in the CTD section along the ice front and where $\bar{v}$ denotes the observed mean speed normal to this section. The production of HSSW will be most efficient over the shallow Berkner Shelf. The current is steered toward the cavity by the topography sloping down toward the west (see model experiments by Grosfeld et al. [1997]). From Figure 11 we note that the bottom topography slopes down toward the west between station 42 at the Berkner Shelf and passing station 14, giving an estimate of $l_v \approx 200$ km. It is often found, however, that currents are concentrated above maximum bottom slopes, in this case, between station 41 and the currentmeter site R2, indicating $l_v \approx 100$ km. Thus we estimate $l_v \approx 150 \pm 50$ km.

The thickness of the HSSW will vary during the year but probably exceeds 200 m in the winter season; see the temperature and salinity records (Figure 7). However, HSSW is absent from both currentmeters typically 4 months of the year. Taking into account these variations seen in Figures 7a and 7b, we choose an effective $h_v \approx 100$ m.

From Table 1b, we obtain $\bar{v} \approx 0.03$ m s$^{-1}$. Using (2), we then estimate the volume flux, $V_{\text{in}} \approx (0.3 - 0.6) \times 10^6$ m$^3$ s$^{-1}$. This estimate we consider conservative since there is reason to believe that the transverse velocities are higher farther to the east between the moorings where the slope is steeper.

4.2.2. Production of HSSW on the Berkner Shelf. The observed annual salinity amplitude in Figure 7c corresponds to the formation of between 1 and 2 m of sea ice on a shelf depth of 400 m. The question is whether this salinity amplitude can be explained by local open ocean freezing or if freezing processes near the ice front need to be invoked. On the basis of a dynamic-thermodynamic model, Lemke et al. [1990] obtained the ice thickness $\approx 1$ m at the mooring site. Tidal divergence will enhance the freezing, and a rough calculation for the semidiurnal tide increases the thickness by about 10%. Still, the resulting thickness of locally frozen ice is not sufficient to account for the observed salinity amplitude. Lemke et al.’s [1990] model also shows that the ice production has a maximum in the deeper southeastern Weddell Sea. The local enhancement of salinity depends on the water depth, but conceivably, the westward advection of this water may account for the observed salinity anomaly. Also, the advection of HSSW from the Ronne Depression should not be ruled out; see model experiments by Grosfeld and Gerdes [1998]. Below we will explore the local salinity production due to offshore winds [Gill, 1973] and due to oscillating tides.

In Appendix A we derive an expression for the average ice production rate (IP) in an area influenced by a tidal constituent:
IP = 0.3CAν₀ω₀⁻¹/₂,

(3)

where λ denotes a length interval of the ice front where ice production takes place, ω is the tidal frequency, ν₀ is the tidal surface velocity amplitude normal to the ice front, and C is a constant depending on the air temperature during freezing (see (A8)). This ice production increases the salinity of the volume V the amount ΔS, thus

\[ V = \rho_ρ \cdot \text{IP} \cdot (S - S_f)/\Delta S, \]

(4)

where \( \rho, \rho_s, \) and \( S_f \) denotes the density of seawater and the density and salinity of sea ice. In Figure 10 we estimate the surface amplitude of the \( M_2 \) velocity component normal to the shelf by extrapolating the \( A_{s1} \) and \( A_s \) circular components of \( M_2 \) to the surface. The sum of these then gives \( v_{M2} \sim 0.27 \) m s⁻¹. In (3) we consider the ice production over the length \( \lambda = 300 \) km of the ice front. The yearly average temperature in the region is about \(-19^\circ C\) [Hanssen-Bauer, 1992], which gives \( C = \)
Table 2a. Major Tidal Constituents for the Current Meter 10907 From 245 m Depth at R2*

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Major Axis, cm s⁻¹</th>
<th>Minor Axis, cm s⁻¹</th>
<th>Angle of Inclination, deg</th>
<th>Greenwich Phase</th>
<th>Period, hours</th>
</tr>
</thead>
<tbody>
<tr>
<td>O1</td>
<td>8.3</td>
<td>−0.3</td>
<td>72.5</td>
<td>270.7</td>
<td>25.82</td>
</tr>
<tr>
<td>P1</td>
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<td>−0.4</td>
<td>60.7</td>
<td>92.3</td>
<td>24.06</td>
</tr>
<tr>
<td>K1</td>
<td>9.2</td>
<td>−1.3</td>
<td>74.9</td>
<td>113.3</td>
<td>23.93</td>
</tr>
<tr>
<td>N2</td>
<td>3.5</td>
<td>1.0</td>
<td>67.0</td>
<td>103.5</td>
<td>12.66</td>
</tr>
<tr>
<td>M2</td>
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<td>3.6</td>
<td>61.2</td>
<td>125.3</td>
<td>12.42</td>
</tr>
<tr>
<td>S2</td>
<td>11.3</td>
<td>1.4</td>
<td>59.5</td>
<td>152.5</td>
<td>12.00</td>
</tr>
<tr>
<td>K2</td>
<td>3.6</td>
<td>0.1</td>
<td>50.8</td>
<td>157.1</td>
<td>11.96</td>
</tr>
</tbody>
</table>

*Positive minor axis denotes counterclockwise rotation. The angle of inclination denotes the orientation of the major axis and is measured in degrees counterclockwise from east. Greenwich phase is referred to the northern major semiaxis.

Table 2b. Same As Table 2a but for the Current Meter 10909 From 400 m Depth at R2

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Major Axis, cm s⁻¹</th>
<th>Minor Axis, cm s⁻¹</th>
<th>Angle of Inclination, deg</th>
<th>Greenwich Phase</th>
<th>Period, hours</th>
</tr>
</thead>
<tbody>
<tr>
<td>O1</td>
<td>8.2</td>
<td>−1.1</td>
<td>72.9</td>
<td>97.0</td>
<td>25.82</td>
</tr>
<tr>
<td>P1</td>
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<tr>
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<td>23.93</td>
</tr>
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</tr>
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<td>53.7</td>
<td>132.6</td>
<td>12.42</td>
</tr>
<tr>
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<td>58.8</td>
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Table 2c. Same As Table 2a but for the Current Meter 9707 From 235 m Depth at RT

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Major Axis, cm s⁻¹</th>
<th>Minor Axis, cm s⁻¹</th>
<th>Angle of Inclination, deg</th>
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<tbody>
<tr>
<td>K1</td>
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<td>0.4</td>
<td>30.4</td>
<td>102.3</td>
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</table>

4 × 10⁻⁴ m s⁻¹/². In (4) we use ΔS = 0.15 ± 0.05 psu, ρ_/ρ = 0.9, and $S - S_i = 30$ psu and obtain $V \sim (0.15 \pm 0.05) \times 10^6$ m³ s⁻¹. Note that we have only considered the contribution to the ice production due to one tidal constituent.

The effect of the wind field on the production of HSSW may be investigated applying the model of Pease [1987]. In Figure 12 we show the wind velocity field during 1 month (August–September) the winter of 1993 [Smedsrud, 1997], and it appears that the wind is systematically blowing off the shelf. Pease [1987] discussed a model for a wind-driven latent heat polynya.

In the model, frazil ice is formed at the surface layer and blown downstream against the pack ice, where it accumulates into a layer of thickness $h$. A steady state is achieved when this layer propagates upwind at the same speed as the downwind velocity of the pack ice. The ice production rate within the polynya is

$$IP = v_f h A,$$ (5)

where $h$ denotes the thickness of the frazil ice layer, $v_f$ is the drift velocity of pack ice (3% of the wind velocity $v_w$), and $A$ (300 km) is the actual length of the ice shelf. Following Pease [1987], we adopt $h = 0.1$ m, and from Figure 12 we estimate the average wind speed at $v_w = 5$ m s⁻¹. The volume of HSSW produced by the latent heat polynya is then obtained from (4) and (5) using the observed amplitude $ΔS \sim 0.15 \pm 0.05$ psu. We then obtain $V \sim (0.9 \pm 0.3) \times 10^6$ m³ s⁻¹, which is almost an order of magnitude larger than the tidal contribution derived above.

Despite our limited knowledge of the wind field outside the Ronne Ice Shelf we thus believe that the ice production due to wind and tides will account for a production of HSSW in agreement with our order of magnitude estimates derived from the current meter measurements. Quite possibly, our estimates for HSSW production underestimate the total production. Markus et al. [1998] found that the ice production in the coastal polynyas accounts for only a small fraction of the total Weddell Sea ice volume.

4.2.3. Heat fluxes. From the observations of temperature and velocity at the R2 current meter site we may estimate the order of magnitude of the heat transport into the domain defined by the floating ice shelf. We define this closed domain as the Filchner-Ronne cavity, and for the present we assume that the inflow can be derived from our R2 station. We also assume that the outflow is defined by the outflow of ISW, which is observed to take place in the Filchner Depression at the depth $z_{out} \sim 400$–600 m and with a fairly constant temperature close to the in situ freezing point (about $-2.3°C$) [Foldvik and Kvinge, 1977]. We want to use the net heat transport into the domain to obtain an estimate of the melting and heating of glacial ice in contact with seawater.
Part of the available heat is used to heat the glacial ice from its undisturbed interior temperature $\theta_{\text{ice}} \sim -24^\circ\text{C}$ [Jenkins, 1991] to the in situ melting point. The melting of glacial ice preferably takes place near the grounding line [Jenkins, 1991], whereas refreezing takes place at shallower depths. The observed mean temperature of the outflowing water ($\theta_{\text{out}}$) reflects the net result of the processes of melting, freezing, heat conduction, and mixing. We may therefore assume that all melting is taking place at this temperature and that the ice has been heated to this temperature $\theta_{\text{out}}$ before melting commences. The heat balance for the cavity is then determined by the heat fluxes into the cavity $Q_{\text{in}}$ and $Q_{\text{out}}$, by the heat flux into the glacial ice $Q_{\text{heat}}$, and by the heat required for melting glacial ice $Q_{\text{melt}}$; thus

$$Q_{\text{in}} + Q_{\text{out}} = Q_{\text{melt}} + Q_{\text{heat}}$$

The corresponding mass fluxes are denoted $M_{\text{in}}$, $M_{\text{out}}$, and $M_{\text{melt}}$. Here $Q_{\text{melt}} = L M_{\text{melt}}$, where $L$ denotes the latent heat of fusion of freshwater ice. Continuity of mass and salt require

$$M_{\text{melt}} + M_{\text{in}} = M_{\text{out}}$$
$$S_{\text{in}} M_{\text{in}} = S_{\text{out}} M_{\text{out}}$$

Let $v$ denote the velocity component normal to the ice shelf (positive out of the cavity) and $\theta$ the potential temperature. The heat flux $Q$ through an area $A$ then may be written

$$Q = -c p A \bar{v} (\theta - \theta_{\text{out}}),$$

where the bar denotes an arithmetic average of the entire record, $\rho$ is the density, $c$ is the heat capacity of seawater, and the temperature is referred to the average temperature of the outflowing water ($\theta_{\text{out}}$). In (9) we may introduce the deviations $v', \theta'$ from the mean values $\bar{v}, \bar{\theta}$:

Figure 8. Tidal ellipses for some dominating components at moorings (a) and (b) R2 and (c) RT. The orientation of the ice edge is also indicated.

Figure 9. Bottom pressure and current in direction 24° at 235 m depth at the mooring RT.

Figure 10. The clockwise ($A_c$) and counterclockwise ($A_{cc}$) circular current components for the M2 constituent at mooring R2. The bottom is indicated with the shading. The dashed lines indicate the extrapolated vertical profiles. The rotation of the current ellipse is clockwise when $A_c > A_{cc}$ and counterclockwise when $A_c < A_{cc}$.
Figure 11. CTD section obtained with R/V Lance in the summer season 1993: (a) temperature and (b) salinity. Positions of current meters R2 and RT are indicated, as well as some CTD stations.
\[ Q = -c\rho A[\bar{v}(\bar{\theta} - \theta_{\text{out}}) + \bar{\nu}' \theta'] \], \\
\bar{v}' = \bar{\theta}' = 0. \quad (10) \]

We have calculated from the observations that the turbulent heat flux \( Q' \sim \bar{\nu}' \theta' \) is small compared to the mean advection since

\[ \bar{\nu}' \theta' < 2 \times 10^{-2}, \quad \bar{v}_\text{in}(\theta_{\text{in}} - \theta_{\text{out}}), \quad \theta_{\text{out}} = -2.3^\circ \text{C}. \]

We believe that \( Q_{\text{out}} \) is even smaller than \( Q_{\text{in}} \) since the temperature fluctuations under the ice shelf must be small. Both \( Q'_{\text{in}} \) and \( Q'_{\text{out}} \) will therefore be neglected. Since

\[ M_{\text{in}} = -\rho A_{\text{in}} \bar{v}_{\text{in}}, M_{\text{out}} = \rho A_{\text{out}} \bar{v}_{\text{out}}, \quad (11) \]

we obtain from (6)

\[ LM_{\text{melt}} = M_{\text{in}}(\bar{\theta}_{\text{in}} - \theta_{\text{out}}) c - Q_{\text{heat}}. \quad (12) \]

In the following we simplify the notation by dropping the averaging bar. All variables then refer to averaged values. Equations (7), (8), and (12) provide two independent expressions for the mass of melted ice:

\[ M_{\text{melt}} = M_{\text{in}}(\bar{\theta}_{\text{in}} - \theta_{\text{out}}) c/L - Q_{\text{heat}}/L, \quad (13) \]

\[ M_{\text{melt}} = M_{\text{in}}(S_{\text{in}} - S_{\text{out}})/S_{\text{out}}. \quad (14) \]

Since

\[ Q_{\text{heat}}/L = c/L M_{\text{melt}}(\bar{\theta}_{\text{out}} - \theta_{\text{ice}}) \sim 10^{-3} M_{\text{melt}}, \quad (15) \]

the last term in (13) may be neglected compared to the left-hand side. Equation (13) then is reduced to

\[ M_{\text{melt}} = M_{\text{in}}(\bar{\theta}_{\text{in}} - \theta_{\text{out}}) c/L. \quad (16) \]

Equations (14) and (16) express a linear relationship for the rate of change in temperature and salinity between the inflow and outflow regions, namely,

\[ \theta_{\text{in}} - \theta_{\text{out}} = (S_{\text{in}} - S_{\text{out}}) L/c S_{\text{out}}. \quad (17) \]

where \( L/c S_{\text{out}} \sim 2.3 \). This straight line is similar to the one derived by Gade [1979] and Nøst and Foldvik [1994] and is shown in the \( \theta-S \) diagram in Figure 13. See also the discussion by Nicholls et al. [1997] and Nicholls and Makinson [1998], who confirmed the relationship (17) by subice measurements.

**4.2.4. Melting rates.** The amount of melting due to the heat flow at the current meter site is obtained from (16):

\[ M_{\text{melt}} = (2.5 \pm 1) \times 10^6 \text{ kg s}^{-1} = (0.8 \pm 0.3) \times 10^{11} \text{ tons yr}^{-1}. \]

The corresponding volume flux is \( 2.5 \times 10^3 \text{ m}^3 \text{ s}^{-1} \). Schlosser et al. [1990] concluded from tracer analysis of the overflowing ISW on the continental slope north of the Filchner Depression that this ISW represented a melting of \( 4 \times 10^3 \text{ m}^3 \text{ s}^{-1} \). Taking into account that tracer analysis gives total melting, in contrast to our net melting estimates, these results compare well. Jacobs et al. [1992] using mass balance considerations found the total melting under the Filchner-Ronne Ice Shelf to be \( 2.0 \times 10^{11} \text{ tons yr}^{-1} \). The difference might reflect the additional heat inflows through the Ronne Depression and also the eastern Filchner Depression.

If the melting of \( 1.0 \times 10^{11} \text{ tons yr}^{-1} \) was evenly distributed over the area of the Filchner-Ronne Ice Shelf (\( 470 \times 10^3 \text{ km}^2 \)) that would amount to 0.2 m of melted ice per year. Numerical model experiments by Gerdes et al. [1999] indicate a melt rate of only 0.1 m yr\(^{-1}\). Their model was forced by CTD data obtained during the summer season, which probably explains the discrepancy.

The melting is not evenly distributed over the underside of the floating ice shelf since local melting depends both on the local temperature difference between the seawater and the in situ freezing point, as well as on the local transport velocity.

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**Figure 12.** The wind velocity field during 1 month (August–September) the winter of 1993 [Smédra, 1997].

**Figure 13.** An \( \theta-S \) diagram with scatterplots from the two current meters at R2. Core values for some typical (HSSW, ISW, and MWDW) water masses obtained on various cruises are indicated. The continuous line represents CTD station 6 obtained in 1993, and the numbers indicate depth. The straight lines are the mixing lines with slope according to (17). The arrows indicate the water mass modifications taking place under the Filchner-Ronne Ice Shelf transforming HSSW on the Berkner Shelf into ISW flowing into the Filchner Depression.
that we shall obtain the same melting rate as above provided that the mass flow from the Ronne to the Filchner Depression is small and that the sub–ice shelf cavity circulation is dominated by closed circulation systems [Determann et al., 1994; Determann and Gerdes, 1994]. Finally, referring to Figure 13, we note that the Berkner Shelf Water occasionally attains salinities above 34.75 during the winter season. Thus a contribution to the ISW in the deep Filchner Depression may conceivably also be supplied from the Berkner Shelf.

4.3.2. Berkner Shelf–Filchner Circulation. The ISW in the Filchner Depression above 600 m depth does not fit (17) with the HSSW in the Ronne Depression (see Figure 13). However, the ISW between about 600 and 300–400 m depth does fit the winter observations from the Berkner Shelf (Figure 13). It is this ISW that forms the core of outflowing water from under the Filchner Ice Shelf and is situated above the temperature minimum (see Figures 13 and 11). This water was observed to leave the shelf break at the sill (sill depth about 600 m) of the Filchner Depression [Foldvik et al., 1985c]. Finally, the ISW overflows the sill of the Filchner Depression originated from this upper core of ISW. The implication then is as follows: Dense HSSW forms on the Berkner Shelf, especially near the ice front during winter, by enhanced freezing due to surface divergence by offshore winds [Gill, 1973] and oscillating tides (see Appendix A). The HSSW crosses the ice front on the western slope of the Filchner Shelf as buoyancy-driven current. Residual tidal forces may also be of importance [Makinson and Nicholls, 1999]. The water subsequently circulates anticlockwise around the Berkner Island as a dynamically balanced current on the slope. The HSSW is transformed into ISW through cooling and melting processes and reappears at the western slope of the Filchner Depression as a core of ISW [Foldvik et al., 1985c]. Finally, the ISW overflows the sill of the Ronne and Filchner Depression. Figure 13 shows our observations of temperature and salinity for the inflow region on the western Berkner Shelf slope (R2) together with a \( \theta-S \) station for the Filchner Depression. Also shown are typical \( \theta \) and \( S \) values for HSSW in the Ronne Depression. The thin straight lines represent the slope of (17). From an inspection of Figure 13 we offer the following comments.

4.3.1. Filchner-Ronne circulation. The Ronne Depression is occupied by high-salinity water at, or slightly below, its surface freezing point (see, for example, Figure 11). This water type does fit the characteristics of the ISW below about 600 m depth in the Filchner Depression, both water types being on the melting slope of (17) (see Figure 13) Nest and Foldvik [1994] also discussed this problem and concluded that the source water (HSSW) for the ISW in the deep Filchner Depression needed a salinity of 34.75. Since they did not find such saline HSSW between the Filchner Depression and the Ronne Depression from their examination of summer cruise data, they proposed that the ISW in the Filchner Depression originated from the Ronne Depression.

However, the discussion based on (17) does not give any information about the fluxes, and from analysis of in situ observations Nicholls and Makinson [1998, p. 316] concluded “Very little HSSW is able to pass from the ice front above the Ronne Depression, over the bedrock sill south of Doake Ice Rumples, and into the Filchner Depression.” Thus, although the water occupying the deep Filchner Depression below sill depth does have characteristics consistent with Ronne Depression HSSW, the flux of this water into the Filchner Depression may be of minor significance. Model experiments also indicate that the mass flow from the Ronne to the Filchner Depression is small and that the sub–ice shelf cavity circulation is dominated by closed circulation systems [Determann et al., 1994; Determann and Gerdes, 1994].

4.3. Implications for the Circulation Under the Ice Shelves

The \( \theta, S \) relation described by (17) is independent of the actual measured inflows and outflows. It simply predicts the changes in temperature and salinity of warmer seawater in contact with melting ice and is valid on any scale. The important assumption made in arriving at (17) for the present application is the assumption of continuity of mass, salt, and heat for the system. In the following we intend to make use of (17) to compare the water types on the Berkner Shelf and in the

**Figure 14.** The circulation of ISW below the Filchner-Ronne Ice Shelf as proposed by Foldvik and Gammelsrød [1988] and confirmed by the present study.
Filchner Depression to contribute to newly formed WSBW [Foldvik et al., 1985a]. Such a scenario for the formation of ISW was proposed by Foldvik and Gammelsrød [1988] (see Figure 14).

The scheme of circulation described above is also indicated by model experiments. Grosfeld et al. [1997] studied the circulation in a three-dimensional ice shelf cavity model connected to the ocean, including sloping side walls, and driven by buoyancy fluxes due to melting and freezing. They found that the wind-driven circulation north of the ice shelf and the thermohaline circulation beneath the ice shelf were largely separated, the ice front acting as a topographic barrier. In areas where the bottom topography varies along the ice front, for example, above the Filchner and Ronne Depressions, contours of $f/H$ cross the ice front and exchange is increased. The appearance of ISW in the two depressions (Figure 11) confirms that exchange takes place here. The solution by Grosfeld et al. [1997] shows currents along the ice front with a small component toward the ice front (compare our progressive vector diagram (Figure 3)) and with eastward return flow in a wide gyre under the ice shelf. Robust features of the solution are maximum melt rates in the southeast corner of the model associated with concentrated southward flow. Jenkins and Holland [1997] largely confirmed these results. In a study of tidal residual concentration, Makinson and Nicholls [1999] show a band of concentrated Lagrangian tracer paths along the western slope of Berkner Island.

Recently, two hot water drill holes were obtained east and south of Berkner Island [Nicholls et al., 1999]. In both locations, CTD and current measurements were made. From seismic surveys they also provide the bedrock and ice base elevations for the two sites. On the basis of these measurements and the assumption of a geostrophic flow, Nicholls et al. [1999] conclude that the volume transport of ISW is of the order of magnitude 1 Sv. These measurements agree well with the observed ISW volume flux passing the sill north of the Filchner Depression [Foldvik et al., 1985b] and also compares well with the volume flux penetrating into the cavity at the Berkner Shelf, as discussed above. Nicholls et al. [1999] estimate the width of the ISW current to about 15 km. If we assume that the travel distance from the Berkner Shelf to the Filchner Depression is 1000 km, the average melting over that stretch would be more than 7 m in 1 year.

### Appendix A: Freezing Due to Tidal Currents Near the Ice Front

Traveling tides introduce periodic fields of divergence and convergence, which influence the rate of surface freezing. This is especially important near an ice shelf or a coast where areas of open water or thin ice polynyas are often observed during the freezing season. Below we develop an expression for the net freezing outside the ice front of an ice shelf.

Let the velocity field $v$ of a tidal constituent of frequency $\omega$ vary in time $t$ as

$$v = v_0 \sin \omega t.$$  

(A1)

Within a tidal cycle the displacement $X$ of a water particle is

$$X = \frac{v_0}{\omega} \left[ (1 - \cos \omega t) \right]$$

(A2)

(see Figure A1). Ice thickness $h$ is a function of freezing time $\tau h = h(\tau)$. Freezing takes place in the domain $x \geq 0, t \geq 0$.

At a distance $x$ off the ice front, ice has been forming for a time $\tau$ where

$$x = \int_0^\tau v(t) \, dt;$$  

(A3)

hence

$$x = v_0 \omega^{-1} \left[ \cos \omega(t - \tau) - \cos(\omega t) \right].$$  

(A4)

This gives the relationship between $dx$ and $d\tau$:

$$dx = v_0 \sin \omega(t - \tau) \, d\tau.$$  

(A5)

The total ice production per unit length ($\lambda$) of the ice shelf (IP$_\lambda$) and within one tidal cycle is obtained by summing up all changes in ice thickness taking place over the stretch of freezing water ($0 \leq x \leq X$) and then integrating over the tidal period $T$:

$$IP_\lambda = \int_0^T dt \int_0^X \frac{dh}{d\tau} \, dx.$$  

(A6)

We replace $dx$ with $d\tau$ from (A5) and insert the freezing formula according to Stefan’s law [see, e.g., Leppäranta, 1993],

$$h = C\tau^{1/2},$$  

(A7)

where $C$ is given by

$$C = \frac{2\kappa_i}{\rho_i L} (\Delta T)^{1/2}.$$  

(A8)

Here $\kappa_i$ is the thermal conductivity of ice, $\rho_i$ is the density of ice, $L$ is the heat of fusion, and $\Delta T$ the (constant) temperature difference between the air temperature and freezing point. Equation (A6) then becomes

$$IP_\lambda = v_0 \frac{C}{2} \int_0^T dt \int_0^t \tau^{-1/2} \sin \omega(t - \tau) \, d\tau.$$  

(A9)

Introducing nondimensional variables, (A9) takes the form

$$IP_\lambda = v_0 C \omega^{-3/2} \int_0^{2\pi} dt \int_0^t \tau^{-1/2} \sin(\omega(t - \tau)) \, d\tau,$$  

(A10)

where the double integral is solved numerically. The result is
\[ IP = K v_0 \omega^{-3/2}, \]  
\[ (A11) \]
where \( K = 1.895. \)

The IP over the length \( l \) of the ice front and per unit of time then becomes

\[ IP = K^* v_0 \omega^{-1/2}, \]  
\[ (A12) \]
where \( K^* = K/2\pi = 0.30. \)

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A. Foldvik, E. Nygaard, and Østerhus, Geophysical Institute, University of Bergen, N-5007 Bergen, Norway. (arne.foldvik@uib.no)

T. Gammelsrød, The University Courses on Svalbard, N-9171 Longyearbyen, Norway.

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