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SUB ICE-SHELF CIRCULATION AND PROCESSES

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doi:10.1006/rwos.2001.0010

Introduction

Ice shelves are the floating extension of ice sheets (*see* **Ice-shelf Stability**). They extend from the grounding line, where the ice sheet first goes afloat, to the ice front, which usually takes the form of an ice cliff dropping down to the sea. Although there are several examples on the north coast of Greenland, the largest ice shelves are found in the Antarctic where they cover 40% of the continental shelf. Ice shelves can be up to 2 km thick and have horizontal extents of several hundreds of kilometers. The base of an ice shelf provides an intimate link between ocean and cryosphere. Three factors control the oceanographic regime beneath ice shelves: the geometry of the sub-ice shelf cavity, the oceanographic conditions beyond the ice front, and tidal activity. These factors combine with the thermodynamics of the interaction between sea water and the ice shelf base to yield various glaciological and oceanographic phenomena: intense basal melting near deep grounding lines and near ice fronts; deposition of ice crystals at the base of some ice shelves, resulting in the accretion of hundreds of meters of marine ice; production of sea water at temperatures below the surface freezing point, which may then contribute to the formation of Antarctic Bottom Water (*see* **Bottom Water Formation**); and the upwelling of relatively warm Circumpolar Deep Water.

Although the presence of the ice shelf itself makes measurement of the sub-ice shelf environment difficult, various field techniques have been used to study the processes and circulation within sub-ice shelf cavities. Rates of basal melting and freezing affect the flow of the ice and the nature of the ice–ocean interface, and so glaciological measurements can be used to infer the ice shelf's basal mass balance. Another indirect approach is to make ship-based oceanographic measurements along ice fronts. The properties of in-flowing and out-flowing water masses give clues as to the processes needed to transform the water masses. Direct measurements of oceanographic conditions beneath ice shelves have been made through natural access holes such as rifts, and via access holes created using thermal (mainly hot-water) drills. Numerical models of the sub-ice shelf regime have been developed to complement the field measurements. These range from simple one-dimensional models following a plume of water from the grounding line along the ice shelf base, to full three-dimensional models coupled with sea ice models, extending out to the continental shelf-break and beyond.

The close relationship between the geometry of the sub-ice shelf cavity and the interaction between the ice shelf and the ocean implies a strong dependence of the ice shelf/ocean system on the state of the ice sheet. During glacial climatic periods the geometry of ice shelves would have been radically different to their geometry today, and ice shelves probably played a different role in the climate system.

Geographical Setting

By far the majority of the world's ice shelves are found fringing the Antarctic coastline (**Figure 1**). Horizontal extents vary from a few tens to several hundreds of kilometers, and maximum thickness at the grounding line varies from a few tens of meters to 2 km. By area, the Ross Ice Shelf is the largest at around 500 000 km^2. The most massive, however, is the very much thicker Filchner-Ronne Ice Shelf in the southern Weddell Sea. Ice from the Antarctic Ice Sheet flows into ice shelves via fast-moving ice streams (**Figure 2**). As the ice moves seaward, further nourishment comes from snowfall, and, in some cases, from accretion of ice crystals at the ice shelf base. Ice is lost by melting at the ice shelf base and by calving of icebergs at the ice front. Current estimates suggest that basal melting is responsible for around 25% of the ice loss from Antarctic ice shelves; most of the remainder calves from the ice fronts as icebergs.

Over central Antarctica the weight of the ice sheet depresses the lithosphere such that the seafloor beneath many ice shelves deepens towards the grounding line. The effect of the lithospheric depression has probably been augmented during glacial periods by the scouring action of ice on the seafloor: at the glacial maxima the grounding line would have been much closer to the continental shelf-break. Since ice shelves become thinner towards the ice front and float freely in the ocean, a typical sub-ice shelf cavity has the shape of a cavern that dips downwards towards the grounding line (**Figure 2**). This geometry has important consequences for the ocean circulation within the cavity.

Oceanographic Setting

The oceanographic conditions over the Antarctic continental shelf depend on whether relatively warm, off-shelf water masses are able to cross the continental shelf-break.

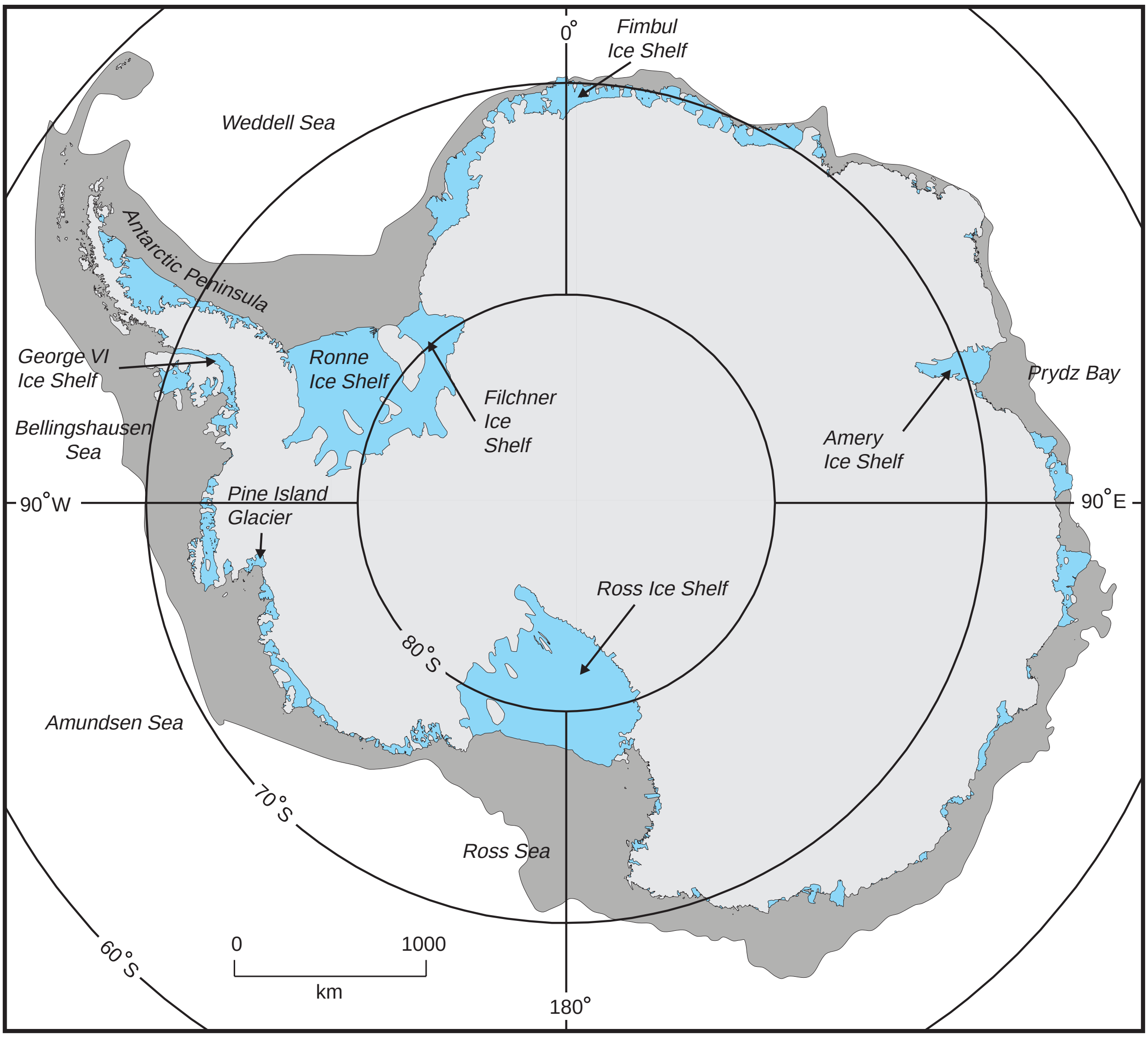

Figure 1 Map showing ice shelves (blue) covering about 40% of the continental shelf (dark gray) of Antarctica.

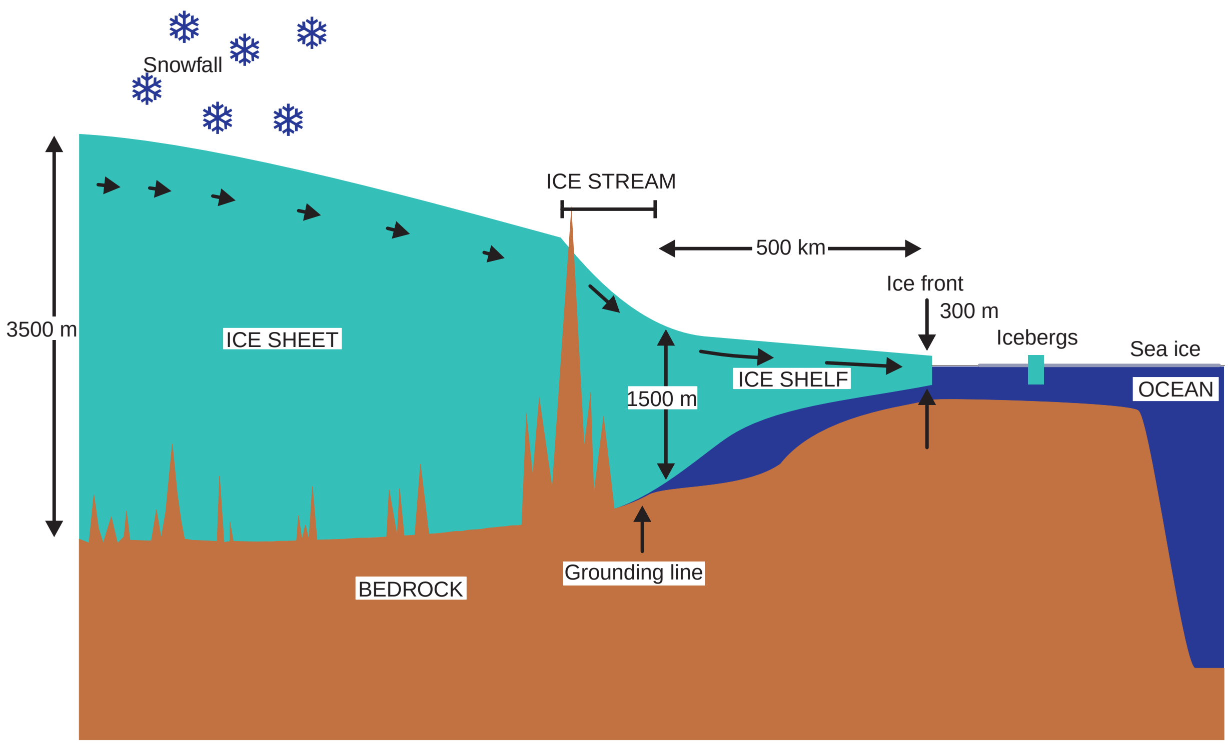

Figure 2 Schematic cross-section of the Antarctic ice sheet showing the transition from ice sheet to ice stream to ice shelf. Also shown is the depression of the lithosphere that results in the deepening of the seabed towards the continental interior.

For much of Antarctica a dynamic barrier at the shelf-break prevents advection of circumpolar deep water (CDW) onto the continental shelf itself. In these regions the principal process determining the oceanographic conditions is production of sea ice (*see* **Sea Ice**) in coastal polynyas (*see* **Polynyas**) and leads, and the water column is largely dominated by high salinity shelf water (HSSW). Long residence times over some of the broader continental shelves, for example in the Ross and southern Weddell seas, enable HSSW to attain salinities of over 34.8 PSU. HSSW has a temperature at or near the surface freezing point (about -1.9°C), and is the densest water mass in Antarctic waters. Conditions over the continental shelves of the Bellingshausen and Amundsen seas (**Figure 1**) represent the other extreme. There, the barrier at the shelf-break appears to be either weak or absent. At a temperature of about 1°C, CDW floods the continental shelf.

Between these two extremes there are regions of continental shelf where tongues of modified warm deep water (MWDW) are able to penetrate the shelf-break barrier (**Figure 3**), in some cases reaching as far as ice fronts. MWDW comes from above the warm core of CDW: the continental shelf effectively skims off the shallower and cooler part of the water column.

What Happens When Ice Shelves Melt into Sea Water?

The freezing point of fresh water is 0°C at atmospheric pressure. When the water contains dissolved salts, the freezing point is depressed: at a salinity of around 34.7 PSU the freezing point is -1.9°C. Sea water at a temperature above -1.9°C is therefore capable of melting ice. The freezing point of water is also pressure dependent. Unlike most materials, the pressure dependence for water is negative: increasing the pressure decreases the freezing point. The freezing point T_f of sea water is approximated by:

$$T_f = aS + bS^{3/2} - cS^2 - dp$$

where $a = -5.75 \times 10^{-2}$°C PSU^{-1}, $b = 1.710523 \times 10^{-3}$°C PSU$^{-3/2}$, $c = -2.154996 \times 10^{-4}$°C PSU^{-2} and $d = -7.53 \times 10^{-4}$ °C dbar^{-1}. S is the salinity in PSU, and p is the pressure in dbar. Every decibar increase in pressure therefore depresses the freezing point by 0.75 m°C. The depression of the freezing point with pressure has important consequences for the interaction between ice shelves and the ocean. Even though HSSW is already at the surface freezing point, if it can be brought into contact with an ice

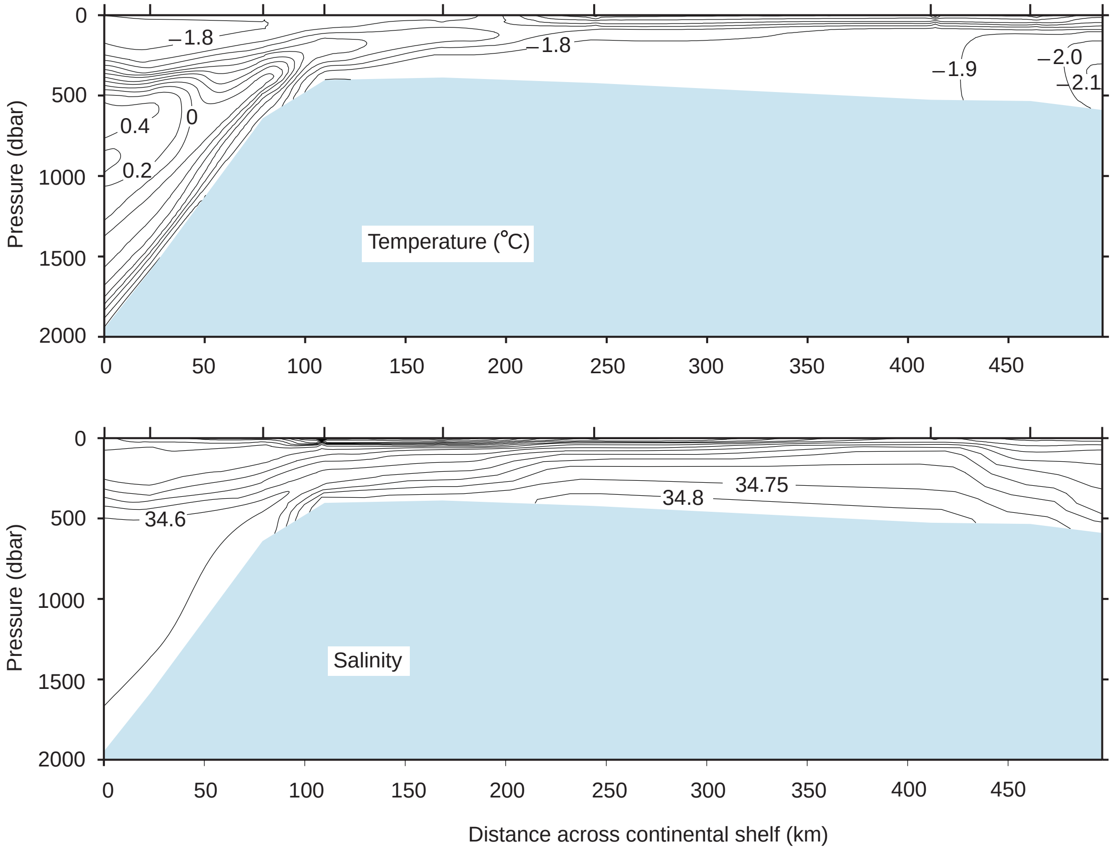

Figure 3 Hydrographic section over the continental slope and across the open continental shelf in the southern Weddell Sea, as far as the Ronne Ice Front. Water below the surface freezing point (-1.9°C) can be seen emerging from beneath the ice shelf. The majority of the continental shelf is dominated by HSSW, although in this location a tongue of warmer MWDW penetrates across the shelf-break. The station locations are shown by the heavy tick marks along the upper axes.

shelf base, melting will take place. As the freezing point at the base of deep ice shelves can be as much as 1.5°C lower than the surface freezing point, the melt rates can be high.

When ice melts into sea water the effect is to cool and freshen. Consider unit mass of water at temperature T_0, and salinity S_0 coming into contact with the base of an ice shelf where the *in situ* freezing point is T_f. The water first warms m kg of ice to the freezing point, and then supplies the latent heat necessary for melting. The resulting mixture of melt and sea water has temperature T and salinity S. If the initial temperature of the ice is T_i, the latent heat of melting is L, the specific heat capacity of sea water and ice, c_w and c_i, then heat and salt conservation requires that:

$$(T - T_f)(1 + m)c_w + m(c_i(T_f - T_i) + L) = (T_o - T_f)c_w$$

$$S(1 + m) = S_o$$

Eliminating m, and then expressing T as a function of S reveals the trajectory of the mixture in T–S space as a straight line passing through (S_0, T_0), with a gradient given by:

$$\frac{dT}{dS} = \frac{L}{S_o c_w} + \frac{(T_f - T_i)c_i}{S_o c_w} + \frac{(T_o - T_f)}{S_o}$$

The gradient is dominated by the first term, which evaluates to about 2.4°C PSU^{-1}. In polar waters the third term is two orders of magnitude lower than the first; the second term results from the heat needed to warm the ice, and, at about a tenth the size of the first term, makes a measurable contribution to the gradient. This relationship allows the source water for sub-ice shelf processes to be found by inspection of the T–S properties of the resultant water masses. Examples of T–S plots from beneath ice shelves in warm and cold regimes are shown in **Figure 4**.

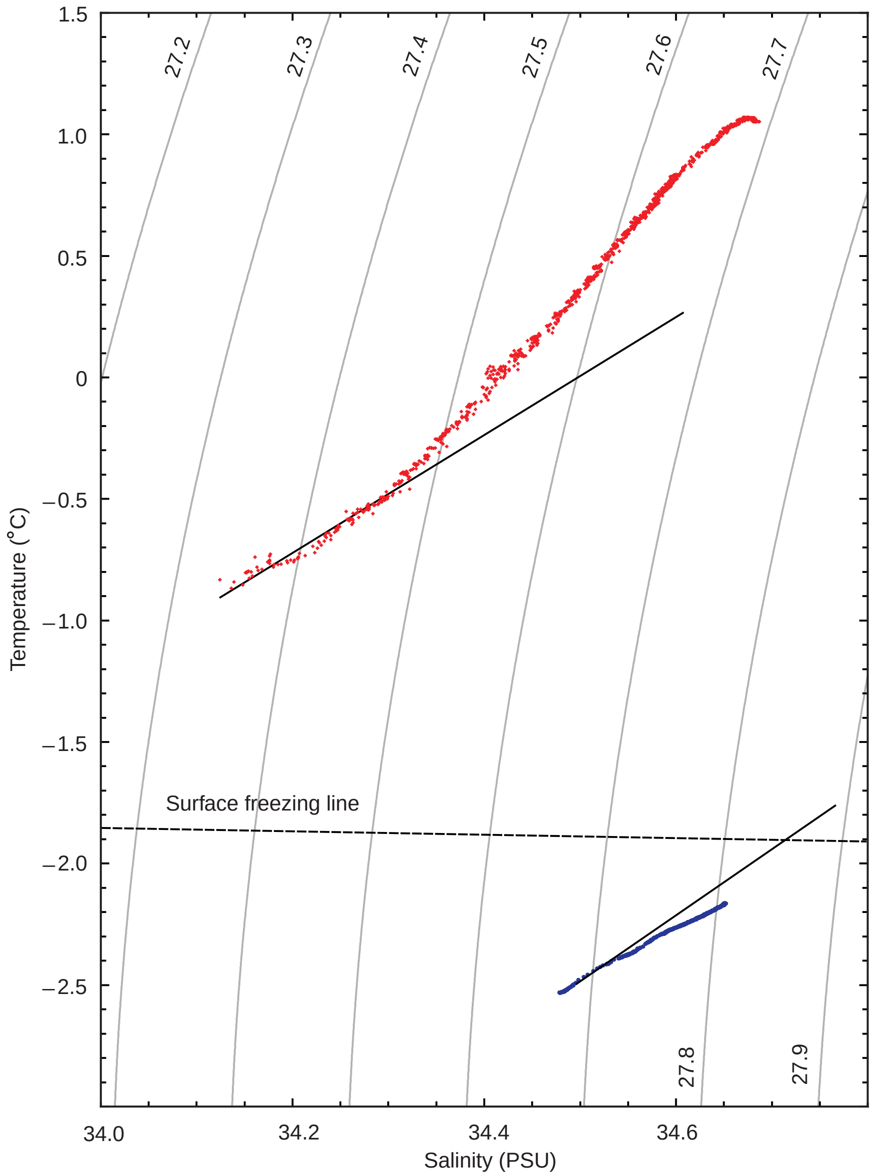

Figure 4 Temperature and salinity trajectories from CTD stations through the George VI Ice Shelf (red) and Ronne Ice Shelf (blue). The cold end of each trajectory corresponds to the base of the ice shelf. The straight lines are at the characteristic gradient for ice melting into sea water. For the Ronne data, as the source water will be HSSW at the surface freezing point, the intersection of the characteristic with the broken line gives the temperature and salinity of the source water. The isopycnals (gray lines) are referenced to sea level.

Two important passive tracers are introduced into sea water when glacial ice melts. When water evaporates from the ocean, molecules containing the lighter isotope of oxygen, ^{16}O, evaporate preferentially. Compared with sea water the snow that makes up the ice shelves is therefore low in ^{18}O. By comparing the $^{18}O/^{16}O$ ratios of the outflowing and inflowing water it is possible to calculate the concentration of melt water, provided the ratio is known for the glacial ice. Helium contained in the air bubbles in the ice is also introduced into the sea water when the ice melts. As helium's solubility in water increases with increasing water pressure, the concentration of dissolved helium in the melt water can be an order of magnitude greater than in ambient sea water, which has equilibrated with the atmosphere at surface pressure.

Modes of Sub-ice Shelf Circulation

Various distinguishable modes of circulation appear to be possible within a sub-ice shelf cavity. Which mode is active depends primarily on the oceanographic forcing from seaward of the ice front, but also on the geometry of the sub-ice shelf cavity. Thermohaline forcing drives three modes of circulation, although the tidal activity is thought to play an important role by supplying energy for vertical mixing. Another mode results from tidal residual currents.

Thermohaline Modes

Cold regime external ventilation Over the parts of the Antarctic continental shelf dominated by the production of HSSW, such as in the southern Weddell Sea, the Ross Sea, and Prydz Bay, the circulation beneath large ice shelves is driven by the drainage of HSSW into the sub-ice shelf cavities. The schematic in **Figure 5** illustrates the circulation mode. HSSW drains down to the grounding line where tidal mixing brings it into contact with ice at depths of up to 2000 m. At such depths HSSW is up to 1.5°C warmer than the freezing point, and relatively rapid melting ensues (up to several meters of ice per year). The HSSW is cooled and diluted, converting it into ice shelf water (ISW), which is defined as water with a temperature below the surface freezing point.

ISW is relatively buoyant and rises up the inclined base of the ice shelf. As it loses depth the *in situ* freezing point rises also. If the ISW is not entraining sufficient HSSW, which is comparatively warm, the reduction in pressure will result in the water becoming *in situ* supercooled. Ice crystals are then able to form in the water column and possibly rise up and accrete at the base of the ice shelf. This 'snowfall' at the ice shelf base can build up hundreds of meters of what is termed 'marine ice'. Entrainment of HSSW, and the possible production of ice crystals, often result in the density of the ISW finally matching the ambient water density before the plume has reached the ice front. The plume then detaches from the ice shelf base, finally emerging at the ice front at mid-water depths.

The internal Rossby radius beneath ice shelves is typically only a few kilometers, and so rotational effects must be taken into account when considering the flow in three dimensions. HSSW flows beneath the ice shelf as a gravity current and is therefore gathered to the left (in the Southern Hemisphere) by the Coriolis force. As an organized flow, it then follows bathymetric contours. Once converted into ISW, the flow is again gathered to the left, following either the coast, or topography in the ice base. If the ISW plume fills the cavity, conservation of potential vorticity would demand that it follow contours of constant water column thickness. The step in water column thickness caused by the ice front then presents a topographic obstacle for the outflow of the ISW. However, the discontinuity can be reduced by the presence of trenches in the seafloor running across the ice front. This has been proposed as the mechanism that allows ISW to flow out from beneath the Filchner Ice Shelf, in the southern Weddell Sea (**Figure 1**).

Initial evidence for this mode of circulation came from ship-based oceanographic observations along

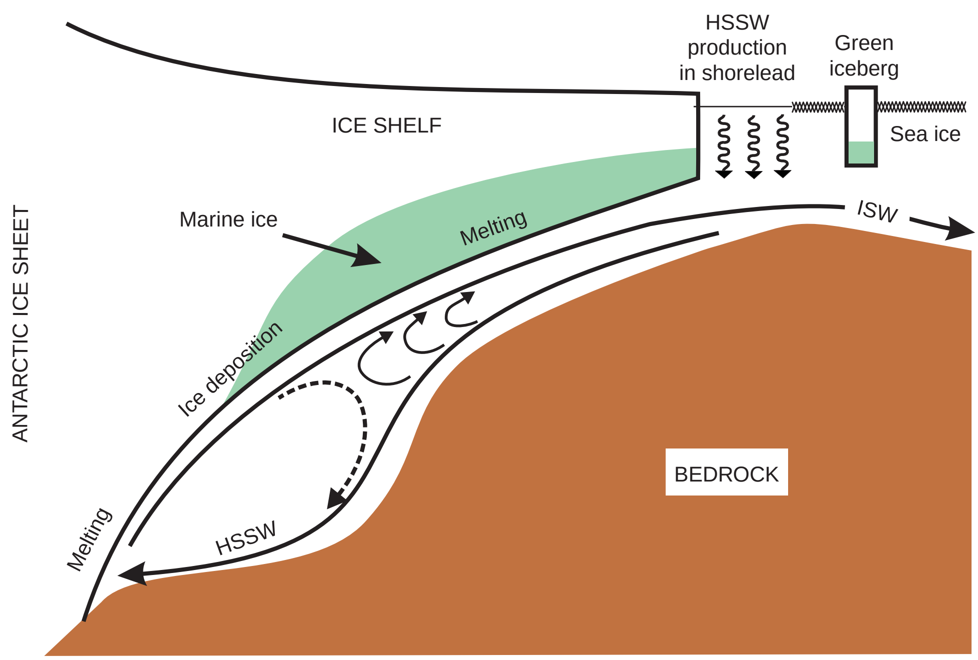

Figure 5 Schematic of the two thermohaline modes of sub-ice shelf circulation for a cold regime ice shelf.

the ice front of several of the larger ice shelves. Water with temperatures up to 0.3°C below the surface freezing point indicated interaction with ice at a depth of at least 400 m, and the $^{18}O/^{16}O$ ratio confirmed the presence of glacial melt water at a concentration of several parts per thousand. Nets cast near ice fronts for biological specimens occasionally recovered masses of ice platelets, again from depths of several hundred meters. The ISW flowing from beneath the Filchner Ice Shelf has been traced overflowing the shelf-break and descending the continental slope, ultimately to mix with deep waters and form bottom water.

Evidence from the ice shelf itself comes in the form of glaciological measurements. By assuming a steady state (the ice shelf neither thickening nor thinning with time at any given point) conservation arguments can be used to derive the basal mass balance at individual locations. The calculation needs measurements of the local ice thickness variation, the horizontal spreading rate of the ice as it flows under its own weight, the horizontal speed of the ice, and the surface accumulation rate. This technique has been applied to several ice shelves, but is time-consuming, and has rarely been used to provide a good areal coverage of basal mass balance. However, it has demonstrated that high basal melt rates do indeed exist near deep grounding lines; that the melt rates reduce away from the grounding line; that further still from the grounding line, melting frequently switches to freezing; and that the balance usually returns to melting as the ice front is approached.

One-dimensional models have been to study the development of ISW plumes from the grounding line to where they detach from the ice shelf base. The most sophisticated includes frazil ice dynamics, and suggests that the deposition of ice at the base depends not only on its formation in the water column, but also on the flow regime being quiet enough to allow the ice to settle at the ice base. As the flow regime usually depends on the basal topography, the deposition is often highly localized. For example, a reduction in basal slope reduces the forcing on the buoyant plume, thereby slowing it down and possibly allowing any ice platelets to be deposited.

Deposits of marine ice become part of the ice shelf itself, flowing with the overlying meteoric ice. This means that, although the marine ice is deposited in well-defined locations, it moves towards the ice front with the flow of the ice and may or may not all be melted off by the time it reaches the ice front. Icebergs that have calved from Amery Ice Front frequently roll over and reveal a thick layer of marine ice. Impurities in marine ice result in different optical properties, and these bergs are often termed ‘green icebergs’.

Ice cores obtained from the central parts of the Amery and Ronne ice shelves have provided other direct evidence of the production of marine ice. The interface between the meteoric and marine ice is clearly visible – the ice changes from being white and bubbly, to clear and bubble-free. Unlike normal sea ice, which typically has a salinity of a few PSU, the salinity of marine ice was found to be below 0.1 PSU. The salinity in the cores is highest at the interface itself, decreasing with increasing depth. A different type of marine ice was found at the base of the Ross Ice Shelf. There, a core from near the base showed 6 m of congelation ice with a salinity of between 2 and 4 PSU. Congelation ice differs from marine ice in its formation mechanism, growing at the interface directly rather than being created as an accumulation of ice crystals that were originally formed in the water column.

Airborne downward-looking radar campaigns have mapped regions of ice shelf that are underlain by marine ice. The meteoric (freshwater) ice/marine ice interface returns a characteristically weak echo, but the return from marine ice/ocean boundary is generally not visible. By comparing the thickness of meteoric ice found using the radar with the surface elevation of the freely floating ice shelf, it is possible to calculate the thickness of marine ice accreted at the base. In some parts of the Ronne Ice Shelf basal accumulation rates of around $1\,\mathrm{m\,a^{-1}}$ result in a marine ice layer over 300 m thick, out of a total ice column depth of 500 m. Accumulation rates of that magnitude would be expected to be associated with high ISW fluxes. However, cruises along the Ronne Ice Front have been unsuccessful in finding commensurate ISW outflows.

Internal recirculation Three-dimensional models of the circulation beneath the Ronne Ice Shelf have revealed the possibility of an internal recirculation of ISW. This mode of circulation is driven by the difference in melting point between the deep ice at the grounding line, and the shallower ice in the central region of the ice shelf. The possibility of such a recirculation is indicated in **Figure 5** by the broken line. Intense deposition of ice in the freezing region salinifies the water column sufficiently to allow it to drain back towards the grounding line. In three dimensions, the recirculation consists of a gyre occupying a basin in the topography of water column thickness. The model predicts a gyre strength of around one Sverdrup ($10^6\,\mathrm{m^3\,s^{-1}}$).

This mode of circulation is effectively an 'ice pump' transporting ice from the deep grounding line regions to the central Ronne Ice Shelf. The mechanism does not result in a loss or gain of ice overall. The heat used to melt the ice at the grounding line is later recovered in the freezing region. The external heat needed to maintain the recirculation is therefore only the heat to warm the ice to the freezing point before it is melted. Ice leaves the continent at a temperature of around $-30°C$, and has a specific heat capacity of around $2010\,J\,kg^{-1}\,°C^{-1}$. As the latent heat of ice is $335\,kJ\,kg^{-1}$, the heat required for warming is less than 20% of that required for melting. To support an internal redistribution of ice therefore requires a small fraction of the external heat that would be needed to melt and remove the ice from the system entirely. A corollary is that a recirculation of ISW effectively decouples much of the ice shelf base from external forcings that might be imposed, for example, by climate change.

Apart from the lack of a sizable ISW outflow from beneath the Ronne Ice Front, evidence in support of an ISW recirculation deep beneath the ice shelf is scarce, as it would require observations beneath the ice. Direct measurements of conditions beneath ice shelves are limited to a small number of sites. Fissures through George VI and Fimbul ice shelves (**Figure 1**) have allowed instruments to be deployed with varying degrees of success. The more important ice shelves, such as the Ross, Amery and Filchner-Ronne system have no naturally occurring access points. Instead, access holes have to be created using hot water, or other thermal-type drills. In the late 1970s researchers used various drilling techniques to gain access to the cavity at one location beneath the Ross Ice Shelf before deploying various instruments. During the 1990s several access holes were made through the Ronne Ice Shelf, and data from these have lent support both to the external mode of circulation, and most recently, to the internal recirculation mode first predicted by numerical models.

Warm regime external ventilation The flooding of the Bellingshausen and Amundsen seas' continental shelf by barely modified CDW results in very high basal melt rates for the ice shelves in that sector. The floating portion of Pine Island Glacier (**Figure 1**) has a mean basal melt rate estimated to be around $12\,m\,a^{-1}$, compared with estimates of a few tens of centimeters per year for the Ross and Filchner-Ronne ice shelves. Basal melt rates for Pine Island Glacier are high even compared with other ice shelves in the region. George VI Ice Shelf on the west coast of the Antarctic Peninsula, for example, has an estimated mean basal melt rate of $2\,m\,a^{-1}$. The explanation for the intense melting beneath Pine Island Glacier can be found in the great depth at the grounding line. At over 1100 m, the ice shelf is 700 m thicker than George VI Ice Shelf, and this results in not only a lower freezing point, but also steeper basal slopes. The steep slope provides a stronger buoyancy forcing, and therefore greater turbulent heat transfer between the water and the ice.

The pattern of circulation in the cavities beneath warm regime ice shelves is significantly different to its cold regime counterpart. Measurements from ice front cruises show an inflow of warm CDW ($+1.0°C$), and an outflow of CDW mixed with glacial melt water. **Figure 6** shows a two-dimensional schematic of this mode of circulation. Over the open

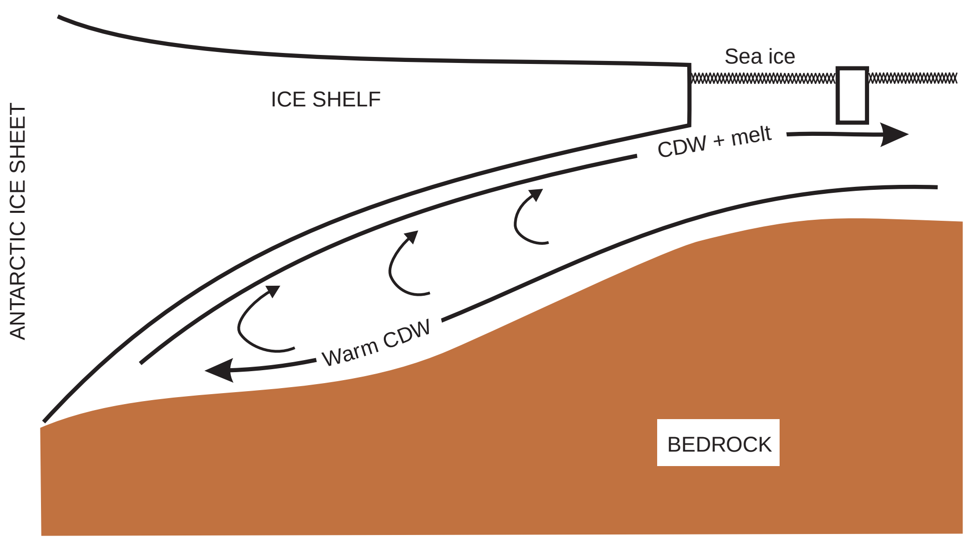

Figure 6 Schematic of the thermohaline mode of sub-ice shelf circulation for a warm regime ice shelf.

continental shelf the ambient water column consists of CDW overlain by colder, fresher water left over from sea ice production during the previous winter. Although the melt water-laden outflow is colder, fresher, and of lower density than the inflow, it is typically warmer and saltier than the overlying water, but of similar density. Somewhat counter-intuitively, therefore, the products of sub-glacial melt are often detected over the open continental shelf as relatively warm and salty intrusions in the upper layers. Again, measurements of oxygen isotope ratio, and also helium, provide the necessary confirmation that the upwelled CDW contains melt water from the base of ice shelves. In the case of warm regime ice shelves, melt water concentrations can be as high as a few percent.

Tidal Forcing

Except for within a few ice thicknesses of grounding lines, ice shelves float freely in the ocean, rising and falling with the tides. Tidal waves therefore propagate through the ice shelf-covered region, but are modified by three effects of the ice cover: the ice shelf base provides a second frictional surface, the draft of the ice shelf effectively reduces the water column thickness, and the step change in water column thickness at the ice front presents a topographic feature that has significant consequences for the generation of residual tidal currents and the propagation of topographic waves along the ice front.

Conversely, tides modify the oceanographic regime of sub-ice shelf cavities. Tidal motion helps transfer heat and salt beneath the ice front. This is a result both of the regular tidal excursions, which take water a few kilometers into the cavity, and of residual tidal currents which, in the case of the Filchner-Ronne Ice Shelf, help ventilate the cavity far from the ice front. The effect of the regular advection of potentially seasonally warmed water from seaward of the ice shelf is to cause a dramatic increase in basal melt rates in the vicinity of the ice front. Deep beneath the ice shelf, tides and buoyancy provide the only forcing on the regime. Tidal activity contributes energy for vertical mixing, which brings the warmer, deeper waters into contact with the base of the ice shelf. **Figure 7A** shows modeled tidal ellipses for the M_2 semidiurnal tidal constituent for the southern Weddell Sea, including the sub-ice shelf domain. A map of the modeled residual currents for the area of the ice shelf is shown in **Figure 7B**. Apart from the activity near the ice front itself, a residual flow runs along the west coast of Berkner Island, deep under the ice shelf. However, this flow probably makes only a minor contribution to the ventilation of the cavity.

How Does the Interaction between Ice Shelves and the Ocean Depend on Climate?

The response to climatic changes of sub-ice shelf circulation depends on the response of the oceanographic conditions over the open continental shelf. In the case of cold regime continental shelves, a reduction in sea ice would lead to a reduction in HSSW production. Model results, together with the implications of seasonality observed in the circulation beneath the Ronne Ice Shelf, suggest that drainage of HSSW beneath local ice shelves would then reduce, and that the net melting beneath those ice shelves would decrease as a consequence. Some general circulation models predict that global climatic warming would lead to a reduction in sea ice production in the southern Weddell Sea. Reduced melting beneath the Filchner-Ronne Ice Shelf would then lead to a thickening of the ice shelf. Recirculation beneath ice shelves is highly insensitive to climatic change. The thermohaline driving is dependent only on the difference in depths between the grounding lines and the freezing areas. A relatively small flux of HSSW is required to warm the ice in order to allow this mode to operate.

The largest ice shelves are in a cold continental shelf regime. If intrusions of warmer off-shelf water were to become more dominant in these areas, or if the shelf-break barrier were to collapse entirely and the regime switch from cold to warm, then the response of the ice shelves would be a dramatic increase in their basal melt rates. There is some evidence from sediment cores that such a change might have occurred at some point in the last few thousand years in what is now the warm regime Bellingshausen Sea. Evidence also points to the possibility that one ice shelf in that sector, the floating extension of Pine Island Glacier (**Figure 1**), might be a remnant of a much larger ice shelf (*see* **Ice-shelf Stability**).

During glacial maxima the Antarctic ice sheet thickens and the ice shelves become grounded. In many cases they ground as far as the shelf-break. There are two effects. The continental shelf becomes very limited in extent, and so there is little possibility for the production of HSSW; and where the ice shelves overhang the continental shelf-break, the only possible mode of circulation will be the warm regime mode. Substantial production of ISW during glacial conditions is therefore unlikely.

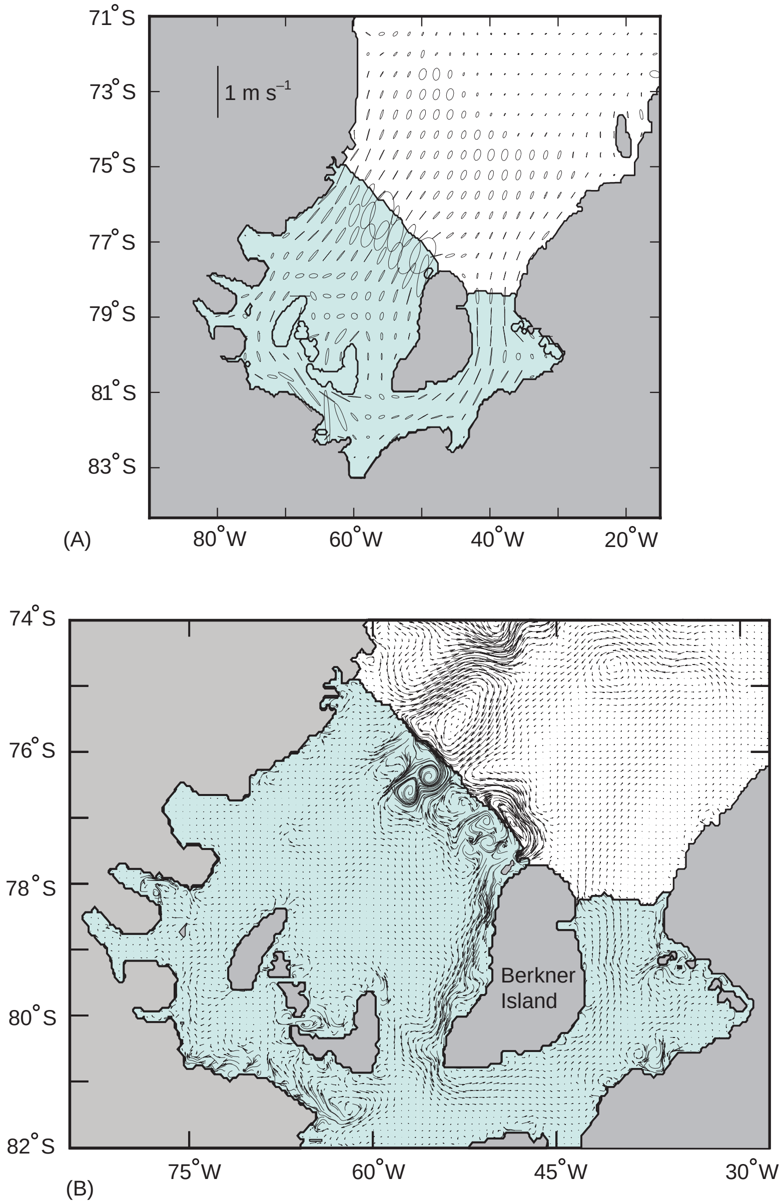

Figure 7 Results from a tidal model of the southern Weddell Sea, in the vicinity of the Ronne Ice Shelf. (A) The tidal ellipses for the dominant M_2 species. (B) Tidally induced residual currents.

See also

Bottom Water Formation. Current Systems in the Southern Ocean. Holocene Climate Variability. Ice–Ocean Interaction. Ice-shelf Stability. Internal Tidal Mixing. Polynyas. Sea Ice: Overview. Shelf-sea and Slope Fronts. Thermohaline Circulation. Tides. Under-ice Boundary Layer. Water Types and Water Masses. Weddell Sea Circulation.

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