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SOUTHERN OCEAN CONTROLS ON CURRENT ICE SHELF EVOLUTION

Annual Progress Report

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Abstract

Our research during the first year has been focused on the eastern sector of the Ross Ice Shelf. Available mooring data covering a one-year cycle of shelf water characteristics and hydrographic data along the eastern Ross Ice Shelf edge indicated this was a good region to study the impact of seasonal shelf water variability on the sub-ice shelf circulation and related basal mass fluxes. Prior one and two-dimensional thermohaline circulation models applied to the sub-ice ocean had been forced by summer shelf conditions only. The modeled sub-ice ocean responds to seasonal forcing at the open boundary with higher spatial and temporal variability. Average melting per year at the ice shelf base increases by > 500%, if a 100-year shelf water temperature increase of 1 °C is assumed. However, these melting rates decrease, if shelf water salinity simultaneously drops, as might be expected from increased melting and lower rates of sea ice formation.

Preliminary results were presented at last year's meeting of the International Glaciological Society in Cambridge, and at an ice shelf workshop of the European Science Foundation held in Germany. A paper based upon the results of this research is nearly ready for submission to a reviewed international journal. Ongoing investigations include the J9-borehole environment of the southern Ross Ice Shelf, and the application of recently obtained hydrographic data to the George VI Ice Shelf cavity.

Annual Progress Report

Up until now, the forcing of thermohaline sub-ice shelf circulation models has been somewhat arbitrary, and based largely upon summer data. Some preliminary results had indicated that the sub-ice circulation and related basal mass fluxes would be sensitive to variable forcing at the model's open boundary. During the first several months of work on this project, hydrographic data covering a one-year cycle at the eastern Ross Ice Shelf edge were used to force a channel-flow model configured to represent sub-ice conditions on two different paths around Roosevelt Island. Model results show that the density distribution along the ice shelf edge supports an anti-clockwise circulation around Roosevelt Island, probably along paths of nearly constant water column thickness. Such flow patterns could exclude the shallow cavity north of Siple Coast from interaction with warmer shelf water masses. By introducing the observed seasonal forcing, average melting increased from 0.07 m/a (no seasonal cycle) to 0.38 m/a. To evaluate the impact of potential long-term changes in shelf water characteristics on the sub-ice environment, a 100-year temperature rise of 1 °C was imposed, resulting in an average melting rate of 2.27 m/a. This rate decreases as shelf water salinity is lowered; a possible scenario, if enhanced melting or less brine drainage were to accompany the warming. Fresher shelf water reduces the horizontal density gradients and, therefore, the strength of the sub-ice circulation.

Preliminary results of this work were presented by SSJ at the Fifth International Symposium on Antarctic Glaciology (VISAG) in Cambridge, UK; abstract and portion of meeting agenda are attached. Following an invitation by the European Science Foundation, HHH attended as the only US participant a workshop on "Ice Shelf - Ocean Interaction" held at the Alfred-Wegener-Institute in Bremerhaven, Germany, 15-16 November 1993. During this meeting HHH learned about ongoing ice shelf research in Norway and France, and renewed contacts with colleagues from Germany and the United Kingdom. This may result in extended visits to Lamont-Doherty by at least one European ice shelf scientist in 1995. In addition we now receive the European Ice Sheet Modelling Initiative Newsletter, which discusses new approaches, latest results, meetings and workshops concerning the evolution of the Antarctic Ice Sheet in a changing climate. Efforts and results summarized above are described in more detail in the attached ms, "Seasonal Circulation Under the Eastern Ross Ice Shelf, Antarctica". This paper is undergoing internal review, following which it will be submitted to JGR or an equivalent international journal.

Continuing the work on this project, we will use an extended configuration of the Ross Ice cavity to simulate the conditions at the J9-borehole site (south-central Ross Ice Shelf). Flow paths discussed in the literature will be tested, and model results will be compared with the sub-ice

observations. Both PI's have just finished a two-month hydrographic program in the Amundsen and Bellingshausen Seas (supported by NSF and DOE) that included measurements along the southern and northern edges of George VI Ice Shelf. This may be one of Antarctica's fastest melting ice shelves, with 2 m/a average melting estimated in the literature. Channel flow simulations will provide the opportunity to model the interaction between this ice shelf and its cavity, which is filled with relatively warm open ocean water masses. Our recent cruise through the Amundsen and Bellingshausen Seas demonstrated that many if not all of the ice shelves in this sector are also underlain by relatively warm water.

Seasonal circulation under the eastern Ross Ice Shelf, Antarctica

by

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Abstract

A seasonal cycle of shelf water temperatures and salinities observed at the eastern Ross Ice Shelf edge was used to force a two-dimensional thermohaline circulation model adapted to two different sub-ice cavity paths around Roosevelt Island. Model results verified by current meter measurements reveal that shelf water flowing into the cavity west of Roosevelt Island might follow a sub-ice path with nearly constant water column thickness of 200 m. This would largely exclude the grounding line of the northern Siple Coast from the contact with open ocean water masses. In contrast to the forcing with time-independent summer profiles, seasonal forcing causes a higher spatial and temporal variability of the cavity's circulation and property distribution. At the model's open boundaries the intermitted inflow of shelf water displaces the meltwater outflow originating from the interior ice shelf base to greater depth and initiates an additional shallow meltwater plume. The average melting along the ice shelf base increases from 0.07 m/a to 0.38 m/a due to seasonality in shelf water characteristics, and to 2.3 m/a, if we introduce a 100-year temperature rise of 1 °C. The rate decreases, if salinity simultaneously decreases; a possible scenario, if the enhanced meltwater outflow mixes with shelf water in front of the ice shelves.

Introduction

In situ observations of the ocean beneath Antarctic ice shelves are rare and of short duration [Jacobs *et al.*, 1979; Nicholls *et al.*, 1991]. Long-term measurements are underway [Orheim *et al.*, 1990; Nixdorf, 1994] or planned [Fahrbach, 1994], but data sets extending for a year or longer in these sub-ice cavities are not yet available. However, an ice shelf cavity interacts with the continental shelf through a circulation driven mainly by thermohaline (density) differences [Robin, 1979], so that observations along the ice shelf edge can provide indirect evidence for sub-ice processes [Jacobs *et al.*, 1979; Weiss *et al.*, 1979; Schlosser *et al.*, 1990]. In recent years attention has been focused on the flow of glacial meltwater from underneath Antarctic ice shelves [Foldvik *et al.*, 1985a; Potter and Paren, 1985; Hellmer and Jacobs, 1992]. With temperatures below the sea surface freezing point, Ice Shelf Water (ISW) is the characteristic ingredient in one of the bottom water formation processes along the Weddell Sea continental slope [Foldvik *et al.*, 1985a; Jacobs, 1986; Gordon *et al.*, 1993]. ISW has also been observed on the continental shelves of the Ross Sea [Jacobs *et al.*, 1970] and elsewhere. As demonstrated by numerical models applied to the Ross and Filchner-Ronne Ice Shelves [MacAyeal, 1985a; Hellmer and Olbers, 1989; Jenkins and Doake, 1991], ISW can result from the interaction of High Salinity Shelf Water (HSSW) with the ice shelf base. With temperatures near the sea surface freezing point, HSSW provides heat for melting ice at deeper levels of the interior cavities due to the reduction of the sea water freezing point with increasing pressure [Doake, 1976].

Year-long current meter measurements from the sill of the Filchner Depression in the southern Weddell Sea reveal a seasonal signal in the ISW flow, with stronger mean current speeds in winter than in summer [Foldvik *et al.*, 1985b]. This might be an indication of sub-ice shelf seasonality, or the imposition of an open ocean seasonal cycle in the 500 km distance from the Filchner Ice Shelf edge to the mooring site. Little is known about the seasonal variability of

shelf water characteristics along ice shelf edges since all hydrographic sections there have been taken during summer. The surface layer is freshened by ice melting and warmed by solar radiation during summer, processes that are reversed during winter sea ice formation. Depending on density stratification of the water column, the volume of brine released, and offshore advection and consequent upwelling, deep convection on the shelf will not everywhere reach the sea floor. With an average draft of 250 m [*Jacobs et al.*, 1992], ice 'barriers' can also block seasonally modified shelf water from flowing into the sub-ice cavities. Close to some ice shelf edges, off-shore winds and tidal cycles can maintain areas of open water [*Zwally et al.*, 1985]. These areas may be favorable for sea ice production during most of the year [*Foster*, 1972], potentially reducing the seasonal variability. In any case, some prior work has shown a general agreement between glaciological and oceanographic observations and model results, achieved by prescribing 'summer' profiles at the open boundary [*Jenkins and Doake*, 1991; *Hellmer and Olbers*, 1991].

Initial attempts to force a channel-flow model with an arbitrary annual, sinusoidal cycle in shelf water salinity (depth-independent amplitude of 0.01 psu superimposed on the summer values) changed the strength of the through-flow, displaced zones of attrition and accumulation, and increased melting and freezing rates [*Hellmer and Olbers*, 1991]. That amplitude, applied to the Filchner-Ronne cavity, is almost an order of magnitude less than has been measured near the Ross Ice Shelf edge [*Pillsbury*, unpublished data]. Hydrographic data through annual cycles were obtained from current meters deployed along the Ross Ice Shelf edge in 1983 and 1984 (**Figure 1**). These Aanderaa instruments recorded current speed and direction together with temperature and, in some locations, conductivity [*Pillsbury and Jacobs*, 1985; *Visser and Jacobs*, 1987]. At mooring *H* on the west side of Roosevelt Island, the flow was weak and meandering into the Ross Ice Shelf cavity (**Figure 2a**). At mooring *I* east of Roosevelt Island a persistent outflow for water depths greater than the ice shelf draft was recorded for most of the year (**Figure 2b**). The latter was apparently responsible for the initial northwesterly drift of

iceberg B9 in 1987 [*Keys et al.*, 1990]. Conductivity-Temperature-Depth (CTD) stations close to the mooring sites provided full-depth temperature and salinity profiles. This unique data set provides the opportunity to study the influence of a recorded seasonal cycle in shelf water characteristics on sub-ice shelf ocean dynamics and related ocean/ice shelf processes. We applied the two-dimensional channel flow model, used by *Hellmer and Olbers* [1991] to investigate the direction of flow south of Berkner Island, to the eastern Ross Ice Shelf domain (**Figure 1**). After a short description of the model, and forcing at the open boundaries, we present the results for two different cavity configurations, one including a shallow sector which approximates parts of the northeastern Siple Coast grounding line. Additional simulations were carried out to investigate how a drift in shelf water characteristics over a period of 100 years could alter mass fluxes at the ice shelf base.

The Model

Description

We outline here the basic elements of a two-dimensional (y/z -plane) channel flow model forced by thermohaline (density) differences between the open boundaries and the interior cavity [*Hellmer and Olbers*, 1989; 1991]. Boussinesq and hydrostatic approximations are used for the momentum balance in the y/z -plane. The direction of y is perpendicular to the ice shelf front and parallel to the paths illustrated in Figure 1 ($y=0$ at the eastern open boundary). The z -direction is positive upward with $z=0$ at the sea surface. The influence of earth rotation cannot be considered due to the reduction of one horizontal dimension. Tides are believed responsible for enhanced water mass exchange at the ice shelf edge [*Gammelsrød and Slotsvik*, 1981] and vertical mixing in shallow regions of the cavity [*MacAyeal*, 1985b], but are not included at this stage of model development. Heat and salt balances with the usual form of diffusion and convective adjustment parameterization, and a nonlinear equation of state supplement the dynamic part of the model.

Following *Jenkins* [1991], heat and salt transfer coefficients at the ocean/ice shelf interface depend on the velocity calculated for the grid-box below the upper boundary. For the molecular diffusive heat flux through the ice (neglected for salt) we assume a linear temperature profile, ranging from a salinity and pressure dependent freezing temperature at the base [*Jacobs et al.*, 1979] to an annual mean surface value of -25°C [*Thomas*, 1976]. These formulations allow for the melting or the formation of ice at the ice shelf base as a result of in situ supercooling; crystal transport with the current is not yet considered in this model.

Configuration

Assuming that the sub-ice flow will tend to conserve potential vorticity by following a path of nearly constant water column thickness (WCT), the cavity shape for the first set of simulations has been derived from topographic data around Roosevelt Island (the 500 km black line in **Figure 1**). With the exception of the ice shelf edge region, WCT is held constant at approximately 200 m. The inclusion of a grounding line segment (gray extension in **Figure 1**) adds 100 km of reduced WCT (80 m) and 100 km of decreasing or increasing WCT for a total length of 700 km. Gradients of sea floor depth and ice shelf thickness are approximated stepwise with constant increments of $\Delta y = 20\text{ km}$ and $\Delta z = 20\text{ m}$. The dynamic response of the ice shelf and temporal thickness variations are neglected due to the short (5-year) simulation time, during which melting rates of 1 m/a remove $\sim 1\%$ of ice. The deposition of sea ice at the base is even less by an order of magnitude. The assumption of net glacial balance might become critical for high melting regions (e.g. close to the grounding line) when integrated over a 100-year period. The idealized model configuration is represented in the figures described below illustrating the two-dimensional fields of sub-ice quantities.

Forcing

All simulations were started from stationary states. These resulted from a 3-year integration of the corresponding cavity configurations, forced by hydrographic conditions at the open boundaries, as represented by the summer profiles of stations 36 and 43. For the seasonal cycle, annual ranges of $\Delta T = 0.2$ °C and $\Delta S = 0.15$ psu were derived from the year-long temperature and salinity records of mooring *H* (Figure 3). By superimposing these values on both summer profiles, we assumed that the seasonal variability does not change along a 200 km length of the eastern ice shelf edge, and that the record from 305 m at site *H* is representative of the whole water column. Temperature time series from two depths at site *I* do show a decrease of the seasonal signal with depth, but this site represents cavity outflow, the modeled profiles of which should reflect the cavity processes. The seasonal variability in temperature and salinity at 550 m at site *H* (Figure 3) shows a coincidence of temperature maximum with salinity minimum, amplifying the annual shelf water density range. Shelf water characteristics are transferred into the cavity where the velocity field at the open boundaries indicates inflow, and the initial profiles are reset to the modeled cavity characteristics when outflow occurs.

The 500 km-Cavity

Results

Quarterly snapshots of an annual cycle beneath the eastern Ross Ice Shelf (Figure 4) begin 3 months after observations started at site *H*. At day 364 the seasonal cycle approximates the results of the time-invariant (summer profile) forcing, but the latter solution is fresher, with smaller (< 1.5 cm/sec) velocities.

The salinity (density) gradient along the ice shelf edge decreases eastward initiating a flow through the sub-ice cavity from the right (west) to the left (east) for most of the year. The strength varies with time, with slightly higher velocities (> 10 cm/sec) in the inflow cavity. A weaker through-flow (< 2 cm/sec) at day 91 just below 500 m separates three counter flows. One represents a meltwater plume on the western side which starts rising along the ice shelf base at kilometer 250. At the western entrance this plume occupies the upper 150 m of the water column, and has a core velocity > 2 cm/sec at 400 m depth. An analogous but weaker circulation with an inflow at 400 m depth appears in the upper 130 m on the eastern side. Both of these shallow plumes result from shelf water contact with the ice ~ 100 km south of the ice fronts. A deeper return flow transports water masses toward the western slope, but mixes with inflowing shelf water and does not emerge from the cavity.

The spatial variability of the salinity field is higher than that of the velocity field during the year. Relatively low late summer salinities (**Figure 3**) at the ice shelf edge result in a fresh cavity at day 91. The minima close to the ice base at both edges are related to the fresher summer inflow and the melt water plumes mentioned above. The interior bottom salinity maximum of day 91 is a remnant from the previous year, apparent at the western end of the cavity at day 182. This feature diminishes by day 182 and has been replaced by day 273 (end of October). The new salinity maximum again ponds in the depression by day 364, and is diluted by mixing with the melt-laden water above and with the year's freshest bottom inflow around day 91. *Hellmer and Jacobs [1992]* suggested that such dense bottom layers may in some locations protect ice shelf grounding lines from potentially warmer shelf waters. But in this cavity configuration, the salty (dense) lens may guide the higher temperature late-summer shelf water closer to the ice shelf base (e.g. day 91).

Variability in the temperature field is confined to the entrances, and to the eastern cavity (sequence not shown). The fluctuations at both edges follow the seasonal cycle illustrated in

Figure 3, i.e. a warm (cold) inflow is fresh (salty). Higher temperatures at the western edge are caused by slightly warmer shelf water observed near station 36 [Pillsbury and Jacobs, 1985; Fig. 2a]. However, during an annual cycle, the core of the main melt water plume leaving the eastern cavity occupies different levels of the water column with decreasing temperatures as it moves upward. During the period of warm and fresh inflow from the east (day 91), the plume detaches from the deep ice shelf base at temperatures above the in situ freezing point. Due to mixing with the inflowing warmer water, the plume loses its cold imprint before the eastern edge is reached. As the strength of the upper inflow decreases, the plume detaches at shallower depths with colder temperatures. A 160 m ascent along the entire eastern ice shelf base produces the lowest outflow temperatures at day 273.

Discussion of the 500 km-cavity results

The year-long current measurements show a predominant flow at both levels toward 350° [Figure 2b], suggesting that the whole water column at site 1 experiences an outflow from the cavity. The depth of the upper current meter (255 m) was roughly equal to the ice shelf draft [Keys et al., 1990]. However, only the upper level shows any temperatures at the surface freezing point for a salinity of 34.31 psu, and that only briefly in mid-December. The warmer temperatures from late December thru April (the austral summer) are presumably caused by seasonal warming in the open sea upstream of the mooring site. We forced the sub-ice ocean with a depth-independent seasonal range, but the modeled water column at the outflow site shows a larger temperature range at the upper level (Figure 5), in agreement with the observations. However, the modeled temperature time series differs primarily by being too cold in the later half of the year. Possible reasons for this discrepancy include a stronger winter forcing from the sea surface, effectively homogenizing the upper water column. This could overwhelm an outflowing plume of limited dimensions or meandering nature. Tidal mixing

might also damp a cold outflow signal. It may be relevant to note that CTD station 43 recorded a minimum temperature of -1.95°C at a depth of 256 m on 28 January 1984, midway between the two model results, but substantially colder than the current meter temperature record.

At 540 m the current speed was more than twice that at 255 m, indicating a strong cavity outflow. In addition to the small but systematic difference between model results and observations (**Figure 5**), the site *I* bottom layer at $\sim -1.8^{\circ}\text{C}$ is at or above the inflow temperatures of station 36 which ranged from -1.83°C to -1.86°C on 27 January 1984. Again, mixing near the ice shelf edge by tides or by the velocity shear might warm the melt water plume on its way to site *I*, ~ 10 km north of the ice shelf edge. But if the flow paths are correct, a through-flow with any ocean/ice shelf interaction should cause outflow temperatures lower than observed. One explanation may be that warmer inflows closer to the west side of Roosevelt Island (e.g. -1.78°C at 450 m on station 39) are more important than the meandering current at mooring *H*/station 36; we chose station 36 because of its proximity to the mooring site.

The site *I* current meter record at 540 m depth can be divided into 3 different periods: a medium outflow (~ 5 cm/sec) from February to May, changes in the flow direction from May through September, and a strong outflow with maximum velocities of 14 cm/sec during the late winter and summer months October to January (**Figure 6**). Model results also show periods of moderate, low and high near-bottom outflow roughly coincident in magnitude with the measurements. The simulated time interval and appearance of the events differs from the observations, with the best fit obtained by a 700 km flow path and 200 m-thick water column in the interior. In effect, this implies that the circulation around Roosevelt Island bypasses the grounding line. The model does not show the observed flow reversals toward the southeast during winter (May to September) which may be driven by changes in the strength of the along-shore flow, if it extends to 540 m-depth. However, the modeled through-flow can change to two separate circulation cells with bottom inflow at each cavity end, if the density gradient along the

ice shelf edge decreases to zero. This situation occurs most likely in winter when atmospheric forcing is strong enough to homogenize the water column in front of the ice shelf.

The 700 km-Cavity

Results

To investigate the effects of a change in the flow path, we extended the sub-ice cavity by 200 km (gray line in **Figure 1**), adding a section of shallower water depth (80 m) below a deeper ice shelf base adjacent to the Siple Coast grounding line. Days 91 and 273 of the velocity and salinity fields are selected from the annual cycle of the longer cavity (**Figure 7**) to illustrate the major differences between the two simulations. By day 273 maximum salinities (> 34.6 psu) had penetrated only 150 km into the cavity, compared with a total filling of the interior in the 500 km-version, due to a 30% weaker inflow. As illustrated by the salinity maximum at ~ 475 km (> 34.56 psu) at day 91, shelf water is diluted upon approaching the shallowest region, which is filled with a less saline and colder water mass. This freshening and cooling in the interior agrees with sub-ice observations from the Ronne Ice Shelf [Nicholls *et al.*, 1991]. Supported by 50 % higher velocities beneath a deeper ice shelf base (620 m), ocean/ice shelf interaction near the grounding line initiates a more vigorous melt water plume, and fresher, cooler water masses dominate the outflow (eastern) cavity for most of the year. The modeled temperature time series for the eastern edge also show a 0.05 °C cooler upper temperature minimum and a ~ 0.1 °C cooler bottom layer (**Figure 5**). Bottom temperatures of -1.98 °C would require a mixing ratio of 1 : 9 between the lower sub-ice outflow and shelf water masses to match the site *I* observations (**Figure 5**). With the exception of the grounding line sector velocities were similar to day 91 or lower than day 273 in the 500 km-cavity. Modeled outflow velocities at 540 m depth reached a minimum in May, followed by a spring maximum from October to November (**Figure 6**). Although the observations showed more variability from May to September, periods of inflow,

and maximum values during November to January, the 700 km-route extended the period of near zero flow, and came closer to depicting the observed maximum late in the year.

It is apparent that enhanced ocean/ice interaction at a deeper ice shelf base does not necessarily increase the transport through the cavity. In a self-regulating effect, the additional melt water reduces the driving density differences and, consequently, the sub-ice velocities. The results of simulations with different interior water depths (80, 160 and 200 m) (**Figure 6**) show that outflow velocities increase with increasing interior water column thickness. The difference between the 500 km and 700 km simulations suggest that the circulation beneath the eastern Ross Ice Shelf may follow a longer path with a preferred water column of ~ 200 m thickness. Since outflow temperatures also approach the observed values with increasing interior water column thickness (**Figure 5**), the northern Siple Coast grounding line may have little, if any influence on the outflow characteristics to the east of Roosevelt Island.

Basal mass fluxes

Velocities and related melting rates might be overestimated for the grounding line sector in the 700 km simulation, since the two-dimensional flow forces major portions of the shelf water through the shallow region. Therefore, we will focus on the basal mass flux distributions along the 500 km ice shelf base (**Figure 8**).

As a result of a slow moving sub-ice ocean (average velocity 1 cm/sec), the maximum melt rate caused by the time-independent forcing is less than 0.15 m/a. By introducing a seasonal cycle, average melting rises from 0.07 m/a to 0.38 m/a, is more variable along the ice shelf base and can be related to changes in ice shelf thickness. Slightly higher melting at the western base results from the direct contact of inflowing shelf water masses with the ice (**Figure 4**, days 273

and 364). Since the in situ freezing temperature is primarily a function of pressure, melting at the shallow ice shelf base is too low to initiate an ascending plume. Instead, the glacial melt mixes with the inflowing water without changing its buoyancy. The melting rate reaches a maximum of ~ 0.7 m/a at the deepest portion of the ice shelf base. Major contributions to the annual rate occur during times the upper water column is occupied by the shallow recirculation (Figure 3, days 91 and 182), and 'undiluted' shelf water interacts with the ice shelf at greater depth. Latent heat lost to the melting process, the increase in freezing temperature with decreasing depth, and no significant entrainment of warmer water from below reduces the plume's heat capacity as it ascends along the eastern base. Close to the ice shelf edge a drop of the plume temperature below the in situ freezing point occurs during periods of up to 30 days. However, accumulations of ice crystals at the base are wiped out in the annual average by melting related to the shallow inflow of seasonally variable shelf water.

Large changes in shelf water characteristics

There is evidence of considerable variability in shelf water characteristics over the past few decades [Jacobs, 1985], and some portions of the Antarctic continental shelf contain substantially warmer water than does the Ross Sea. Future climate change that involves rates of sea ice and cold, dense shelf water formation, or ocean circulation strength much different from today might bring such conditions to the eastern Ross Sea. To evaluate the impact of substantially different shelf water characteristics on basal mass fluxes, we assumed that over a 100-year period modified Circumpolar Deep Water [the Ross Sea's "warm core"; Jacobs et al., 1985] might replace today's shelf water in the eastern Ross Sea. We superimposed the same seasonal range as in the previous simulations, but allowed temperatures at stations 36 and 43 to increase linearly by 1.0°C over a century's time, with salinities held constant at today's values. In a second model run a simultaneous decrease in shelf water salinity of 1 psu was implemented

to help balance increasing density differences between the open ocean and the cavity as higher basal melting freshens the sub-ice ocean. A desalinization of the shelf water masses initiated by the flow of Modified Weddell Deep Water (MWDW) with temperatures $< -1.3^{\circ}\text{C}$ into the sub-ice cavity has been discussed by Jenkins [1991]. Lower salinities could also result from less sea ice formation and higher precepitation.

The temperature and temperature/salinity trend simulations, did not change the spatial structure of the melting curve caused by the seasonal forcing, but increased the melt rates (**Figure 8**). For the temperature trend alone, average melting amounted to 2.27 m/a, 500% of the trend-free value, and similar to that predicted by Jenkins [1991] for the Ronne Ice Shelf cavity filled with MWDW. For the temperature/salinity trend simulation, average melting relaxed to 1.76 m/a, mainly due to a 50% decrease of the cavity through-flow. The fact that the shape of the melting curve remains the same for different model runs indicates that the ice shelf thickness distribution has major influence on the strength of the ocean/ice shelf interaction. Therefore, simulations covering more than several decades will require an adjustment of the ice shelf thickness distribution related to the ice dynamics, which was not considered here.

Conclusions

Model results show that the density structure in front of the eastern Ross Ice Shelf edge supports an anti-clockwise circulation around Roosevelt Island, regardless whether the sub-ice ocean is forced with time-invariant summer profiles or with the observed seasonal cycle in shelf water characteristics. Hydrographic features observed in the outflow region between Roosevelt Island and Edward VII Land are more closely approximated when the forcing at the model's open boundaries includes seasonal variability. From differences between the measurements and model results we infer that the inflowing shelf water may tend to follow sub-ice paths of almost

constant water column thickness in order to conserve potential vorticity. This could exclude much of the open-sea, warmer shelf water from direct contact with the grounding line. However, the sub-ice topography might as well direct water masses flowing in at site *H* to the southeastern Ross Ice Shelf cavity. *Nøst and Foldvik* [1994] showed that those water masses might have been observed at J9. Modified in the shallow cavity off Siple Coast, they also could have caused the accumulation of sea ice as observed at the borehole by *Zotikov et al.* [1980].

At depths below the ice shelf draft the coastal current mixes with a relatively warm sub-ice outflow. Previously, plume temperatures warmer than -1.9°C have been calculated only for plumes detaching from the base near an ice shelf edge with almost zero draft [MacAyeal, 1985; Table 1]. An average melt rate of 0.38 m/a results in net melting along the path of $1.84 \times 10^5 \text{ m}^3/\text{yr}$ (water equivalent). A total outflow of $2.8 \times 10^8 \text{ m}^3/\text{yr}$ to the east implies that the fraction of glacial melt water transported out of the cavity per unit width in meter amounts to 0.6 ‰. This is an order of magnitude smaller than calculated for the ISW plume in the Filchner Depression [Schlosser et al., 1990]. However, Western Shelf Water flowing underneath the Filchner-Ronne Ice Shelf already contains a 3.2‰ fraction of glacial melt [Weiss et al. 1979]. The temporal and spatial variability of the upper circulation cell, which at maximum strength forces the interior melt water plume to flow out at greater depth cannot yet be verified by observations in front of the ice shelf. Although, current meter measurements to the west of site *H* indicate the existence of a shallow circulation. Its interaction with the ice base close to the ice shelf edge results in the outflow of Shallow Ice Shelf Water, the warmer 'relative' of the main Deep Ice Shelf Water outflow [Jacobs et al., 1985].

Existing seasonal shelf water increases the average basal melting rate above that of a static environment. Interannual variability and long-term trends have the potential to change the sub-ice circulation and basal mass balance even more. Long-term monitoring of shelf water along the Ross Ice Shelf edge would help to define the temporal variability of shelf water

characteristics and water mass exchange between the open ocean and the sub-ice ocean. If done in sufficient detail, combined with a 3-dimensional numerical model and supplemented by a few boreholes through the ice we should eventually be able to simulate the behavior of the ice shelf in a changing climate.

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Figure Captions

Fig. 1 Map of water column thickness distribution beneath the eastern Ross Ice Shelf in meters including the position of the grounding lines around Roosevelt Island and along Siple Coast (dashed lines) [after *Greischar and Bentley*, 1985]. Superimposed are the locations of mooring sites *H* and *I* (open squares) and hydrographic stations 36 and 43 (crosses). Model geometries for the 500 km and 700 km simulations correspond to the sub-ice cavity shape along the black line and the black plus grey lines, respectively.

Fig. 2a Progressive vector diagram of ocean currents at mooring site *H* ($78^{\circ} 09' \text{ S}$, $170^{\circ} 37' \text{ W}$) recorded at 305 m depth for a 373-day period starting on 27 January 1984. Data are low-pass filtered (tides removed), and the beginning of each month is indicated by a lettered square. Negative values correspond to southward/westward flow.

Fig. 2b Progressive vector diagrams of ocean currents at mooring site *I* ($77^{\circ} 41' \text{ S}$, $160^{\circ} 24' \text{ W}$) 244 km east of mooring site *H*. Records are from 255 m depth (solid line, bottom and right scales) and 540 m depth (dashed line, bottom and left scales) for a 371-day period starting on 28 January 1984. Data are low-pass filtered (tides removed), and the beginning of each month is indicated by a lettered square (255 m) or triangle (540 m). Negative values correspond to southward/westward flow.

Fig. 3 Temperature and salinity time series from mooring site *H* (dotted lines), and the approximations (thick lines) applied to the model's open boundaries. The salinity record has been adjusted to the salinity profile obtained during the recovery of the mooring. Both data sets were smoothed by a 14-day running mean filter.

Fig. 4 Four snapshots of the 500 km sub-ice cavity route (Figure 1) illustrating the annual variability of the velocity field superimposed on the corresponding salinity distribution. Inflow occurs at the right side and outflow at the left. The maximum ice shelf thickness and deepest depths represent conditions south of Roosevelt Island. Julian days 91, 182, 273, 364 correspond to the ends of April, July, October and January. The contour interval is 0.5 cm/sec for days 91 and 364, and 1 cm/sec for days 182 and 273.

Fig. 5 Temperature time series recorded at 255 m and 540 m depth at mooring site *I* (solid lines) compared with time series from the eastern ice shelf edge resulting from the 500 km (thick solid lines) and 700 km (gray lines) simulations. The thin solid line in the top graph represents the surface freezing point for a salinity of 34.31 psu measured at 255 m depth at station 43. An additional time series has been added to the bottom graph (dotted line) to illustrate the warming of the outflow due to an 120 m-increase of water column thickness in the grounding line sector of the 700 km long cavity.

Fig. 6 Daily averaged northward component of the currents at site *I* at 540 m depth (thin solid line) compared with modeled velocities from the corresponding level at the eastern cavity end. Time series result from the simulation of the 500 km cavity (thick solid line) and three 700 km cavities, which have 80 m (dotted line), 160 m (short dashed line) and 200 m (long dashed line) interior water columns.

Fig. 7 Same as in Figure 3 but for those two states of the 700 km cavity which show major differences compared to the 500 km cavity simulation. The contour interval changes from 0.5 cm/sec at day 91 to 1 cm/sec at day 273.

Fig. 8 Melting rate distributions at the ice shelf base of the 500 km cavity and the annual averages for different shelf water states prescribed at the model's open boundaries; summer only (dashed line), seasonal cycle (solid line), 100-year 1 °C temperature increase (thick solid line), and 100-year 1 °C temperature increase/1 psu salinity decrease (dotted line).

Figure 1

Figure 2a

Figure 26

Figure 3

Figure 4

Figure 4

Figure 5

Figure 6

Salinity

Figure 7

Figure 8

Tuesday 7th September: Rank Room

ICE/OCEAN. Chairman: Ian Allison

- 0830 Nixdorf, U.; Oerter, H.; Miller, H. *First access to the ocean beneath the Ekstroem Ice Shelf by means of hot-water drilling.* (38)
- 0850 Orheim, O.; Osterhus, S. *Heat exchange and mass balance at the underside of the ice shelf Fimbulisen.* (162)
- 0910 Robinson, A.; Makinson, K.; Nicholls, K. *Ocean circulation beneath Ronne Ice Shelf.* (164)
- 0930 Bombosch, A.; Jenkins, A. *Modelling the formation and deposition of frazil ice beneath Filchner- Ronne Ice Shelf.* (118)
- 0950 Determann, J.; Gerdes, R. *Melting and freezing beneath ice-shelves: implications from a 3D-ocean circulation model.* (28)
- 1010 Hellmer, H.H.; Jacobs, S.S. *Ross ice shelf cavity response to annual shelf water variability.* (80)

1030 Morning Coffee in Foyer

- 1100 Jenkins, A. *Thermohaline convection driven by the melting of Antarctic ice shelves and icebergs.* (157)

MODELLING I. Chairman: Ian Whillans

- 1120 Huybrechts, P. *Formation and disintegration of the Antarctic ice sheet.* (123)
- 1140 Budd, W.F.; Jenssen, D.; Mavraklis, E.; Coutts, B. *Modelling the Antarctic ice sheet changes through time.* (109)
- 1200 Genthon, C. *Antarctic climate modelling with general circulation models of the atmosphere.* (101)
- 1220 Connolley, W.M. *Accumulation over Antarctica as represented by the UK Met. Office GCM.* (30)

1240 Lunch in Hall

ICE CORES II. Chairman: Bernhard Stauffer

- 1400 Petit, J.R.; Martin, M.; Jouzel, J.; Barkov, N.I.; Vaikmae, R. *The dust record in Antarctic ice: new data and comparison with other records.* (178)
- 1420 Higashi, A.; Fujii, Y. *Studies on microparticles contained in medium depth ice cores retrieved at East Queen Maud Land, Antarctica.* (23)
- 1440 Thompson, L.G.; Peel, D.A.; Mosley-Thompson, E.; Mulvaney, R.; Dai, J.; Lin, P.N.; Davis, M.E.; Raymond, C.F. *Climate since 1510 AD on the Dyer Plateau: evidence for recent climate change.* (190)
- 1500 Nishio, F.; Watanabe, O.; Jacobs, P.; Oerter, H. *Weddell polynya recorded in ice core.* (46)
- 1520 Clais, P.; White, J.W.C.; Petit, J.R.; Jouzel, J. *The origin of Antarctic precipitations from surface snow deuterium excess data.* (180)

1540 Afternoon Tea in Foyer

- 1610 Martinerie, P.; Ya Lipenkov, V.; Chappellaz, J.; Raynaud, D. *The air content record in the Vostok ice core.* (98)

Ross Ice Shelf Cavity Response to Annual Shelf Water Variability

by

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Basal melting of ice shelves is an important component in the mass balance of the Antarctic ice sheet. This mass loss is controlled by the strength of the sub-ice thermohaline field which drives the inflow and outflow of "warm" and cold shelf waters. Numerical models demonstrate the strong dependence of the thermohaline circulation on the characteristics and seasonal variability of shelf water. In results reported to date, only summer observations have been used as open boundary conditions, and seasonality has been imposed in an arbitrary way. Here we use year-long temperature and salinity measurements near the front of Ross Ice Shelf to force the sub-ice circulation in a 740-km channel around Roosevelt Island, with realistic ice shelf thickness and sea floor depth distributions. Several simulations were conducted to isolate the effects of the spatial and temporal variability of temperature and salinity. Seasonal fluctuations eliminate a freezing regime east of the Island and extend melting over the entire channel with an average rate ~ 1 m/yr along the 120-km Siple Coast grounding line. An inflow at mid-depth draws shelf water with open-ocean properties ~ 100 km into the cavity. This water recirculates at shallower depths, still above the freezing temperature, and is a heat source for enhanced melting near ice shelf edge.

**DATE
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9/12/94

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