

ICE-OCEAN PROCESSES OVER THE CONTINENTAL SHELF OF THE SOUTHERN WEDDELL SEA, ANTARCTICA: A REVIEW

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[1] Interactions between the Southern Ocean and the Weddell Sea ice shelves are important both to the Antarctic Ice Sheet and to the production of globally significant water masses. Here we review the interaction between the Filchner-Ronne Ice Shelf and the shelf sea in which it floats. The continental shelf processes leading to the production of Weddell Sea deep and bottom waters from the original off-shelf source waters are discussed, and a new view is offered

of the initial production of High-Salinity Shelf Water. Data from ship-based measurements at the ice front, from glaciological methods, and from measurements made within the sub-ice shelf cavity itself are used to describe the pattern of flows beneath the ice shelf. We also consider the variability observed within the cavity from tidal to interannual time scales and finish with a discussion of future research priorities in the region.

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1. INTRODUCTION

[2] The wind is the principal source that powers the global ocean circulation [Wunsch, 2002]. The vertical structure of water currents, however, and the properties of the water masses are largely determined by fluxes of heat and salt at the ocean boundaries (the surface and edges) and by vertical mixing. For example, loss of buoyancy at high latitudes as a result of intense cooling by the atmosphere and glacial ice leads to a sinking of cold water to the ocean abyss. As the buoyancy loss takes place at the sea surface, the dense water that ends up at the bottom of the ocean basins is also rich in oxygen and plays an important role in cooling and ventilating the deep ocean.

[3] In the southern high latitudes the cold water at the seafloor is called Antarctic Bottom Water (AABW). Around 10 sverdrups ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) of newly formed AABW is exported from the Southern Ocean, representing an important component of the global thermohaline circulation. The Weddell Sea is traditionally viewed as a major source of AABW

[e.g., Deacon, 1937; Orsi *et al.*, 1999]. Although it is unclear whether the Weddell Sea is the dominant source and estimates of the fraction of the total AABW flux that it contributes have been falling over the decades [Naveira Garabato *et al.*, 2002], it is certainly the source of the coldest and most oxygen-rich bottom waters in the Southern Ocean. The seafloor values for potential temperature, dissolved oxygen, and salinity [Orsi and Whitworth, 2005] shown in Figure 1 highlight the influence of the Weddell sector, where bottom waters that are cold and oxygen rich, but also relatively fresh, are introduced into the Southern Ocean. The key processes in the production of these bottom waters are the formation of sea ice over the broad continental shelf of the southwestern Weddell Sea and the chilling of shelf waters through contact with ice shelves. The cold and dense shelf waters resulting from these processes interact with off-shelf water masses at the continental shelf break and slope and contribute to AABW formation either directly by forming Weddell Sea Deep Water (WSDW) or indirectly by first forming Weddell Sea Bottom Water (WSBW), which is ultimately converted to WSDW.

[4] The dominant glacial feature in the Weddell sector is the huge Filchner-Ronne Ice Shelf (FRIS) floating over the southern Weddell Sea continental shelf. FRIS plays a major role in converting a sizable fraction of shelf waters into a form that is capable of descending to the deep Weddell Sea and forming WSBW [Foldvik and Gammelsrød, 1988; Foldvik

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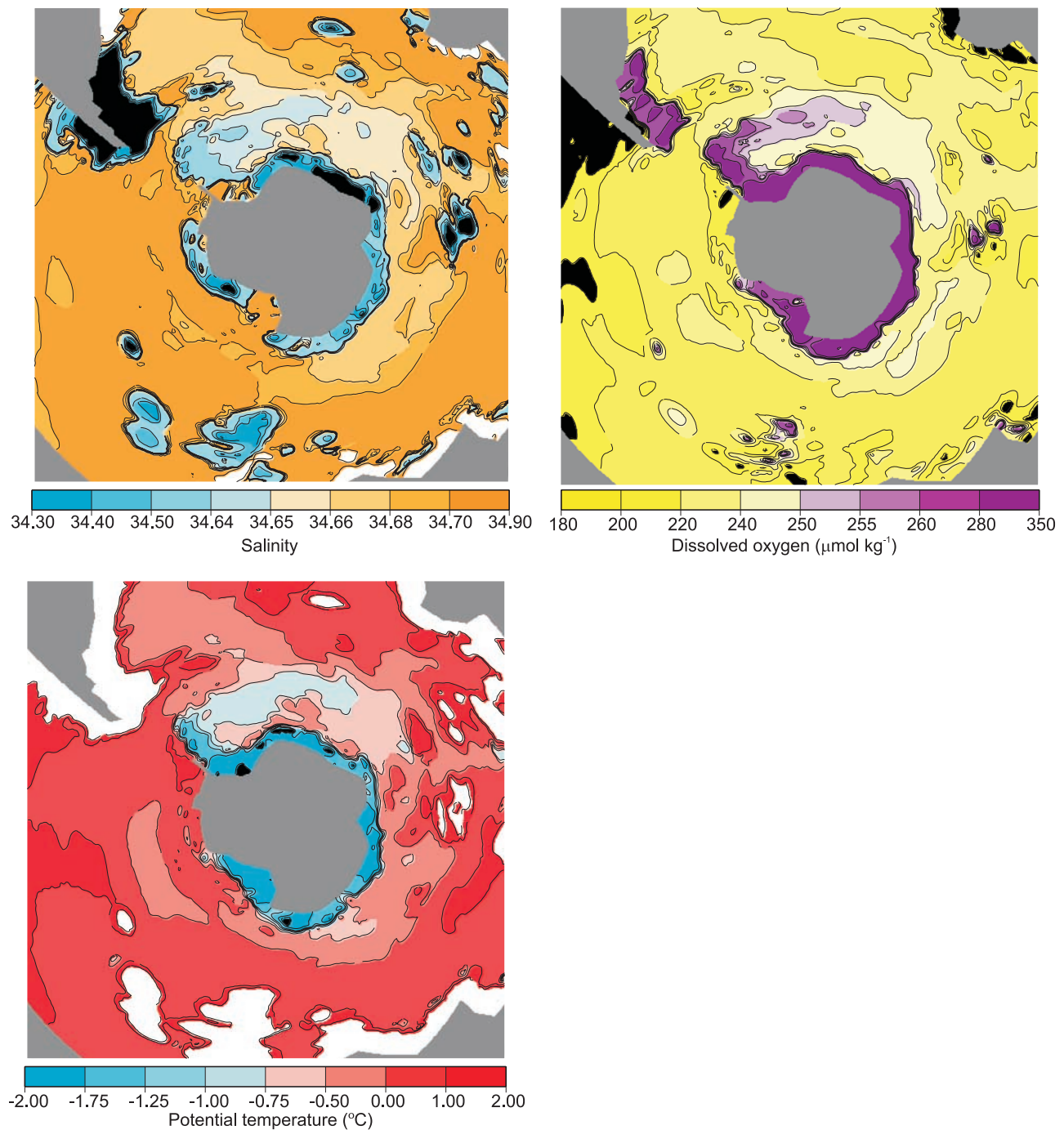


Figure 1. Bottom salinity, potential temperature, and dissolved oxygen for the Southern Ocean. Note the nonlinear color scales.

et al., 2004]. Of increasing concern is the role played by ice shelves in regulating the seaward flow of the Antarctic Ice Sheet and the consequences for future sea level rise. Recent breakouts of small ice shelves on the Antarctic Peninsula have triggered an increase in ice flux from the continent [Scambos *et al.*, 2004; Dupont and Alley, 2005] and have led to concerns that the thinning or loss of the much larger ice shelves farther south might result in a significant rise in sea level [Payne *et al.*, 2004]. There is, therefore, a clear interest in the interaction between ice shelves and the Southern Ocean

from the perspectives of both Antarctic glaciology and sea level change.

[5] The modern era of oceanographic research in the southwestern Weddell Sea began with the first International Weddell Sea Oceanographic Expedition (IWSOE), 1968–1973. The principal aim of IWSOE was to discover the mechanisms controlling deep water formation, an objective that has motivated much of the work in the region that followed. U.S., Norwegian, German, and U.K. research vessels have made a series of cruises to the area since the

first IWSOE, but the coverage was necessarily limited by sea ice conditions and the presence of the permanent ice shelf in the south. Determining the ocean conditions beneath FRIS is particularly challenging. Knowledge of the ice shelf's basal melting or freezing rate has provided important clues [Robin *et al.*, 1983; Jenkins and Doake, 1991]. Since the mid-1980s, a series of FRIS-based glaciological field campaigns has studied the response of the ice shelf to basal melting and freezing. More recently, satellite observations have allowed ice shelf-wide estimates of basal mass balance, albeit with some important assumptions. Satellite observations have also improved our knowledge of the tides over the ice shelf-covered region. Since the late 1980s, German and U.K. research groups have used hot water-drilled access holes to make direct observations of the ocean beneath the ice shelf.

[6] With the second International Polar Year it seems timely to draw together the various strands that have led to our present understanding of the processes over the southwestern Weddell Sea continental shelf, and this paper is an attempt to do that. We are aiming not at a comprehensive history of the research in the region but rather an attempt to outline the present consensus and the key lines of research that have led us there.

[7] This paper is concerned with the processes that shape the water masses of the southwestern Weddell Sea continental shelf and, in particular, the interactions between the floating ice shelves and the shelf seas. We will review our present understanding of the oceanographic regime of FRIS by following the source water mass from where it arrives on the continental shelf to where it leaves the shelf to contribute to the production of deep and bottom waters. We will discuss the processes that act on and modify the water mass at each stage.

[8] Section 2 discusses the source waters for the continental shelf regime and how they gain access to the continental shelf. The conversion of the source waters to High-Salinity Shelf Water (HSSW) is considered in section 3, while section 4 addresses the question of how the HSSW so formed is able to enter the sub-ice shelf cavity. Processes that take place within the cavity itself are discussed in section 5, and known sources of temporal variability are described in section 6. Section 7 concludes the paper with an outlook for future research in the southwestern Weddell Sea region.

1.1. Geographic Setting

[9] Ice shelves form when the ice at an ice sheet's oceanic boundary does not calve as icebergs at the point where it goes afloat (the grounding line) but remains connected to the grounded ice sheet. Ice shelves may thus be regarded as floating extensions of the ice sheet, and whether an ice shelf forms depends to a large extent on the coastal geometry. The ultimate fate of the ice in an ice shelf is either to be calved off as icebergs from its seaward edge, the ice front, or to be melted from the ice shelf base.

[10] FRIS, lying in the southern Weddell Sea (Figure 2a), is by ice volume the largest of the Antarctic ice shelves. It has an area of about 450,000 km², with typical horizontal

dimensions of 500 km. The thickness of the ice shelf has been mapped by using aircraft-based downward looking radar [Robin *et al.*, 1983] and, under the assumption that the ice shelf floats in hydrostatic equilibrium, by inverting satellite altimetry [Vaughan *et al.*, 1995]. Lambrecht *et al.* [2007] have compiled ice thickness data for FRIS from various sources. The average thickness is around 700 m, with a maximum of around 1800 m at the deepest grounding lines.

[11] The depth of the bedrock beneath FRIS has been determined using seismic techniques (Figure 2b). An extensive Russian seismics campaign was conducted during the 1970s [Pozdeyev and Kurinin, 1987] and was supplemented in some areas during the 1990s by British [Johnson and Smith, 1997] and German [Lambrecht *et al.*, 1997] groups. The Filchner Ice Shelf (FIS) lies on the eastern side of the Filchner-Ronne embayment. Beneath FIS there is a deep trough that runs out across the continental shelf, intersecting the shelf break to form the Filchner Sill. The trough, known as the Filchner Depression, reaches a maximum depth of 1800 m at the deepest grounding lines beneath FRIS. Mirroring the Filchner Depression, the Ronne Depression is a weaker feature that lies beneath the western side of the Ronne Ice Shelf. It is smaller in extent and does not reach far north of the ice front. The continental shelf extends another 500 km seaward of the ice front with seabed depths of between 300 m and 500 m, except for within the Filchner and Ronne depressions.

1.2. Oceanographic Context

[12] This paper is primarily concerned with the modification of the temperature and salinity of water masses in the southern Weddell Sea. Water masses in the Southern Ocean are usually defined in terms of a mix of characteristics, namely, their neutral density, their potential density, and their potential temperature and salinity [Whitworth *et al.*, 1998]. Our definitions will use potential temperature (θ) and salinity (S) only, which is in keeping with traditional practice for the Weddell Sea region. The properties of the key water masses are shown in the θ - S diagram in Figure 3.

[13] The Antarctic Circumpolar Current, girdling the continent of Antarctica, for the most part is composed of Circumpolar Deep Water (CDW). Beneath the CDW is AABW, with Antarctic Surface Water above. Closer to the continent, within the Weddell Basin, there exists an elongate cyclonic (clockwise) gyre reaching from the Antarctic Peninsula in the west to around 40°E (Figure 2a). CDW enters at the northern and eastern edges of the gyre and is converted within the gyre to the slightly cooler and fresher Warm Deep Water (WDW). Fahrbach *et al.* [1994a] found that boundary currents associated with the basin's continental slopes account for 90% of a 29.5-Sv gyre circulation.

[14] Within the Weddell Sea, the water mass beneath WDW is WSDW [Orsi *et al.*, 1993; Fahrbach *et al.*, 1994a], the Weddell Sea's manifestation of AABW [Foldvik *et al.*, 1985a]. The bulk is in transit from the east, but some is newly formed at the continental margins of the Weddell Sea, and still more is formed from upwelled deeper and colder

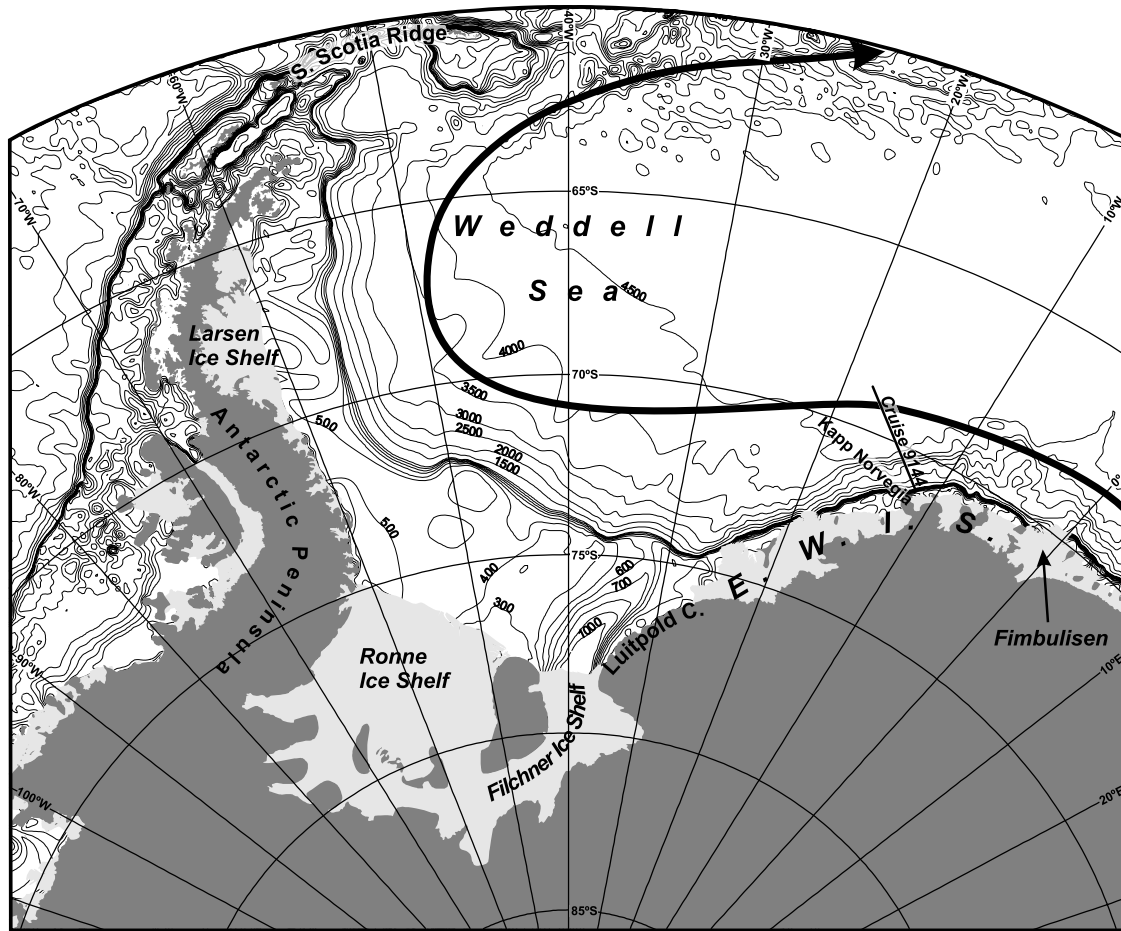


Figure 2a. Map of the western Weddell Sea. Contours are of bathymetry and are from BEDMAP [Lythe and Vaughan, 2001] and the *British Antarctic Survey* [2000]. Areas shaded in light gray are ice shelves; those in dark gray are land. The large arrow is a schematic representation of the western portion of the Weddell Gyre. The line labeled “Cruise 9144” from Kapp Norvegia shows the track of a hydrographic section discussed in the text.

WSBW. WSDW is light enough to be able to escape the Weddell Basin via passages through the South Scotia Ridge. WSBW lies below WSDW and has its origins in processes over the continental shelves: cold, dense waters descend the continental slope, entraining WDW and WSDW from above. Too dense to escape the Weddell Basin, WSBW is destined to remain in the Weddell Gyre until it mixes up through the water column and is converted to WSDW.

[15] Above the Warm Deep Water lies the Antarctic Surface Water (ASW). Heat loss to the atmosphere during the Antarctic winter cools the ASW to the freezing point. Further heat loss causes the production of sea ice, and the associated salt rejection results in convective overturning within the upper 100–200 m. This results in a mixed layer of Winter Water (WW), which has a higher salinity as a result of diapycnal mixing with the underlying, more saline WDW and the salt rejection due to freezing. During the summer, the warming of the surface waters and melting of sea ice restratifies the upper part of the water column to reestablish the layer of ASW.

[16] The structure of the pycnocline that separates the upper mixed layer from the warmer waters below plays an

important role in controlling the access of those warmer waters to the southern continental shelf. In the center of the gyre the combination of upwelling and convective overturning has the effect of reducing the mixed layer thickness and sharpening the pycnocline (WW-WDW interface). However, along the southern continental shelf break of the eastern Weddell Sea the prevailing easterly winds generate an onshore surface Ekman flux that depresses the pycnocline. *Fahrbach et al.* [1994b] found that an additional, important contribution to the deepening of the interface was provided by wintertime convection in the coastal polynyas, which often extend beyond the region’s very narrow continental shelf. The south-north horizontal density gradient that results from the southward deepening of the isopycnals is termed the slope front, and the associated baroclinic current component is the slope front current [Whitworth et al., 1998].

[17] In the vertical, the slope front is substantially broader than the pycnocline farther offshore. The broadening and deepening of the pycnocline is illustrated by the potential density section in Figure 4 [Fahrbach et al., 1994a], the track for which is given in Figure 2a (cruise 9144). Whichever processes are responsible for the thickening of the pycno-

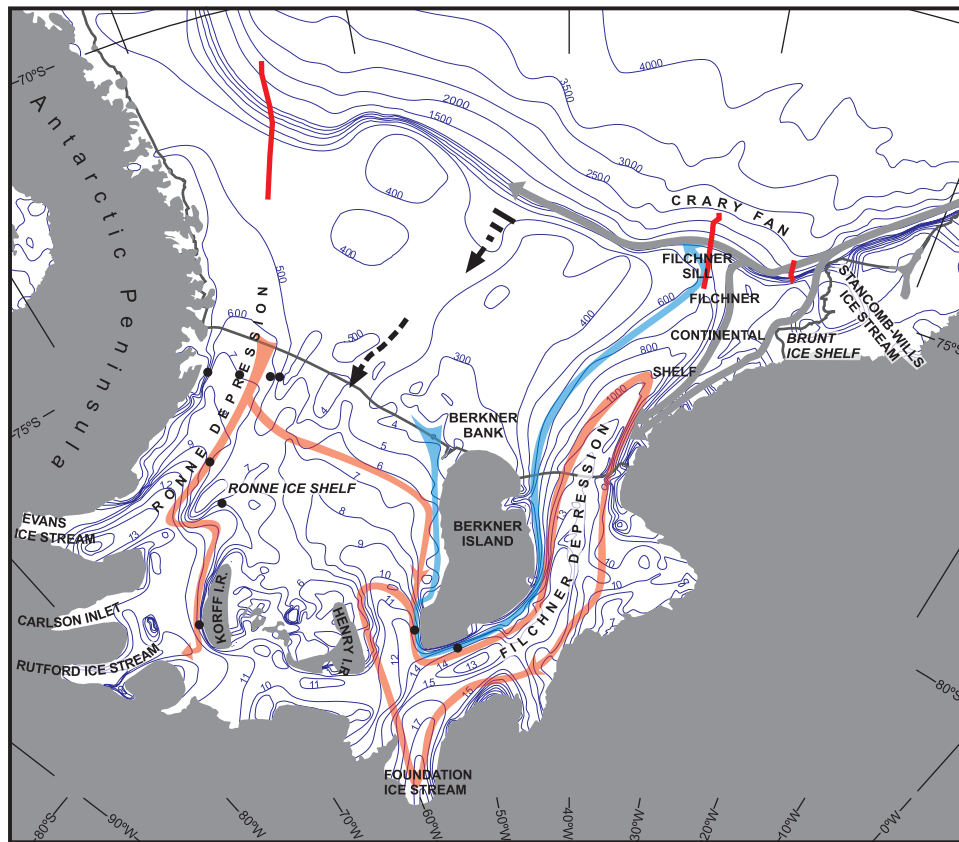


Figure 2b. Map showing the southwestern Weddell Sea. Bathymetric contours are labeled in hundreds of meters beneath the ice shelves. Ice fronts are shown by black lines. Black circles show drill hole locations. The gray arrows show the flow of the slope front and coastal currents. Blue and red arrows show the circulation beneath the ice shelf of waters originating from the eastern and western ends of the Ronne Ice Front, respectively. The top thick black broken arrow indicates the location of an MWDW inflow at the central trough in the Ronne continental shelf break, and the bottom thick black broken arrow indicates the location along the ice front where an MWDW core is observed. The three solid red lines at the shelf break show the position of the sections in Figure 8.

cline, the result is important, as the transition water mass between WW and WDW, Modified Warm Deep Water (MWDW) (Figure 3), plays key roles both as the likely source water for the shelf water masses and as a principal constituent of WSBW in one of the processes thought to be responsible for bottom water production.

[18] The final water mass that contributes to the oceanographic context of the southwestern Weddell Sea is found south of the slope front, over the narrow continental shelf of the southeastern Weddell Sea. This coastline is fringed by the eastern Weddell Sea ice shelves (EWIS), small ice shelves that occasionally reach or even overhang the continental shelf break. The narrow shelf is flooded with WW, diluted slightly by an admixture of glacial meltwater from the EWIS [Fahrbach et al., 1994b], with an additional contribution from MWDW incursions across the shelf break (Figure 3). Termed Eastern Shelf Water (ESW), the mixture has a lower density than the offshore WW, and the resulting south-north density gradient forces a coastal current. There is also some evidence for a current following the topographic step of the ice fronts. This is likely to be driven by a combination of

easterly winds, the density difference between the sub-ice shelf and open continental shelf regimes [Heywood et al., 1998] and by tidal residuals [Makinson and Nicholls, 1999].

2. SOURCE WATERS FOR THE SOUTHERN WEDDELL SEA CONTINENTAL SHELF

[19] HSSW, the key water mass in the production of WSBW and, ultimately, AABW, is formed over the southern Weddell Sea continental shelf as a result of brine rejection during sea ice formation. HSSW has several potential source waters: upper mixed layer waters crossing the continental shelf break, transported southward in a surface Ekman layer; waters conveyed along the narrow continental shelf to the east of the FRIS embayment before turning southward toward the Filchner Ice Shelf; and MWDW crossing directly southward across the shelf break north of FRIS.

2.1. Surface Ekman Flux

[20] Gill [1973] included the Ekman flux in his estimates of the vertical circulation across the shelf. His analysis

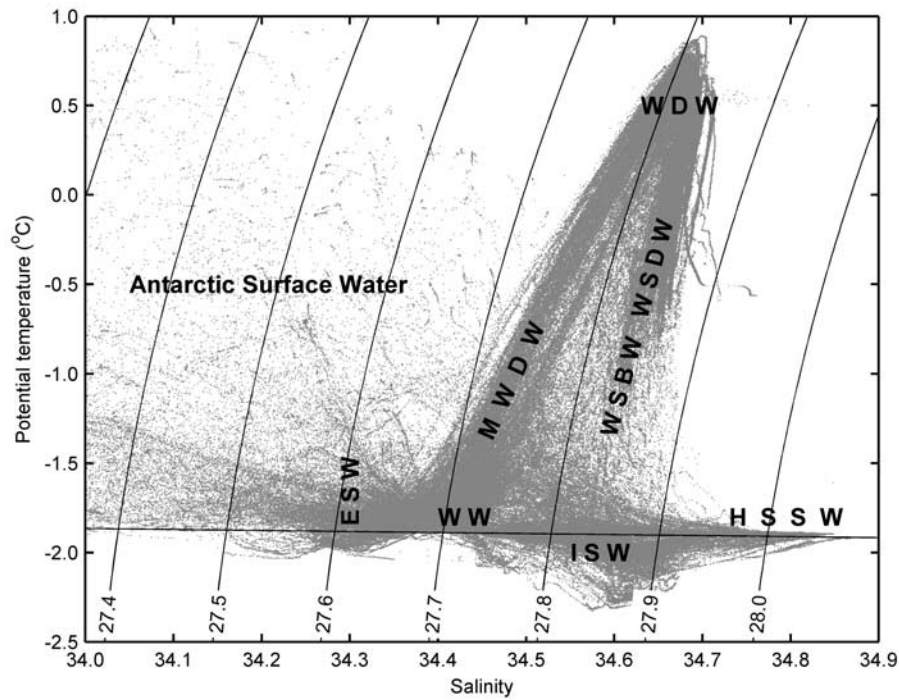


Figure 3. A θ - S diagram showing data from 554 CTD profiles from the Weddell Sea south of 70°S and west of 0°. The diagram shows approximate θ - S characteristics for the water mass types mentioned in the text. Isopycnals are referenced to surface pressure, and the near-horizontal line shows the surface pressure freezing temperature [Fofonoff and Millard, 1983]; all data below that line are from Ice Shelf Water.

indicated an Ekman flux of about $0.5 \text{ m}^3 \text{ s}^{-1}$ per meter of shelf break. For a shelf break of length around 1000 km, this suggested a large flux of surface waters being driven southward onto the continental shelf. In this calculation, Gill's estimate for the surface stress was 0.07 Pa. Given an effective

drag coefficient between air and water (usually via pack ice) of ~ 0.002 [Uotila et al., 2000], this would suggest an average wind speed of around 5 m s^{-1} . Kottmeier and Sellmann [1996] find weaker average wind strengths, around 1 or 2 m s^{-1} , suggesting that Gill's estimate of 0.5 Sv southerly flow is

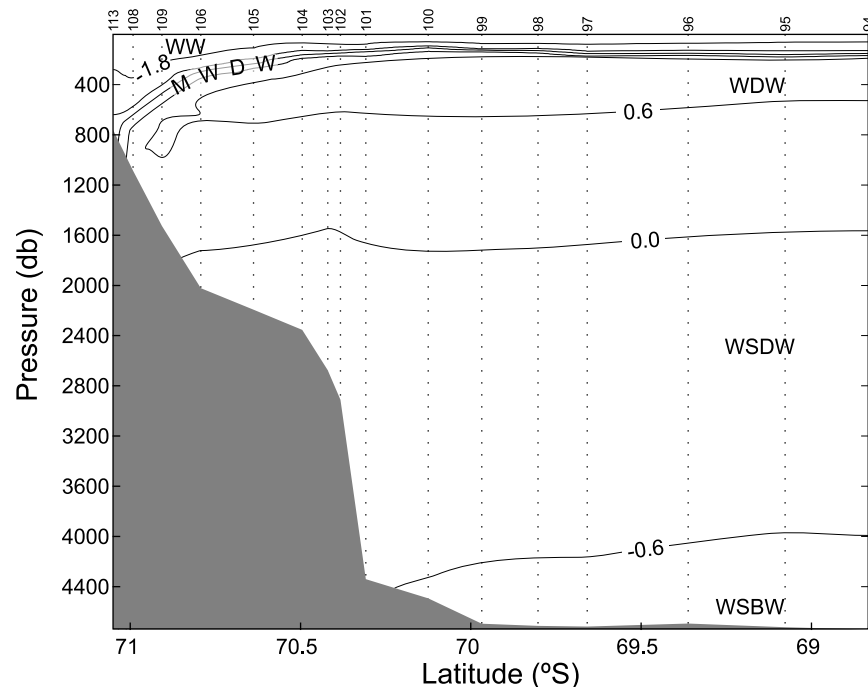


Figure 4. Potential temperature section from the central Weddell Sea to the southern Weddell Sea coast, obtained during cruise ANTARKTIS IX/2 of R/V *Polarstern* [Bathmann et al., 1992]. The track for the section is shown in Figure 2a (as cruise 9144).

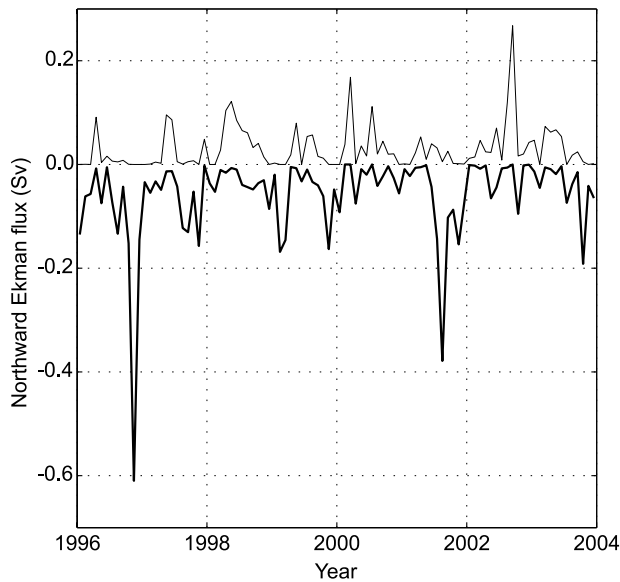


Figure 5. Ekman transport across $73^{\circ}5.5'S$ using ECMWF winds. The southward (thick line) and northward (thin line) contributions were integrated separately.

an overestimate. Figure 5 shows the surface Ekman fluxes perpendicular to $73^{\circ}7.5'S$ calculated using wind fields from the European Centre for Medium-Range Weather Forecasts (ECMWF) operational surface analyses, provided by the ECMWF data server. They are calculated from ECMWF wind data (1996–2004) in 6-h intervals and a grid size of 1.125° . Monthly means of the wind components were formed for each grid point, and the wind stress was calculated using a drag coefficient of 1.8×10^{-3} . For the wind stress calculation, the MATLAB routine WSTRESS which computes wind stress using the formulation from *Large and Pond* [1981], was used (see <http://woodshole.er.usgs.gov/operations/sea-mat/RPSstuff-html/wstress.html>). The northward (southward) Ekman transport, F_{Ekman} , was then calculated from the eastward (westward) wind stress, τ , with $F_{\text{Ekman}} = \tau / (f\rho_w)$, where f is the Coriolis parameter and ρ_w is the water density. All northward (positive) and southward (negative) transports were integrated separately along $73^{\circ}7.5'S$ from $60^{\circ}11.25'W$ to $18^{\circ}37.5'W$, a distance of a little under 1350 km.

[21] Figure 5 shows that there are periods when there appears to be a strong, time-limited southward flux (late 1996, for example), and such events might have short-term consequences for the flux of newly formed WSBW and WSDW. There is little evidence, however, of a significant contribution to the overall mean flushing of the continental shelf (<0.1 Sv).

2.2. Transport From the Eastern Continental Shelf

[22] The water flowing along the EWIS continental shelves, forced by a combination of ice front processes and direct wind action, contributes to the Filchner continental shelf regime. The size of that contribution is as yet unknown, but whether or not it is the dominant source for the entire FRIS continental shelf, as suggested by *Markus et al.* [1998],

it certainly influences conditions on the shallow shelf on the eastern side of the Filchner Depression.

[23] Conductivity-temperature-depth (CTD) sections have been only sporadically obtained in the area where the coastal current enters the Filchner continental shelf. The relevant sections are indicated in Figure 6, labeled with the year in which they were acquired. We use these sections to trace the likely flows from the east. As the continental shelf broadens west of the Stancomb-Wills Ice Stream, the flow that had been following the shelf break splits into two branches: one carrying along the shelf break and the other heading south, initially along the 400–450 m depth contour. The southward branch skirts around the Brunt Ice Shelf, evidently not interacting directly with the ice shelf itself, and arrives at about $76^{\circ}S$ as a 200-m thick layer between $29^{\circ}W$ and the coast (Figure 7a). From there the water follows the coast at the eastern margin of the Filchner Depression, occupying the entirety of the somewhat thinner water column as seen in the section at $77^{\circ}S$ (Figure 7b).

[24] The gray arrow in Figure 6 represents the coastal current that undoubtedly exists during the summer (Figure 7) and is presumably composed of water originally following the EWIS ice fronts and possibly water from the EWIS cavities. The eastern part of the 2003(1) section exhibits distinct Ice Shelf Water characteristics in the deeper half of the water column.

[25] All the CTD measurements in the area have been made during summer, and the majority of the measurements are from different years, spread over almost 3 decades. We do not know the nature of any seasonal variability, though it is likely that during winter the water column over the shallower

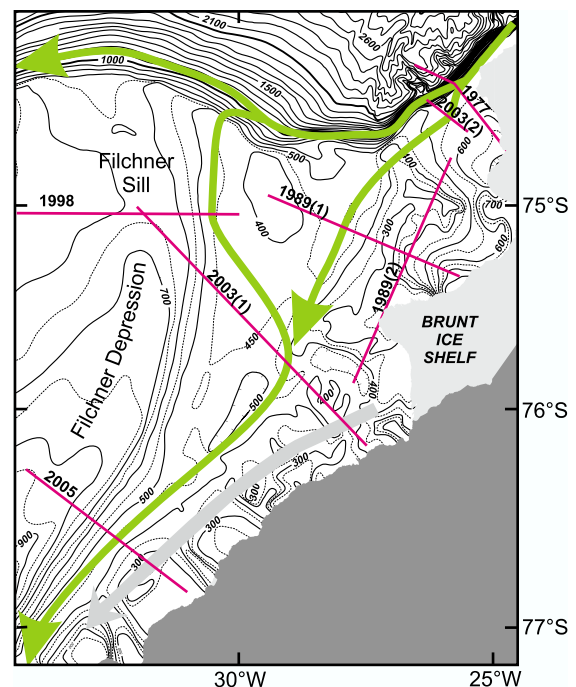
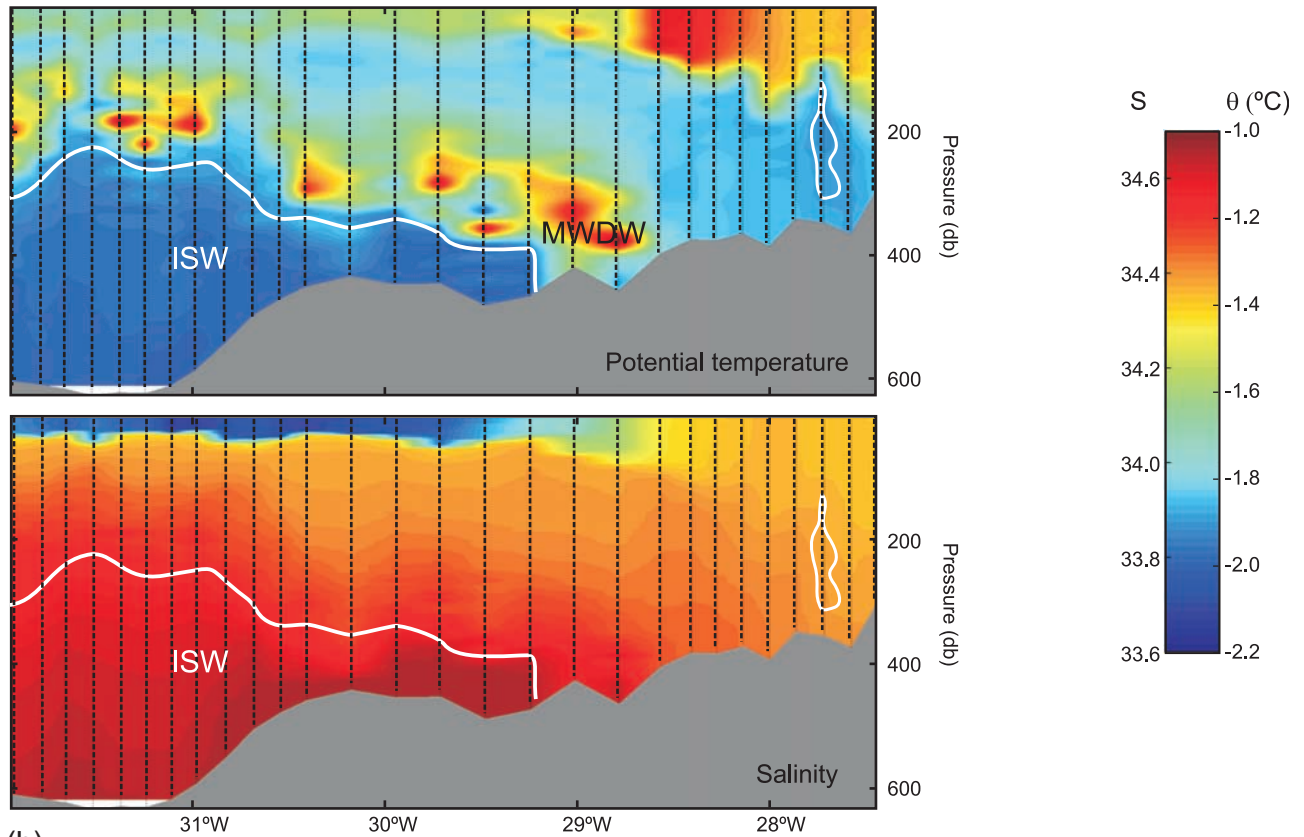


Figure 6. Map showing inflow pathways onto the Filchner continental shelf, determined from the indicated hydrographic sections. The green arrows represent flows of MWDW; the gray arrow is a coastal current.

(a)



(b)

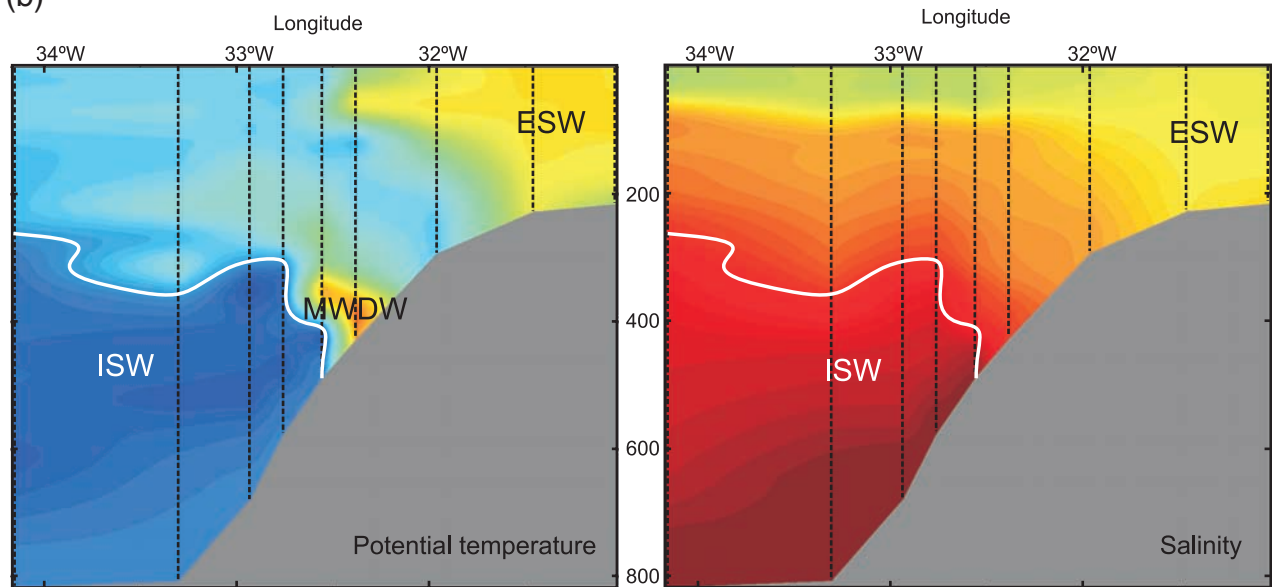


Figure 7. Potential temperature and salinity sections from (a) RISOC (2003(1)) and (b) JR97 (2005) cruises. The tracks for the sections are shown in Figure 6. The white contours indicate the position of the isotherm for the surface freezing point; all colder water is Ice Shelf Water.

continental shelf (that is, not within the Filchner Depression itself) will be entirely mixed and cooled to the freezing point as a result of sea ice production.

[26] A lowered acoustic Doppler current profiler (LADCP) section was obtained at the same time as the temperature and salinity sections in Figure 7b. The LADCP data were detided using CATS02.01, a barotropic tidal model [Padman *et al.*,

2002]. The results indicate maximum fluxes of less than 0.4 Sv. As will be discussed in section 3, we expect an HSSW production rate of ~ 3 Sv, implying that the ESW contribution from the coastal current is relatively small, certainly during the late summer when the data were obtained. The wintertime contribution of ESW to the shelf remains unknown.

2.3. Direct Transport of MWDW Across the Continental Shelf Break

[27] The final route for water getting onto the southern Weddell Sea continental shelf is direct flow of MWDW across the continental shelf break. A perennial MWDW incursion is seen on the eastern flank of the Filchner Depression. Crossing the Filchner Sill at around 32°W, this feature can be traced at 400 dbar for at least 200 km (Figures 7a and 7b), and there is some summertime evidence that it intermittently reaches the Filchner Ice Front [Foldvik *et al.*, 1985b].

[28] Extensive incursions of MWDW have been observed at various locations over the shelf north of the Ronne Ice Shelf [Nicholls *et al.*, 2003], a notable and apparently perennial example being the incursion that reaches as far south as the Ronne Ice Front at around 53°W. This incursion is the only one for which year-round data are available. Foldvik *et al.* [2001] present the results from a mooring deployed near the ice front at the location of the observed MWDW core. The mooring included a current meter and temperature and conductivity sensors at the approximate depth of the core (245 m) and showed that in 1993 traces of the MWDW characteristics survived through the winter until early September. From that point through to late November the water column was well mixed and at the freezing point. The MWDW characteristics then returned.

[29] The results reported by Foldvik *et al.* [2001] imply a flow of MWDW steered along the ice front at a mean speed of around 10 cm s^{-1} , with no mean component into the cavity. It is therefore unlikely to play an important direct role beneath the ice shelf. Nicholls *et al.* [2008] used instrumented Weddell seals to obtain a wintertime CTD survey over the central continental shelf north of the Ronne Ice Shelf. The results showed a coherent full-depth inflow, estimated at $\sim 2.6 \text{ Sv}$, leading Nicholls *et al.* [2008] to conclude that this is the route for the majority of the inflowing water. The mean temperature and salinity of the inflow was -1.76°C and 34.50, respectively.

[30] The mechanism by which MWDW intrudes onto the continental shelf is not clear. We follow the mixing path arguments of Foster and Carmack [1976] to show whether potential energy has to be supplied to raise MWDW onto the shelf. They showed that west of the central Filchner Sill no potential energy was required for off-shelf waters to mix onto the continental shelf. In Figure 8 we have shown neutral paths calculated using more recent CTD sections (the tracks of which are shown in Figure 2b), which indicate possible trajectories along which water parcels would feel no buoyancy forces. These are slightly different from the mixing paths of Foster and Carmack [1976], where the water parcels exchange their properties with the ambient water en route. The paths in Figure 8 demonstrate that the potential energy obstacle starts to be removed at the transition at around 28°W from a steep continental slope to the gentler slope associated with the Crary Fan. At a longitude of 32°W the off-shelf waters to a depth of around 1200 m can move freely onto the shelf. Farther west, at a longitude of 55°W, water parcels from any depth in the profiled water column are able to rise onto the shelf without an increase in potential energy. Foster *et al.*

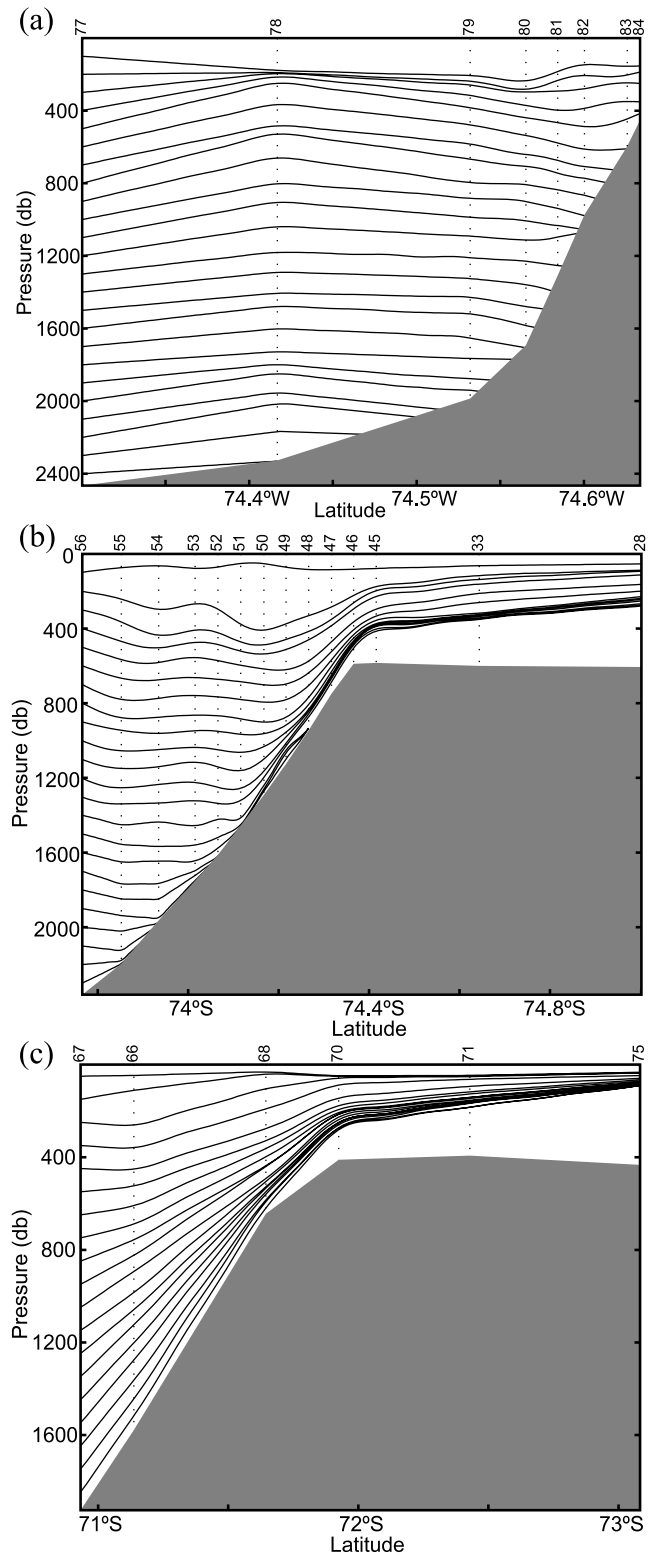


Figure 8. Neutral paths for sections across the southern Weddell Sea slope at longitudes (a) 27°W, (b) 32°W, and (c) 55°W. The positions of the sections are shown as red lines in Figure 2b.

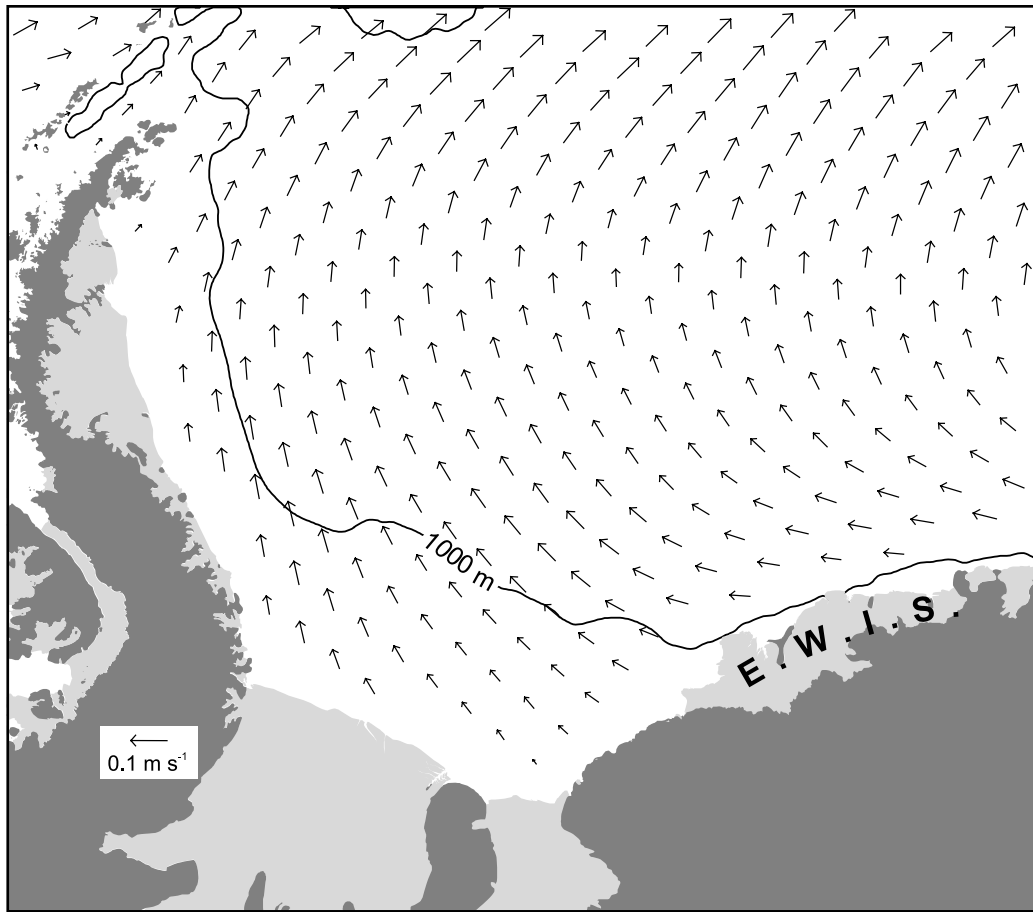


Figure 9. Vectors showing ice velocity averaged over the months March–November for years 1979–1997. Data are from C. Schmitt et al. (2004, available at <http://imkhp7.physik.uni-karlsruhe.de/~eisatlas/>). The 1000-m contour has been included to indicate the line of the continental shelf edge.

[1987] suggest that continental shelf waves acted as the agent to transport MWDW onto the shelf, though they consider this process to be confined to the shelf break region, where the MWDW is mixed with HSSW before descending the slope to form WSBW.

3. PRODUCTION OF HSSW

[31] The key processes concerned with HSSW formation are related to the sea ice budget over the continental shelf. This comprises import of sea ice from the northeast, export of sea ice northward, mainly wintertime sea ice formation, and summertime melting. Feature tracking using data from satellite-borne radiometers (C. Schmitt et al., *Atlas of Antarctic Sea Ice Drift*, 2004, available at <http://imkhp7.physik.uni-karlsruhe.de/~eisatlas/>) yields ice motion vectors that suggest the import of sea ice to be a small component in the balance (Figure 9). Although these data sets do not resolve the rapid sea ice advection that might occur in the narrow coastal current sweeping around the EWIS, the implication is that the majority of the sea ice observed over the southern continental shelf is formed locally.

[32] Sea ice production and its subsequent export from the continental shelf acts to fractionate the water arriving on the

shelf from the north. Fresh water is exported in the form of sea ice, leaving the cold, saline component, HSSW. To generate HSSW, heat must first be removed to cool the seawater to the freezing point, and then additional heat must be lost to create sufficient sea ice to raise the salinity of the water to that of HSSW. For reasons mentioned in section 2, we will proceed on the assumption that the principal source water for the southern continental shelf is MWDW directly transported across the shelf break and not ESW.

3.1. Rate of HSSW Production

[33] A lower bound on the estimate of the rate of conversion of MWDW to HSSW is given by estimates of the flow of HSSW across the Ronne Ice Front into the sub-ice shelf cavity. Nicholls et al. [2003] used current meter moorings along the western Ronne Ice Front to yield an estimate of 0.9 ± 0.3 Sv of inflowing HSSW. Combined with the estimate of Foldvik et al. [2001] for the inflowing HSSW for the eastern portion of the Ronne Ice Front, $0.3\text{--}0.6$ Sv, the total inflow at the ice front is estimated to be 1.4 ± 0.4 Sv. This is consistent with the finding of Foldvik et al. [2004] that about 1.6 ± 0.5 Sv of water at a temperature below -1.9°C (that is, Ice Shelf Water (ISW)) leaves the continental shelf via the Filchner Sill. The flux at the Filchner Sill is likely to have

been enhanced by an admixture of MWDW and HSSW within the Filchner Depression [Grosfeld et al., 2001]. There is a possible flaw in this calculation. If ISW leaves the Ronne Ice Front and is then reconverted to HSSW before reentering the cavity, the flux of HSSW would remain correct, but the water being converted would no longer be pure MWDW. Although there are locations where ISW is seen to emerge at the Ronne Ice Front before being steered along the ice front and reentering the cavity farther west, $\delta^{18}\text{O}$ data suggest that none of the HSSW entering the Ronne cavity is reconverted ISW [Nicholls et al., 2003].

[34] Some HSSW does not enter the sub-ice shelf cavity but is thought to leave the continental shelf via a mechanism first proposed by Gill [1973]: mixing of HSSW with MWDW at the shelf break to form WSBW that then descends the continental slope. The idea was subsequently developed by Foster and Carmack [1976]. There are difficulties in estimating the volume of HSSW involved in this process, the principal one being the inaccessibility of a region prone to heavy sea ice cover, usually year round. However, by analyzing the stable isotopes of helium and oxygen in the deep waters sampled during the drift of Ice Station Weddell, Weppernig et al. [1996] derived the relative contributions of ISW and HSSW to the sub-0°C component of the water column. The southern end of the drift is at the western extreme of the southern Weddell Sea continental slope (Figure 2), and Weppernig et al. [1996] calculated an equal contribution from the two sources. Thus, our best estimate of the HSSW production rate over the southern continental shelf is around 2.8 Sv.

[35] This estimate does not include HSSW that exits the southern continental shelf by being advected northward along the Antarctic Peninsula coast to the continental shelf east of the Larsen Ice Shelf. Gordon [1998] suggests that the HSSW they found flowing down a shelf edge canyon at a latitude of about 70°S originated from the southern shelf. That flow should therefore be included in our HSSW production budget. We do not know the size of the contribution, however, and assume it to be small.

[36] Another uncertainty is connected with the assumption that the water sampled during experiments at Ice Station Weddell contained the full ISW signal from the Filchner Sill. Foldvik et al. [2004] suggest that some of the ISW heads north from canyon features west of the sill (at around 36°W) rather than along the slope. If this is the case, the amount of HSSW leaving the shelf without first being converted to ISW is likely to have been overestimated. For example, if only two thirds of the ISW contribution had been sampled during experiments at Ice Station Weddell, then the total HSSW production rate would have been 2.3 Sv rather than our estimate of 2.8 Sv.

3.2. Heat Loss Required

[37] For a given sea ice production rate, the rate of formation of HSSW depends on the residence time of the water over the shelf [e.g., Markus et al., 1998]: the longer it is there, the smaller the flux but the higher the salinity. The residence time depends on both the rate of supply of MWDW

to the shelf and the rate at which the HSSW and fresh water (in the form of sea ice) can be removed from the shelf [Gill, 1973]. Nicholls et al. [2008] show that the MWDW reaching onto the shelf has a temperature and salinity at the shelf break of around -1.76°C and 34.50, respectively. Typical HSSW characteristics are -1.90°C and 34.75. To become HSSW, therefore, the temperature of the MWDW needs to decrease by 0.14°C , and then sufficient ice needs to form to increase the salinity by 0.25. Over a year, the total heat loss required is given by

$$H_T = F\rho_w T_Y \left(c_w \Delta T + \frac{L\Delta S}{S_{\text{MWDW}} - S_{\text{ice}}} \right),$$

where F is the flux of water to be converted (2.8 Sv), ρ_w is the density of seawater (1027 kg m^{-3}), T_Y is the number of seconds in a year, L is the specific heat of fusion of ice ($335,000 \text{ J kg}^{-1}$), c_w is the specific heat capacity of seawater ($4000 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$), S_{MWDW} is the salinity of MWDW (34.50), S_{ice} is the salinity of young sea ice (~ 5), and ΔT and ΔS are the required changes in temperature and salinity. From this, $H_T \approx 3 \times 10^{20} \text{ J}$. We note that around 80% of the heat loss is needed to accomplish the increase in salinity. The increase in salinity requires the formation of an average of 2.2 m of ice per year across the $370,000\text{-km}^2$ continental shelf.

[38] We have already suggested that to a good approximation all the sea ice over the southern continental shelf is formed locally. The system then lends itself to a simple analysis to determine the total heat loss from the water column during the winter. We assume that the ice is advected away from the fronts of the ice shelves at a constant speed, its rate of thickening diminishing as the ice moves to the north and becomes a more effective insulator. We start the simple model from what we assume to be a typical end-of-summer ice distribution, thereby incorporating the effect of a summer shore lead (50 km wide) and thinner ice cover (0.5 m thick). Eisen and Kottmeier [2000] use a thermodynamic model to calculate the ice growth in newly formed leads in the Weddell Sea, and we refer to their table of heat loss as a function of ice thickness. A correction to the heat loss figures was necessary to account for the insulating effect of snow cover. The snow thickness was allowed to build at a rate of 150 mm a^{-1} (water equivalent) on the basis of precipitation results from the regional model of van Lipzig et al. [2004], with an assumed density of 400 kg m^{-3} . The correction was applied by multiplying the Eisen and Kottmeier [2000] heat fluxes by the factor $(1 + k_i h_s / k_s h_i)^{-1}$, where k and h are thermal conductivity and layer thickness, respectively, and the subscripts i and s refer to the ice and snow layers, respectively. The value for the thermal conductivity of the snow was taken from the results of Singh [1999].

[39] When calculating the ice growth, we need to take into account the heat flux from the ocean. We estimate the heat flux by assuming that 2.8 Sv of MWDW flows onto the shelf and that this water needs to be cooled to the freezing point. We are therefore prescribing the inflow of MWDW into the shelf regime, and our calculation of ice growth indicates the

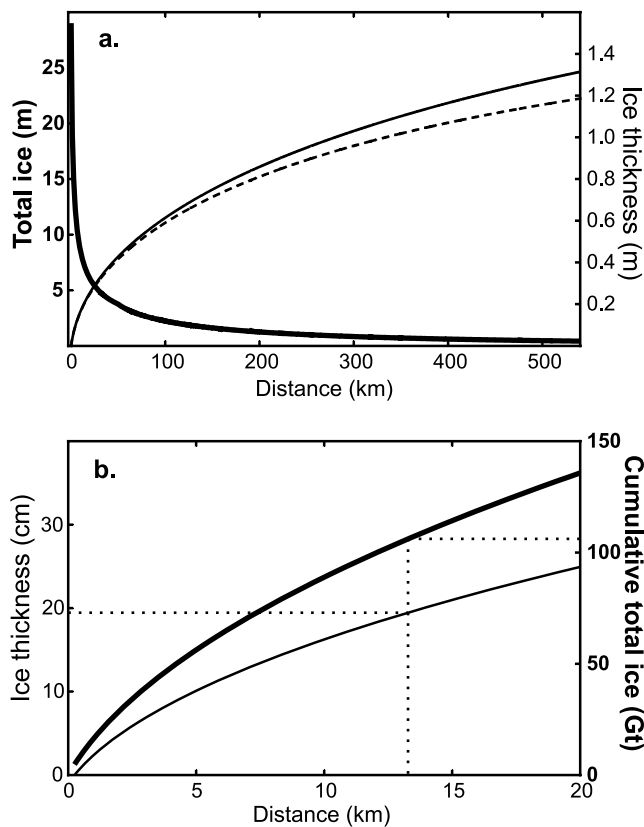


Figure 10. The modeled ice growth across the southwestern continental shelf of the Weddell Sea for 1 March to 5 September. (a) The heavy line shows total ice formation during the period; the dashed line indicates the final thickness distribution for the combined ice and snow cover. The thin line indicates the thickness of the ice alone. (b) Ice growth near the “shore lead.” The thin line again shows the ice thickness; the heavy line is the seasonal ice formation integrated from the ice front out across the continental shelf, multiplied by the length of the continental shelf. The straight dotted lines indicate the extent of the shore lead (see text for explanation).

ability of the shelf regime to increase the salinity of that flux of MWDW.

[40] Figure 10 shows the result of running this simple model over the period from 1 March to 5 September, the period for which the *Eisen and Kottmeier* [2000] values have been averaged and, from the work of *Renfrew et al.* [2002], the period over which most of the ice in the shore lead is formed. The graph shows the ice and snow thickness profile from ice front to shelf break for the end of the winter together with the total ice formation. The average ice formation (that is, the mean of the total ice formation) is 2.3 m, close to the 2.2 m required to produce 2.8 Sv of HSSW.

[41] This calculation assumed a lead fraction of 2%, with the high accompanying rates of heat loss. Ice formed in the leads did not contribute to the overall thickness of the ice cover, although it did contribute to the figure for the total ice formed. The value of 2% for the lead fraction was selected on the basis of interpretation of passive microwave data sets and

on the basis of results from *Padman and Kottmeier* [2000], but the lead fraction is one of the more poorly constrained parameters [*Koentopp et al.*, 2005; *Geiger and Drinkwater*, 2005] and is one that the system is relatively sensitive to: increasing it from 2 to 3% increases the average ice production to 2.5 m. Reducing the accumulation rate by a third from 150 to 100 mm a⁻¹ increases the average ice production by only 3%; decreasing the previous summer shore lead by 50% from 20% of the shelf area to 10% has little effect, reducing the average ice production by 0.04 m. We use an ice speed of 0.05 m s⁻¹ from Figure 9; increasing the ice speed by 20% yields an increase in average ice formation of 10% to 2.47 m. The ice velocity vectors in Figure 9 show the ice moving at an angle to the shelf break of around 45°, which is what we assume in our simulation. Increasing the angle, thereby shortening the average path across the continental shelf by 10%, say, increases the average ice production by 6%.

[42] This crude calculation of the heat and sea ice budgets, which follows a simple analysis originally undertaken by *Gill* [1973], uses the underlying premise that all the ice over the southwestern continental shelf is formed locally. If that assumption is largely correct, then the model highlights some of the sensitivities in the formation of HSSW and gives an estimate of the ice production rate over the continental shelf that is consistent with our estimate of the flux of HSSW that is formed.

3.3. Role of the Shore Lead

[43] The shore lead, more accurately termed an ice front polynya, is an area of open water adjacent to the ice front. It is maintained through the freezing season by the offshore component of the prevailing winds and by the opening and closing effect caused by the component of tidal motion perpendicular to the ice front [*Foldvik et al.*, 2001]. Antarctic ice front polynyas have long been recognized as important routes via which the atmosphere can extract heat from the ocean [e.g., *Bromwich and Kurtz*, 1984]. For the Ronne Ice Front *Grumbine* [1991] modeled the process using highly simplified topography, and *Markus et al.* [1998] used a combination of satellite and other data to determine how the water column would be salinized as it progressed around the coast and ice fronts, starting at the EWIS and ending up at the Antarctic Peninsula. *Foldvik et al.* [2001] carried out an analysis to show the effectiveness of tidal motion and offshore winds at maintaining an opening for heat loss.

[44] *Renfrew et al.* [2002] found an average wintertime ice production in the shore lead of 1.11×10^{11} m³, or an average of 1.48×10^5 m³ km⁻¹ along a 750-km shore lead. To be consistent with our model the figure needs to be reduced by about 4% (to 1.42×10^5 m³ km⁻¹) to account for the oceanic heat flux. The wintertime heat flux necessary to cool the MWDW and also to remove the heat gained during the summer (reported by *Renfrew et al.* [2002]) is 14 W m⁻², about 4% of *Renfrew et al.*'s [2002] average heat flux from open water of 320 W m⁻². Figure 10b shows the modeled ice thickness distribution near the ice front. The cumulative total ice thickness is also shown, that is, the total ice produced during the freezing season per meter of shore lead, integrated

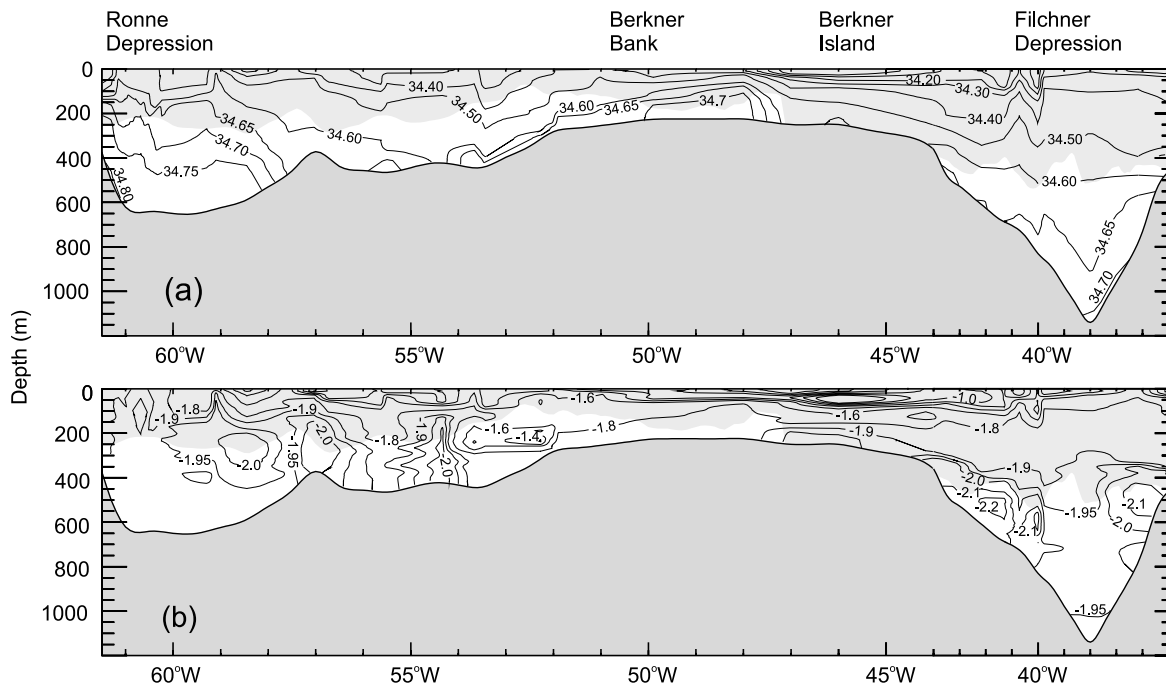


Figure 11. (a) Salinity and (b) potential temperature sections along the Filchner and Ronne ice fronts obtained in February 1993 by the Nordic Antarctic Research Programme 1992–1993 [Gammelsrød *et al.*, 1994]. The light shading indicates the draft of the ice shelf at the ice front.

from the ice front out across the continental shelf. This indicates that the *Renfrew et al.* [2002] shore lead ice production is achieved at a distance of about 13 km, consistent with *Renfrew et al.*'s [2002] average shore lead breadth of 10 km. In our model, therefore, the shore lead corresponds to ice thickness below around 20 cm (Figure 10b).

[45] For the period between 1992 and 1998, *Renfrew et al.* [2002] calculated the average heat loss from the shore lead during the freezing season to be 3.5×10^{19} J. Considering the size of the shore lead, which averaged 10 km in width during the freezing season, this is indeed an intense loss of heat, but it is only a small fraction of the 3×10^{20} J required to generate the necessary quantity of HSSW. Although *Renfrew et al.* [2002] did not include latent heat polynyas known to appear regularly during the winter along the coast between the Filchner Ice Front and the Brunt Ice Shelf (the Luitpold Coast, see Figure 2a), it is clear that the majority of the necessary ice production must take place over the remainder of the continental shelf.

[46] During the freezing season, the shore lead clearly acts as a region of intense salinization. This production could be viewed as giving the salinity a final boost before it passes into the sub-ice shelf cavity. The process of gaining access to that cavity is discussed section 4, but it seems not to be a steady flux: a strongly seasonal signal is seen both from ice front moorings and from instruments moored beneath the ice shelf. Thus, the intense sea ice production in the shore lead might act as a pump, supplying pulses of salt and helping to drive the HSSW into the cavity.

[47] A region of intense sea ice production over continental shelves, such as in an ice front shore lead, creates a pool of

high-salinity water with strong horizontal density gradients at its perimeter. The rim current set up in response to those gradients can become baroclinically unstable and shed eddies that exchange water across the boundary [Jones and Marshall, 1993; Gawarkiewicz and Chapman, 1995; Visbeck *et al.*, 1996]. Those eddies are then responsible for exporting the HSSW from the region of formation.

[48] However, according to our foregoing discussion about the amount of heat that needs to be extracted from the continental shelf waters to make up the required HSSW production rate, the majority of HSSW production must take place over the sea ice-covered region, requiring an average heat loss per unit area of around 16% of that in the shore lead, or around 50 W m^{-2} . The scenario in which convection takes place over a large area but with a smaller zone of more intense heat loss has not been covered by numerical experiments to date, and the shelf circulation and distribution of salinity that would result is unclear. The well-defined and sharp boundary between strong HSSW formation in the shore lead and no HSSW formation beneath the ice shelf would, however, remain.

3.4. Distribution of HSSW Salinity

[49] Several cruises along the Filchner-Ronne ice front have shown a general increase in salinity of the waters from east to west to be a robust feature of the oceanographic regime [e.g., Foldvik *et al.*, 1985b]. There is a local maximum in salinity over the relatively shallow Berkner Bank (Figure 11), with a reduction toward the west where the persistent core of MWDW reaches the ice front at 52–53°W. From the MWDW core to the western end of the ice front, the

increase in salinity is pronounced, with a maximum value of around 34.84 over the Ronne Depression, near the Antarctic Peninsula. There is no existing east to west CTD section from farther north, although *Nicholls et al.* [2003] show four shelf break to ice front sections, from three different years, which span the continental shelf seaward of the Ronne Ice Shelf. These indicate a similar pattern of increasing salinity from east to west.

[50] *Markus et al.* [1998] assume a westward flow of water around the coast, the salinity of the water gradually increasing as a result of sea ice production causing the conversion of ESW to HSSW, primarily within coastal and ice front polynyas. Their study confirmed that the salinity ultimately attained depended chiefly on the residence time of the water in the region of sea ice formation, thus providing one possible explanation of the overall pattern of salinity increasing westward. If, as we have suggested, MWDW, rather than ESW, is the main source of water over the southern continental shelf, then the cause of the observed salinity distribution must be reassessed.

[51] The data from the tagged Weddell seals reported by *Nicholls et al.* [2008] add significantly to our knowledge of the salinity distribution over the wintertime continental shelf, the principal result being the full-depth, low-salinity incursion from shelf break to ice front associated with the ice front MWDW core at 52–53°W. Thus, the principal inflow of MWDW occurs at the shelf break at around 43°W, near a known and significant depression in the bathymetry (marked by a heavy black broken arrow at the shelf break in Figure 2b). The ice front current meter mooring discussed in section 2 [*Foldvik et al.*, 2001] showed that the MWDW maintains its properties through most of the freezing season, finally being replaced by HSSW in late September. The water at the site reverted to MWDW in early December.

[52] Although the depression near 43°W might be the main focus of the flow onto the shelf, MWDW has been found whenever observations have been made within 100 km of the shelf break from between the western Berkner Bank and the Ronne Depression. The Ronne Depression itself, however, appears always to be dominated by HSSW, implying that the MWDW has had ample opportunity to be largely salinized by the time the water arrives there and suggesting a cyclonic (clockwise) circulation of water west of the MWDW inflow.

[53] The explanation that emerges for the observed salinity distribution is quite simple. The salinity over the Filchner Depression is kept low by a combination of the inflowing ESW from the east, the inflowing core of MWDW along the eastern flank of the depression, the outflowing ISW from beneath the Filchner Ice Shelf, and the comparative depth of the depression itself. Higher salinity over Berkner Bank results from sea ice production over a thinner water column. The highest salinity is found at the western end of the continental shelf, where the topography of the Antarctic Peninsula directs barrier winds northward [*Schwerdtfeger*, 1975]. The consistency in the direction of the barrier winds results in the sea ice at the western side of the continental shelf remaining thinner and the ice production being enhanced. The salinity in the central region of the Ronne

Ice Front is reduced by the presence of the inflowing MWDW.

3.5. Summary

[54] The principal conclusions to be drawn from section 3 are as follows. Although the shore lead is an intense sink of heat from the continental shelf, its small area means that it is responsible for only a small fraction of the heat loss needed to explain the conversion of source waters into HSSW in the volumes that we estimate to be necessary. A simple model of ice formation over the southern Weddell Sea continental shelf suggests that the conversion of 2.8 Sv of MWDW to HSSW can be accomplished, the primary assumption of the model being that the majority of the winter sea ice cover is locally generated. The observed salinity distribution across the continental shelf can be simply explained by the different water masses flowing onto the continental shelf from beneath the ice shelf and from across the shelf break and by the uniformity of the direction of the barrier wind on the western side of the continental shelf [*Schwerdtfeger*, 1975].

4. FLOW OF HSSW INTO THE CAVITY

[55] HSSW is produced during the freezing season, when the water column would be expected to be vertically homogeneous as a result of convective mixing. The Taylor-Proudman theorem suggests that water deeper than the ice shelf draft would not be able to move beneath the ice shelf farther than the distance traveled during an inertial half period. To conserve potential vorticity, the water should follow contours of f/h , where f is the Coriolis parameter and h is the water column height. The ice front provides a step in the water column thickness equal to the draft of the ice shelf. Along the FRIS ice front the draft is generally between 200 and 350 m in a water depth between 250 and 700 m. This represents a large topographic barrier to any wintertime flow into the sub-ice shelf cavity.

[56] There have been only four successful long-term current meter moorings deployed at the Ronne Ice Front and none at the Filchner Ice Front. The moorings show a predominantly along-ice front flow, from east to west, with a seasonally varying component of flow into the cavity at some locations [*Nicholls et al.*, 2003]. Measurements from beneath the ice shelf have not only shown that there is a substantial flow into the cavity but have also confirmed the strong seasonality in the flow [*Nicholls*, 1996].

[57] In an attempt to understand how HSSW enters the cavity at the western end of the ice front, *Jenkins et al.* [2004] applied an isopycnic coordinate ocean model to the southern Weddell Sea. The model, which was originally based on MICOM, had been modified to include sub-ice shelf domains [*Holland and Jenkins*, 2001]. For this model run its only forcing was a relaxation to a seasonally varying surface salinity. The modeled flow into the cavity occurred at the western end of the ice front twice each year: once in late winter and again in late summer. The late summer inflow has been observed using instruments moored at the western end of the ice front [*Nicholls et al.*, 2003]. As the model places the

wintertime inflow in the central Ronne Depression, the moored instruments, which were located on either side of the depression, did not sample the full strength of such an inflow, should it exist. However, a late winter inflow was detected during a year of particularly heavy sea ice [Makinson and Schröder, 2004].

[58] The late summer inflow is a result of the decoupling effect of the summertime mixed layer descending beyond the depth of the ice front. The wintertime inflow results from the generally westward flowing current along the ice front meeting the Antarctic Peninsula coast but having its northward escape blocked by a pool of dense HSSW. That flow is then forced southward beneath the ice shelf. How the vorticity budget is balanced remains an open question.

[59] Grosfeld et al. [1997] proposed a different mechanism for ventilation of the cavity. They pointed out that where bedrock depressions crossed the ice front, the interruption in the f/h contours was less severe. They presented results from a numerical model using idealized topography that showed water crossing the ice fronts, making use of the reduced step in the f/h contours over the bedrock slope. However, Nicholls et al. [2003] have shown that HSSW is in plentiful supply even during the summer, yet the flow into the cavity appears to arrive in pulses, suggesting that the cross-ice front depressions do not supply corridors through the f/h barrier.

[60] Tidal activity offers important additional mechanisms that can introduce HSSW into the sub-ice shelf cavity. Makinson and Nicholls [1999] use a modeling study to show that tidal residual flow, also known as tidal rectification, has the capacity to flush the cavity at a rate of ~ 0.35 Sv. Another tidal mechanism is based on the fact that the tidal excursion at the ice front is principally perpendicular to the ice barrier and therefore transports HSSW a few kilometers beneath the ice shelf every tidal cycle before returning the water to the shore lead. East of the Ronne Depression, there is a region of the sub-ice shelf cavity with a relatively narrow water column. This is a zone in which tidal currents are thought to exceed 1 m s^{-1} , with high levels of energy available for mixing [Makinson and Nicholls, 1999]. It is therefore likely that HSSW is able to exchange some of its properties with the relatively cold and fresh water beneath the ice shelf, effecting a degree of flushing even during the freezing season when the water column north of the ice front is well mixed.

5. PROCESSES WITHIN THE CAVITY

[61] A property of water that is the key to most of the ice-ocean interactions beneath the ice shelf is the depression of the freezing point with increasing pressure: for every kilometer of water depth, the freezing point is depressed by 0.75°C [Fofonoff and Millard, 1983]. Water that interacts with the base of a deep ice shelf can therefore attain temperatures below the freezing point at the sea surface, enabling HSSW to melt basal ice. If the water produced by such interactions has a temperature below the surface freezing point (when raised adiabatically to the surface), then that water mass is defined to be ISW.

[62] Once within the sub-ice shelf cavity the only influences on the HSSW are tidal forcing, the direct interaction with the ice shelf base, and the production of ice crystals within the water column when it becomes in situ supercooled. The direct interactions with the ice shelf are conductive heat loss into the ice and melting and freezing at the ice-ocean interface. The process of direct freezing at the interface (the formation of congelation ice, as contrasted with the deposition of ice crystals formed within the water column) is thought to be of little significance as the newly frozen ice rapidly creates an insulating layer that hinders further heat loss from the water column.

5.1. Basal Melting

[63] The rate of melting at the ice base depends on how fast heat and salt can be transported across the boundary layer to the ice-ocean interface. This, in turn, depends on the amount of turbulence present in the water column and the temperature and salinity gradient across the boundary layer. The turbulence largely depends on the roughness of the ice base and the water speed, though in the case of strong melting it is possible that the induced stratification could play a role in suppressing turbulent vertical fluxes. The usual name for the temperature difference between the mixed layer and the ice base is “thermal driving.” Away from the vicinity of the ice front, the warmest water in the sub-ice shelf cavity is at the surface freezing point of seawater ($\sim -1.9^\circ\text{C}$), which therefore places an upper limit on the mixed layer temperature. Similarly, the highest salinity found beneath the ice shelf is that of HSSW at around 34.8. The temperature of the ice-ocean interface is related to the ice draft and interfacial salinity via the pressure-dependent freezing point formula, and the interfacial salinity itself depends on the basal melt rate. Heat loss into the ice shelf reduces the basal melt rate, though this is not a large effect: the heat required to raise the ice to the freezing point is small compared to the heat needed to melt it.

[64] Holland and Jenkins [1999] discuss different formulations for the heat and salt transfer rate across the boundary layer. A solution that is thought to be reliable for FRIS is quadratic in temperature and very nearly linear in water speed. This is a “three-equation formulation” that is derived from the equations for heat and salt conservation and the formula relating freezing point, pressure, and salinity. It uses bulk transport coefficients for the diffusion of heat and salt [Kader and Yaglom, 1972], which depend on the assumption of a neutral boundary layer, so the melting at the ice base does not cause boundary layer stratification that impedes the vertical transport of heat and salt. It is therefore not appropriate for warmer continental shelf environments in which basal melt rates are likely to be relatively high ($>5 \text{ m a}^{-1}$, for example).

[65] The efficacy with which tidal activity is able to raise heat through the water column appears to be limited in all but the regions with the thinnest water column. A one-dimensional vertical-mixing model [Makinson, 2002] showed that only in the region west of Berkner Island was tidal activity capable of generating basal melt rates of meters

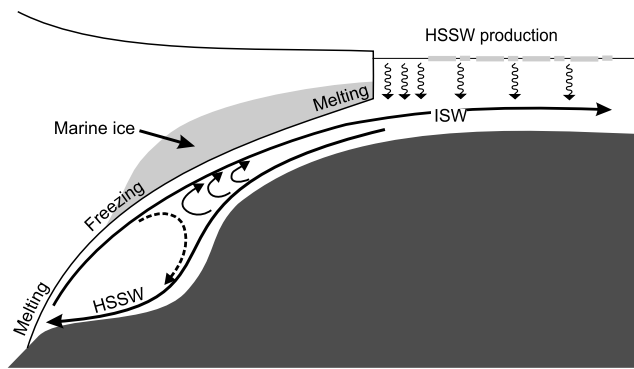


Figure 12. Two-dimensional schematic of the melting and freezing processes beneath FRIS.

per year. Field experiments have confirmed high melt rates in this area [Grosfeld and Blindow, 1993; Grosfeld et al., 1994]. Elsewhere, the rate supported by locally induced mixing, including contributions from shear and internal wave breaking within regions of strong vertical density gradients, was only a few centimeters per year, an order of magnitude lower than measured melt rates.

[66] The indirect dependence of melt rates on ice shelf draft means that the deep grounding lines where the ice sheet originally goes afloat are susceptible to relatively high melt rates. There the ice base can be as much as 1800 m below sea level, with an in situ freezing point more than 1.3°C below that at surface pressure. Basal melt rates near the grounding line of the Rutford Ice Stream, for example, have been found to be as high as 5 m a⁻¹ [Jenkins et al., 2006; Jenkins and Doake, 1991], with even higher values from satellite-derived estimates [Rignot and Jacobs, 2002].

[67] Average melt rates for the entire ice shelf have been estimated by comparing concentrations of oceanographic tracers in water entering the ice shelf cavity with concentrations measured leaving the cavity. Most useful among the tracers have been temperature [e.g., Nicholls et al., 2003] and oxygen isotopes [e.g., Weiss et al., 1979]. By noting the reduction in concentration of the heavier O¹⁸ isotope that results from the melting into the ocean of the strongly O¹⁸-depleted glacial ice and the associated reduction in salinity, it is possible to estimate the total meltwater input. All of these methods have problems, however, and the estimates for the average basal melt rate vary from around 0.20 to 0.34 m a⁻¹.

[68] An estimate of basal mass balance for the entire ice shelf has been obtained in a study that used a combination of ice thickness and snow accumulation data and ice velocities from interferometric synthetic aperture radar [Joughin and Padman, 2003]. Such estimates rely on the assumption that the ice shelf is in steady state, that is, that the distribution of ice thickness is not changing. The pattern of melting and freezing shown by Joughin and Padman [2003] is in good agreement with the pattern determined by other techniques, and their overall assessment of the average net basal melt rate was 0.20 ± 0.06 m a⁻¹.

5.2. Formation of Marine Ice

[69] Although direct freezing of congelation ice onto the ice shelf base does not play a major role in FRIS, this is not the case for marine ice that is formed by the buildup of frazil ice crystals precipitating out of the water column. Ice of marine rather than meteoric origin composes a substantial fraction of the ice shelf [Thyssen et al., 1993]. Figure 12 shows a simplified 2-D diagram of the process. ISW forms when basal melting occurring at great depths is relatively buoyant as a result of the added meltwater. As the ISW flows up the base of the ice shelf and the pressure reduces, the in situ freezing point increases. While enough warmer water is entrained into the ISW from below to maintain it above the freezing point, it is able to continue melting the ice shelf base. Once the pressure falls enough or the rate of entrainment of heat slows sufficiently, the in situ freezing point rises above the ISW temperature, which then becomes supercooled, and ice crystals are able to form. Eventually, the ice precipitates onto the ice shelf base. The building up of the “inverted snowpack” ultimately results in a layer of marine ice that can be hundreds of meters thick [Oerter et al., 1992; Lambrecht et al., 2007].

[70] Marine ice has interesting properties. As it has no air bubbles, it is clear ice and is consequently slightly denser than even deep meteoric ice [Oerter et al., 1992]. Unlike sea ice, marine ice has low salt concentrations: near the interface with the meteoric ice the bulk salt concentration peaks at about 0.1‰, while the concentration for the majority of the marine ice column is around 0.03–0.05‰ [Moore et al., 1992]. The mechanism for the highly efficient expulsion of salt from marine ice is not yet fully understood; it is thought that compaction and compositional convection are probably the important processes [Tabraham, 1997].

[71] Whichever way the desalination process is accomplished, the effect on the underlying seawater of the production and deposition of ice is the same: a reduction in buoyancy resulting from a combination of the increase in the density of the liquid fraction by the expulsion of salt from the individual ice crystals and the resultant increase in bulk density from the loss of the ice crystals when they deposit at the ice shelf base [Holland and Feltham, 2005]. In some situations the buoyancy loss coupled with the topography of the cavity constrains the outward flow of the ISW plume and causes it to recirculate toward the grounding line. In some model simulations this causes the formation of gyres driven purely by the difference in freezing point between deeper and shallower drafting parts of the ice shelf [Gerdes et al., 1999].

5.3. Effect on Water Properties of Interaction With the Ice Shelf

[72] The impacts of interactions with the ice shelf base on the properties of the inflowing HSSW are most easily seen on a plot of potential temperature (θ) against salinity (S). When seawater above its in situ freezing point comes into contact with the ice shelf base, the mass of ice that is melted, and, therefore, the freshening of the water column, depends principally on the thermal driving. The result is that the water mass properties execute a very nearly straight line trajectory

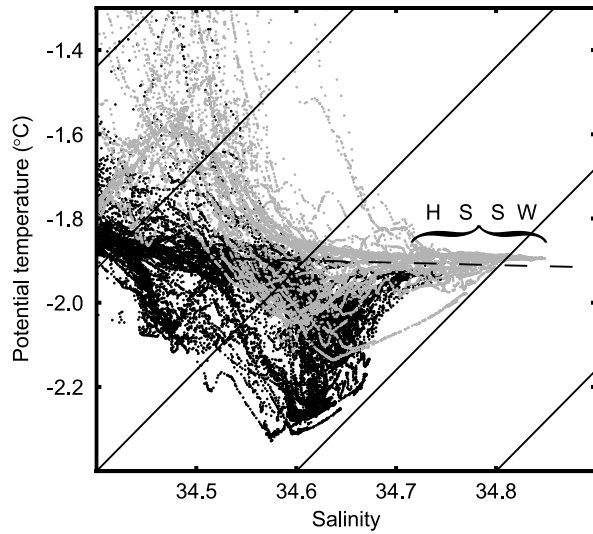


Figure 13. A θ - S plot illustrating the use of a meltwater mixing line to indicate the HSSW source area derived from ISW properties. When HSSW melts ice from the base of the ice shelf, its θ - S properties evolve along a line parallel to the straight meltwater mixing lines shown.

in θ - S space passing through (S_0, θ_0) , the salinity and potential temperature of the original seawater, with a gradient given by [Gade, 1979]

$$\frac{d\theta}{dS} = \frac{L}{S_0 c_w} + \frac{(\theta_f - T_i) c_i}{S_0 c_w} + \frac{(\theta_0 - \theta_f)}{S_0}, \quad (1)$$

where θ_f is the potential temperature at which ice melts at the ice shelf base; T_i is the temperature of the core of the ice shelf; c_i and c_w are the specific heat capacities of ice and water (around 2010 and 4000 J kg⁻¹ °C⁻¹), respectively; and L is the latent heat of ice (3.35×10^5 J kg⁻¹). At ~ 2.4 °C, the first term dominates. The second term, which applies only for melting regions, approximates the effect of the heat lost by diffusion into the ice itself; the third term arises from the need to cool the seawater to the freezing point. For values of T_i appropriate to FRIS (~ -25 °C), the second (ice warming) term is ~ 1 order of magnitude lower than the first but makes a measurable contribution to the θ - S gradient, increasing it to ~ 2.8 °C. The third (water cooling) term is smaller by another order of magnitude and is not significant. When ISW ascends sufficiently for ice to form within the water column, the θ - S properties of the water traverse back along the trajectory toward warmer and more saline conditions. In this case the second term is 0, as is the third term, in the absence of any in situ supercooling.

[73] The tightly constrained evolution of seawater θ - S properties when interacting with the base of an ice shelf, and the observation that the warmest water beneath the ice shelf is at the surface freezing point, means that it is, in principle, possible to determine the salinity of the source water of any waters observed beneath the ice shelf, given their temperature and salinity [Nøst and Foldvik, 1994]. We merely need to intersect the mixing line whose gradient is

given by (1) with the surface freezing point line in the θ - S diagram. As the salinity of the water along the Ronne Ice Front varies in a known way, it is then possible, in principle, to determine the point at which the source water originally entered the cavity. The method clearly breaks down if sources with different salinities enter the cavity and mix together or if ISW exits the cavity, is warmed indirectly by the atmosphere, and then reenters the cavity.

5.4. Circulation Within the Sub-Ice Shelf Cavity

[74] The gradient in salinity from east to west along the ice front makes the FRIS cavity well suited to this type of analysis. The θ - S diagram in Figure 13 illustrates the method. The data are from the vicinity of the Filchner Ice Front (black dots) over plotted by data from near the Ronne Ice Front (gray dots). The coldest of the ISW from the Filchner Ice Front clearly originates from HSSW with salinity in excess of 34.75, which places the source at the western end of the Ronne Ice Front. We note that the most saline HSSW observed at the western end of the Ronne Ice Front ($S > 34.80$) seems to have no related ISW emerging from beneath the Filchner Ice Shelf.

[75] Foldvik et al. [2001] describe the large-scale circulation beneath the ice shelf. Broadly, HSSW entering via the Ronne Ice Front circulates south of Berkner Island to emerge as ISW from beneath the Filchner Ice Front. Ship-based observations from along the ice fronts have revealed that much. Access to the sub-ice shelf environment via boreholes drilled through the ice shelf have allowed sequences of CTD profiles to be obtained together with time series from instruments left suspended beneath the ice shelf. The locations of the boreholes are given in Figure 2b. These data sets have been used to elaborate the picture of the circulation within the cavity [Nicholls et al., 2001].

[76] The densest HSSW ($S > \sim 34.75$) enters via the Ronne Depression (Figure 2b), adjacent to the Antarctic Peninsula. Some of this inflow flushes the Ronne Depression beneath the ice shelf, presumably as far south as the deep grounding lines. Measurements made by Nicholls et al. [1997] west of the Korff Ice Rise showed that the deeper, warmer water was below the depth of the elevated bedrock south of the Korff and Henry ice rises, which they assumed acts as a barrier to its eastward flow. Another arm of the Ronne Depression HSSW inflow follows the 600-m bedrock contour through a gap in the ridge that forms the eastern boundary of the depression. This flow reaches Berkner Island's western coast before turning south to round its southern tip. The water, now ISW with a temperature of -2.3 °C, emerges at the Filchner Ice Front at a depth of around 600 m. Too dense to escape the Filchner Depression, this water returns beneath the ice shelf along the eastern margin of the Filchner Depression to undertake a long journey south to the mouth of the Foundation Ice Stream [Nicholls et al., 2001]. More melting of the deep ice in this area drives the ISW up the ice shelf base, along the eastern coast of the Henry Ice Rise, where much ice deposition occurs. The majority of this ISW is thought to join the water flowing southward along the Berkner coast, thus forming an elongate gyre. The way in which this dense ISW

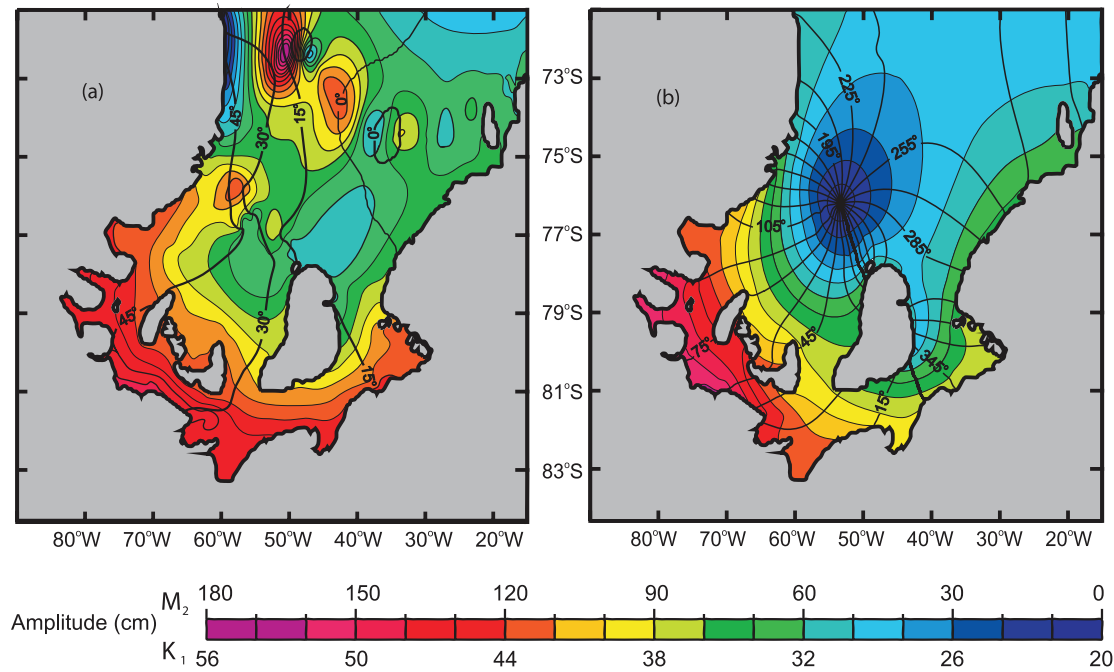


Figure 14. (a) Diurnal (K_1) tidal phase and amplitude. (b) Semidiurnal (M_2) tidal phase and amplitude [Makinson and Nicholls, 1999].

within the Filchner Depression is converted to a form capable of escaping the continental shelf, presumably at the Filchner Sill, is not clear.

[77] Inflow of a less saline version of HSSW occurs from the Berkner Bank on the eastern side of the Ronne Ice Front [Foldvik *et al.*, 2001]. This HSSW travels around Berkner Island, arriving at the Filchner Ice Front as ISW with a temperature of about -2.05°C and at a depth shallower than around 480 dbar. This version of ISW is able to escape the depression via the Filchner Sill and has indeed been observed at the sill [Nicholls *et al.*, 2001].

[78] Given so few boreholes, some of the details of the pattern must be regarded as speculative. Although the broad picture presented in this section has gained credibility from its consistency with some numerical simulations [Jenkins *et al.*, 2004], other model studies have suggested a different interpretation, one that places less emphasis on forcing from conditions north of the ice shelf [Gerdes *et al.*, 1999] and more emphasis on internal, energetic barotropic gyres that follow the contours of f/h . However, we believe that the observation of strong seasonal and interannual variability within the cavity (section 6), a variability that can only be forced by conditions north of the ice front, supports the configuration of currents shown in Figure 2b.

6. TEMPORAL VARIABILITY WITHIN THE CAVITY

[79] Instrument moorings deployed through boreholes on the Ronne Ice Shelf have shed light on the scales of variability within the cavity, from tidal and shorter time scales [Nicholls, 1996], through seasonal [Nicholls, 1996, 1997] to interannual variability [Nicholls and Østerhus, 2004].

6.1. Tidal Forcing

[80] Tidal heights over FRIS have been measured using tiltmeters near grounding lines [Smith, 1991], gravimeters, and, more recently, geodetic GPS receivers and satellite altimetry [Fricker and Padman, 2002]. They show a principally semidiurnal tidal variation, with a Kelvin wave propagating around the FRIS grounding line. The diurnal tidal wave propagates east to west across the Weddell Sea, and cotidal lines for the K_1 tide, as simulated by a barotropic tidal model, are shown in Figure 14a [Makinson and Nicholls, 1999]. Cotidal lines for the M_2 tide are shown in Figure 14b. The semidiurnal tides are the principal contributors to a tidal range of almost 8 m in the western Ronne Ice Shelf, near the Rutford, Carlson, and Evans grounding lines. Tidal models suggest that the tidal current speeds are largely controlled by the water column height within the cavity and are not significantly amplified in the Rutford-Carlson-Evans area. They are, however, on the order of 1 m s^{-1} in the relatively narrow water column in the area west of Berkner Island, resulting in the vigorous vertical mixing and high basal melt rates discussed in section 5 [Grosfeld *et al.*, 1994; Joughin and Padman, 2003; Makinson and Nicholls, 1999].

6.2. Seasonal Variability

[81] Seasonal variation in water temperature has been observed using moorings deployed through drill holes [Nicholls, 1996]. That variation presumably causes seasonality in basal melt rates. However, near the Rutford grounding line, an area deep within the cavity (Figure 2b) where there are no direct measurements of subglacial oceanographic conditions, Jenkins *et al.* [2006] show that basal melt rates averaged over a few weeks during the summer are the same as those obtained from averaging over an entire year,

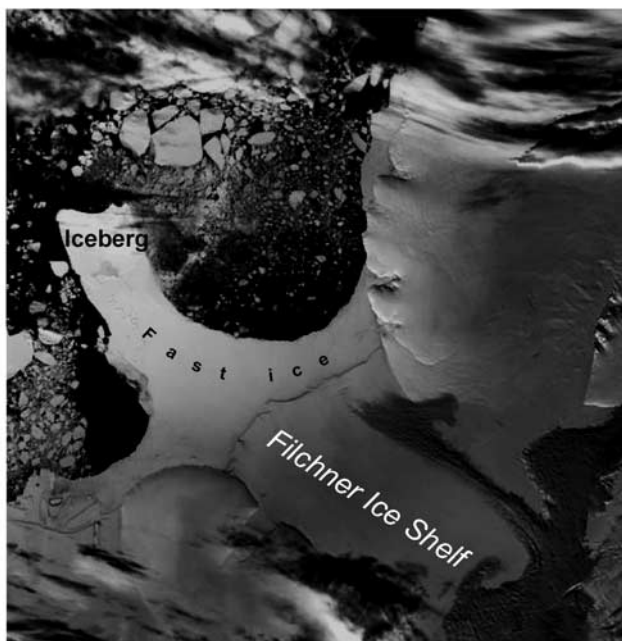


Figure 15. AVHRR visible image from 13 December 2000 showing fast ice between the Filchner Ice Front and grounded iceberg A-23 [Scambos et al., 2005].

suggesting an absence of a strong seasonal signal in the ocean conditions in that area.

[82] Nicholls [1997] suggested that the response of the cavity to seasonal changes in conditions north of the ice shelf might be a proxy for the cavity's response to a changing climate. The idea was that a warming in winter conditions would result in a reduction in the rate of formation of sea ice and therefore of HSSW, thus leading to a reduction in the flushing of the cavity and a consequent reduction in basal melting. More recent measurements discussed in section 4 have shown this argument to be too simplistic; the continental shelf is dominated by HSSW even at the end of summer [Nicholls et al., 2003], suggesting that it is the control on the flow of HSSW into the cavity rather than the HSSW production rate that determines the rate of flushing of the cavity. The arguments in section 4 indicate that although instruments moored at the ice front have shown an HSSW pulse into the cavity during a winter with a particularly high HSSW production, the summer stratification exerts an important control on the flushing of the cavity.

6.3. Interannual Variability

[83] Anomalous sea ice conditions during the austral summer of 1997–1998 resulted in a shore lead of unprecedented breadth at the front of the Ronne Ice Shelf [Hunke and Ackley, 2001]. Instruments deployed through the ice shelf in the summer of 1998–1999 captured the response of the cavity to what was presumed to be a particularly intense flux of HSSW as the shore lead froze over in the early winter [Renfrew et al., 2002]. The instruments were deployed off the southern coast of Berkner Island (Figure 2b) and recorded currents and temperatures for ~3 years. The interpretation of

the results are given in section 5 as a flow that recirculates within the Filchner Depression before it is presumably of sufficiently low density to allow it to escape over the Filchner Sill and descend the continental slope. This anomalous flushing event allowed an estimate of 4–5 years for the total flushing time for the cavity [Nicholls and Østerhus, 2004]. Instruments deployed at the drill sites west of the Korff Ice Rise (Figure 2b) showed no anomalous behavior during or just after 1998. This is consistent with Jenkins et al.'s [2004] model results; of the two inflow events into the sub-ice shelf cavity from the Ronne Depression that were suggested by their application of MICOM and were discussed in section 4, the late winter inflow flushed the western cavity (the sub-ice shelf extension of the Ronne Depression), while the late summer inflow supplied the water that flowed eastward toward the Filchner Depression. It is the second inflow that is thought to respond directly to the intensity of the winter HSSW production (section 4). Clearly, if the conditions seen during the 1997–1998 summer became the norm, with more southerly winds regularly producing large areas of open water, then we might expect a more vigorous flushing of the cavity, with higher overall melt rates.

[84] Other causes of interannual variations are changes in the shape of the cavity. Significant changes in the thickness of the ice shelf have not yet been observed, but major iceberg calving events occur every few decades in which tens of kilometers of ice shelf break out at once. Strong tidal mixing and the associated strong basal melting in the region near the ice front west of Berkner Island have already been noted. The loss of a few tens of kilometers of ice shelf from this region substantially reduces the area of ice shelf subject to that strong melting [Joughin and Padman, 2003].

[85] Calving is likely to have other consequences, too. The ice shelf thins rapidly toward the ice front, probably as a result of a combination of the unconstrained glacial flow and the effect of tidal action flushing warmer waters from the shore lead into the first few kilometers of the cavity [Jenkins and Doake, 1991]. Model studies [Jenkins et al., 2004] and the results from ice front moorings [Makinson and Schröder, 2004] have suggested that HSSW is able to flow into the cavity when the summer pycnocline reaches the depth of the ice shelf base at the ice front, thus decoupling the water column deeper than the ice shelf draft from the water above (see section 4). A calving event such as the one that occurred in 2000 has the effect of increasing the draft of the ice shelf at the ice front by up to 100 m or so, with potentially dramatic consequences for the timing, or even the existence, of an end of summer HSSW inflow.

[86] Another possible source of longer time scale variability is the stranding of icebergs, with consequent changes to the flow patterns over the open continental shelf to the north. In 1986 a major calving event from the Filchner Ice Front left three icebergs grounded on the Berkner Bank (Figure 2b). These giant ice islands have the effect of substantially changing the topography of the region, and their presence has, on several occasions, stabilized a large region of multi-year fast ice, extending from the icebergs to the Filchner Ice Front (Figure 15). Even in the absence of the fast ice, the

icebergs block the westward flow of sea ice, increasing the ice concentration over the Berkner Bank north of the Filchner Ice Shelf. Conversely, west of the icebergs the sea ice concentration is reduced, leading to higher ice formation rates and, therefore, HSSW production. Thus, the primary impact of the bergs is to reduce HSSW formation on the east side of the Berkner Bank and to increase it farther west. This change led to the cooling and freshening of the Filchner Depression as the inflow of relatively warm (-1.9°C) HSSW from the Berkner Bank reduced and the water column became dominated by ISW from beneath the Filchner Ice Shelf [Nøst and Østerhus, 1998].

7. FUTURE DIRECTIONS

[87] The ISW-rich water that overflows the Filchner Sill and makes its way to the abyss of the Weddell Sea is an important ingredient in Weddell Sea Bottom Water and, ultimately, in the Antarctic Bottom Water that has its origins in the Weddell Sea sector. The flux and properties of the plume are the result of a series of upstream processes; monitoring the plume over the long term would therefore supply a powerful indication of the temporal variability of those processes and of their contribution to the climate system as a whole. Source waters flowing onto the southwestern continental shelf appear to be a combination of a coastal current, observed flowing along the Luitpold Coast (Figure 2a), and incursions of MWDW directly across the shelf break. In the absence of year-round data from the coastal current, the relative contribution of these two sources is not yet clear and is the subject of continuing research.

[88] We have presented estimates suggesting that the majority of the transformation of source waters to HSSW takes place over the continental shelf north of the shore lead. This is a region that is difficult to access even during summer, and there are no ocean-atmosphere heat flux data from the important winter months. Techniques presently under development, however, offer some hope that it will soon be possible to infer heat loss through the winter ice pack using satellite-derived snow depth and ice thickness data products.

[89] Direct study of the flow of HSSW into the sub-ice shelf cavity using ice front moorings is difficult. The ice front moves outward during the period that moorings are deployed, meaning that the gradients associated with topographic effects due to the ice front move over the mooring location. Calving of icebergs also presents difficulties, and the passage of icebergs along the ice front is a substantial danger to the moorings themselves. The study of processes near ice fronts is presently the province of laboratory models and tidally enabled numerical models. However, we are now in a position to identify locations beneath the ice shelf where instrument moorings deployed through access holes and maintained over the long term would provide a reliable record of the flux of HSSW and its products through the sub-ice shelf system. Although creating access holes is logistically taxing and therefore expensive to undertake, in a small number of locations they have given direct evidence for the conditions beneath the ice shelf.

[90] Good progress has been made in improving our understanding of the circulation beneath the ice shelf through the application of numerical models [Williams *et al.*, 1998; Gerdes *et al.*, 1999; Holland *et al.*, 2007]. It is the belief of the authors, however, that numerical models have not yet reached the stage of being able to simulate with sufficient fidelity the key processes in the ice shelf ocean system to answer the central question of how the system will respond to future changes in climate. Models have proven to be useful tools in assisting in the interpretation of field data [e.g., Jenkins *et al.*, 2004].

[91] In the absence of a spatially more extensive data set from sub-ice shelf observations, measurements of the rate of basal melting can supply a useful diagnostic for numerical models. Satellite-derived data sets can be used to give an indirect estimate of basal melt rates, but a powerful, newly applied technique using phase-sensitive radio echo sounding (PRES) has proved itself capable of direct and precise melt rate measurements [Jenkins *et al.*, 2006]. The principal disadvantage of the PRES technique is that it cannot be used in regions where the ice column has a substantial marine ice component. It could, however, be used to validate the satellite methods and model results in areas wholly composed of glacial ice.

[92] There are two potent motivations for further developing our understanding of the interactions between ice shelves and the underlying ocean. The first is the need to predict the future state of ice shelves in order to determine the likely contribution of the Antarctic Ice Sheet to sea level change; changes in ice shelves affect the restraint they exert on the seaward flow of the inland ice sheet [Scambos *et al.*, 2004; Dupont and Alley, 2005]. The second is the need to predict the impact of the ice shelves on the ocean. Ocean cavities beneath ice shelves cover about 40% of the area of the Antarctic continental shelf. Of particular importance is the interaction between FRIS and the underlying ocean; as a result of its influence on the characteristics of Antarctic Bottom Water originating from the Weddell Sea, the interaction between FRIS and the Southern Ocean makes FRIS an active component in the global climate system. However, the cavity beneath FRIS is possibly the least accessible part of the world ocean, and it will continue to stretch the ingenuity of researchers as they attempt to fathom its secrets.

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