

Ocean/ice shelf interaction in the southern Weddell Sea: Results of a regional numerical helium/neon simulation

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Ocean Dynamics, Volume 57, Number 1 / Februar 2007,
doi: 10.1007/s10236-006-0073-2
hdl:10013/epic.23220

Received: 1 June 2005 Accepted: 11 February 2006 Published online: 10 May 2006

Abstract

Ocean/ice interaction at the base of deep drafted Antarctic ice shelves modifies the physical properties of inflowing shelf waters to become Ice Shelf Water (ISW). In contrast to the conditions at the atmosphere/ocean interface, the increased hydrostatic pressure at the glacial base causes gases embedded in the ice to dissolve completely after being released by melting. Helium and neon with an extremely low solubility are supersaturated in glacial meltwater by more than 1000%. At the continental slope in front of the large Antarctic caverns ISW mixes with ambient waters to form different precursors of Antarctic Bottom Water. A regional ocean circulation model is presented which uses an explicit formulation of the ocean/ice shelf interaction to describe for the first time the input of noble gases to the Southern Ocean. The results reveal a long-term variability of the basal mass loss solely controlled by the interaction between waters of the continental shelf and the ice shelf cavern. Modeled helium and neon supersaturations from the Filchner-Ronne Ice Shelf front show a "low-pass" filtering of the inflowing signal due to cavern processes. On circumpolar scales the simulated helium and neon distributions allows us to quantify the ISW contribution to bottom water which spreads with the coastal current connecting the major formation sites in Ross and Weddell Seas.

keywords: Tracers – Ice Shelf Water – Antarctic Bottom Water – Southern Ocean – numerical model – Weddell Sea

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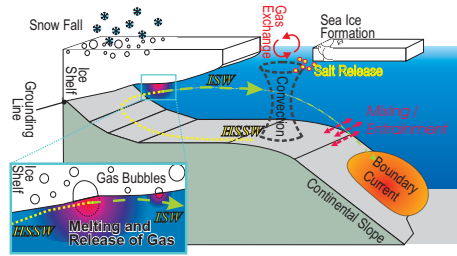


Figure 1: Schematic of the formation of High Salinity Shelf Water in front and its transformation to Ice Shelf Water within an ice shelf cavern. The inset illustrates the transfer of gas from the glacial ice to the meltwater plume.

1 Introduction

The sinking of dense water masses at high latitudes drives the ventilation of the global abyss. In the Southern Ocean the formation of deep and bottom water primarily occurs along the slopes of continental shelves placed in front of the large Antarctic ice shelves Ross, Amery, and Filchner-Ronne (Baines and Condie, 1998) (locations are shown in Fig. 9). Here, heat loss to the atmosphere and related salt gain due to sea-ice formation reduce the stability of the water column. As a result, deep convection homogenizes the whole water column and transfers salt from the surface to the bottom thus forming High Salinity Shelf Water (HSSW). This water mass contributes both to the ventilation of the ice shelf caverns and to the formation of new water masses at the continental shelf break. Two mixing processes are known to form deep and bottom water:

- the Foster-Carmack process (Foster and Carmack, 1976): pycnocline waters penetrate onto the continental shelf to mix with locally formed HSSW,
- the Foldvik process (Foldvik et al., 1985): Ice Shelf Water (ISW), characterized by temperatures below the surface freezing point (Carmack, 1974), mixes directly with ambient deep water as it descends the continental slope.

The second process includes the so-called “ice pump” (Lewis and Perkin, 1985) circulation, by which near-bottom HSSW flows towards the ice shelf grounding line, driven by gravity and the density contrast between ice shelf front and cavern’s interior, and controlled by the water column thickness (Gerdes et al., 1999). HSSW is transformed to ISW by cooling below the surface freezing temperature (due to the pressure dependence of the in-situ freezing point (Gade, 1979)) and freshening due to deep basal melting (Fig. 1). At the same time, gases captured in the ice matrix during the transformation from snow to ice are released and, due to the high hydrostatic pressure, completely dissolved in the meltwater plume. This results in extreme supersaturations of helium and neon of 1400% and 1060%, respectively (Schlosser, 1986). Although Hohmann et al. (2002) report helium and neon supersaturations of 1060% and 770% from the Southeast Pacific, we make use of the former values because our noble gas data in front of Ronne Ice Shelf support the supersaturation ratio of 0.757 (Fig. 2) reported by Schlosser (1986).

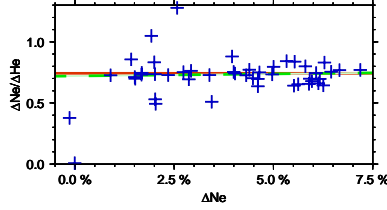


Figure 2: Ratio of neon/helium supersaturation vs. neon supersaturation observed in front of Ronne Ice Shelf during 1995-cruise *ANT XII/3*. Included are the regression line (green dashed) and the ratio $1060\%/1400\% \approx 0.757$ (red line) reported by Schlosser (1986).

To investigate the importance of ISW for the formation of deep and bottom water, a realistic tracer input was implemented to the regional coupled ice-ocean model BRIOS-1.0 (Beckmann et al., 1999). This paper describes the model setup, and the implementation of gas exchange at the atmosphere-ocean and ice shelf-ocean interfaces as this is a prerequisite for realistic tracer concentrations in an ocean circulation model. For model validation and the identification of the ISW source waters in the Weddell Sea, the paper compares observed and modeled temperature/salinity vs. helium distributions. Finally, the spreading path of ISW as part of the new bottom water formed in front of Ross- (RIS), Amery- (AIS), and Filchner-Ronne (FRIS) ice shelves is analyzed.

2 Model description

We use the regional coupled ice-ocean circulation model BRIOS-1.0, which is a modified version of the s-coordinate primitive equation model SPEM (Haidvogel et al., 1991; Beckmann et al., 1999). The model comprises the circumpolar Southern Ocean from 50°S to 82°S with a higher isotropic resolution of 1.5° in the Atlantic sector. The major subice shelf caverns are included but computational restrictions required a water column thickness greater than 200 m. More details of the physical model and its validation are provided in Beckmann et al. (1999), Timmermann et al. (2002), and Timmermann et al. (2002a). The ocean model is initialized using temperature and salinity data from the Hydrographic Atlas of the Southern Ocean (Olbers et al., 1992). At the ocean-atmosphere interface, it is driven with a climatological seasonal cycle derived from the stand-alone sea ice / mixed-layer model BRIOS-0, which was forced with the ECMWF (European Centre for Medium-Range Weather Forecasts) data of the period 1985–1993. The ice shelf–ocean interaction, however, is an integral part of the ocean model, thus computed at every time step.

Due to the lack of reliable gridded data, helium and neon concentrations are initialized with saturation concentrations of 100%. To avoid any supersaturation signal from the northern boundary to influence the results in front of the ice shelves, like FRIS, we restored the subsurface saturation at the northern boundary to 97% of the saturation

concentration related to the actual temperature and salinity. The timescale of restoring decreased northward from six to one day over the last six grid boxes. The comparison of integrated supersaturations in front of FRIS, derived from runs with restoring at the northern boundary to reduced saturation and equilibrium saturation, shows that the northern influence is negligible.

The model was integrated for 20 years to reach a quasi-stationary dynamical state (Beckmann et al., 1999). Additional 80 years of integration were performed with noble gas input at the ocean surface. The results were analyzed starting with year 30 since tracer release (year 50 of total model integration) to guarantee quasi-steady state tracer distributions in the area of our scientific interest.

2.1 Tracer boundary condition

The exchange of gaseous tracer at the atmosphere-ocean interface is implemented as a flux boundary condition according to Wanninkhof (1992), and Asher and Wanninkhof (1998). The flux of low-solubility, non-reactive gases depends on the concentration gradient and the wind velocity at the interface. Since the sea-ice cover, an important element of the polar climate system, reduces the gas exchange and supports the concentration difference, it is described according to England et al. (1994).

The gas exchange is completely described by:

$$F_G = k_0(\alpha_s p_{\text{atmos}} - c_w)(1 - A) \frac{u_{10 \text{ m}}^2}{\sqrt{\text{Sc}}}, \quad (1)$$

with k_0 the piston velocity¹, $\alpha_s \cdot p_{\text{atmos}}$ the solubility concentration after Weiss (1971) with p_{atmos} representing the atmospheric partial pressure² of helium/neon supposed to be constant in space and time, c_w the concentration at the ocean surface, and (Sc) the Schmidt number described by Jähne et al. (1987). The sea-ice cover A was derived from the sea ice / mixed-layer model BRIOS-0; its results agree well with satellite and ULS (Upper Looking Sonar) data (R. Timmermann, pers. comm.).

For convenience, the tracer distributions are presented as deviation from their saturation concentration, defined by:

$$\Delta X \stackrel{\text{def}}{=} \left(\frac{X_{\text{obs}} - X_0}{X_0} \right) \cdot 100, \quad (2)$$

with $X_{\text{obs}} = (\text{Ne}, {}^4\text{He})$ and X_0 the equilibrium values according to Weiss (1971).

This supersaturation, which depends on the temperature (Θ) and salinity (S), is an appropriate description because a pronounced supersaturation in the water column clearly identifies the ISW contribution. Such distinction cannot be done solely based on the concentration. However, some doubts exist about the accuracy of the solubility curves presented by Weiss (1971). Hamme and Emerson (2004) showed that these curves underestimate neon in cold sea water by 2% and Top et al. (1987) claim that the

¹Derived from bomb produced radiocarbon invasion rates into the ocean and hence k_0 includes implicitly the Schmidt number for ${}^{14}\text{C}$.

²We assumed a constant total atmospheric pressure of 992.94 hPa \approx 0.97996 atm.

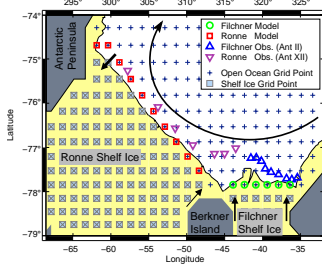


Figure 3: Locations of hydrographic stations and model grid points related to the model (dark-gray) and observed (yellow) landmask in the southwestern Weddell Sea. Note that the positions of ice shelf edge differ due to ice advance and retreat. Arrows schematically represent the main shelf circulation. For details see text.

helium curves are probably too low. Since we initialize the model with a equilibrium saturation, the surface flux³ is driven according to deviations from the saturation concentration, and the results are presented as relation to the saturation concentration, a slightly questionable solubility curve does not significantly affect our results.

2.2 Input by ice shelf basal melting

The input of helium and neon is assumed to be proportional to the ice shelf melt rate which is calculated at every time step as part of the coupled ice shelf-ocean model according to Hellmer and Olbers (1989). For melting zones the concentration in the uppermost ocean layer $c_{k=N}$ of the cavern is calculated by:

$$c_{k=N} = -c_0 \hat{S} \left(\frac{\frac{\partial h_{ice}}{\partial t} dt}{\Delta z_{k=N}} \right) \frac{\partial h_{ice}}{\partial t} < 0, \quad (3)$$

where c_0 is the saturation concentration, \hat{S} the supersaturation, $\Delta z_{k=N}$ the height of the upper most ocean grid box, $\partial h_{ice}/\partial t$ the melt rate, and dt the model's time step. This formulation also describes melting at the ice front as the vertical wall is approximated by an inclined interface due to the use of vertical s-coordinates. The influence of the formation of ice crystals within the cavern and their successive accumulation at the ice shelf base on the noble gas concentration was considered insignificant and thus neglected.

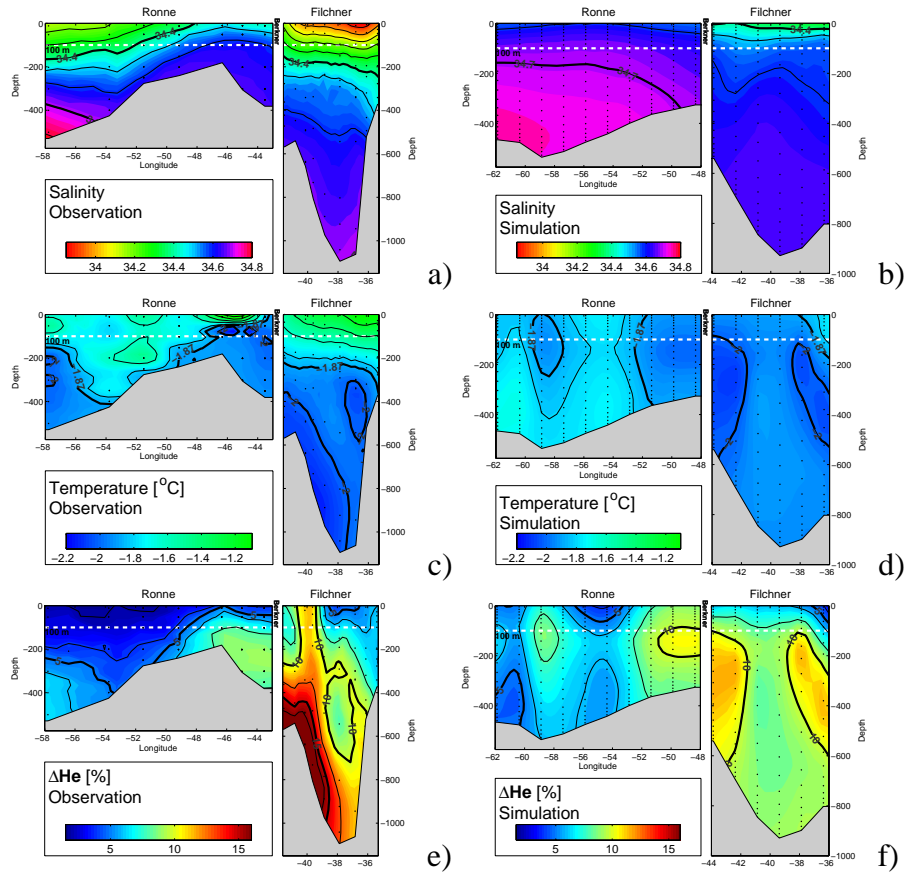


Figure 4: Cross sections of salinity (upper), potential temperature (middle), and helium supersaturation (lower) in front of Filchner Ronne Ice Shelf (Fig. 3). Left column (a, c, e) shows the observations up to 58°W while right column (b, d, f) represents the simulated distribution along the entire ice edge on 1 January of year 60 (since the release of the tracer). Black dots mark sample location or model grid point in the water column. Colorbars are identical for observations and simulations. Thick Contour-lines are: $S = 34.4, 34.7$, $\Theta = -1.87^{\circ}\text{C}, -2^{\circ}\text{C}$, and $\Delta\text{He} = 5\%, 10\%, 15\%$. The white dashed line indicates the cut-off depth at 100 m above which values are excluded from the analysis as explained in the text.

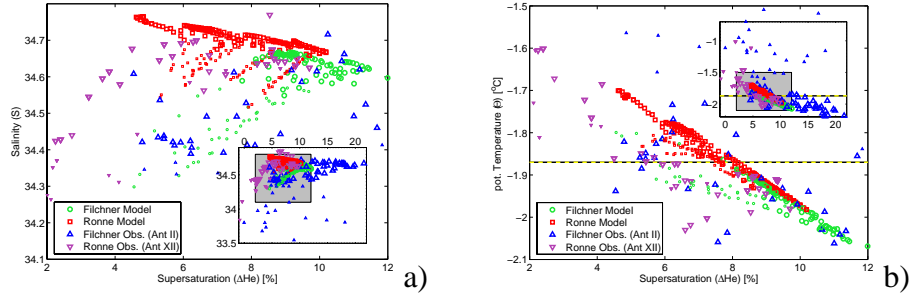


Figure 5: Salinity vs. helium supersaturation (left) and temperature vs. helium supersaturation (right) diagrams for the whole water column on the southern Weddell Sea continental shelf separated into the Filchner (east) and Ronne (west) regimes. The larger diagrams cover the simulated values while the insets show the entire range of the measurements with the gray square representing the range of the larger diagrams. According to the legend the measurements are represented by triangles while circles and squares represent simulated values of model year 60 (since the release of the tracer). Smaller symbols represent values of the uppermost 100 m, which are excluded from our analysis. The yellow line in b) marks the surface freezing temperature (T_f).

3 Model validation

The noble gases were sampled close to the Filchner Ronne Ice Shelf edge at that time (Fig. 3, observed shelf ice edge were taken from a GMT data set⁴). Due to advance and retreat inherent in the ice shelf movement, its position slightly differs from the edge position used in the model (Fig. 3) but has no influence on our comparison of the properties at the ice edge. The distribution of noble gases in front of Filchner Ice Shelf is presented in detail by Schlosser et al. (1990) while a description for the area off Ronne Ice Shelf is in preparation (O. Huhn, pers. comm.). The hydrographic conditions on the southern continental shelf to the east and west as described by Grosfeld et al. (2001) and Nicholls et al. (2003), respectively, can be summarized as follows: a general east to west advection of shelf waters along the ice shelf front in combination with a continuous salt gain due to sea ice formation causes high salinities near the Antarctic Peninsula (Fig. 4a, b). Here, the main flow underneath FRIS takes place feeding most of the cavern's ISW plumes (Nicholls and Østerhus, 2004).

The main ISW outflow occurs on the western slope of the Filchner Trough with observed helium supersaturations of up to $\Delta\text{He} \approx 20\%$ (Fig. 4c–f). The less supersaturated core on the eastern flank can be associated with the return flow as part of the cyclonic trough circulation (Nicholls and Østerhus, 2004). The observed ISW cores — areas of maximum ΔHe and minimum Θ — are deeper and modeled saturations are significantly lower both for reasons discussed below.

³The uncertainty in the parameterization of the flux dependence to wind speed has probably a larger effect.

⁴GMT (General Mapping Tool) topography is mainly based on the World Vector Shore data set of the National Geophysical Data Center in Boulder, Colorado, USA.

On the continental shelf west of Berkner Island a shallow ISW core was observed with supersaturations of $\Delta\text{He} = 9.7\%$ and $\Delta\text{Ne} = 7.1\%$, which are lower than the simulated values, $\Delta\text{He} = 10.2\%$ and $\Delta\text{Ne} = 7.7\%$. The model results indicate that this core is not influenced by the deeper core in the Filchner Trough but is fed by a direct outflow from the Ronne Ice Shelf cavern (compare Fig. 9). An additional but weaker ISW core was observed at the western Ronne Ice Shelf edge below 300 m. Although at the same longitude, the simulated core is shallower by about 100 m. Its higher supersaturation might be explained by the measurements not covering the entire Ronne Depression. The ISW cores are clearly visible in the sections of potential temperature but hardly traceable in the salinity distributions along the ice shelf edge (Fig. 4a–f). This is caused by the strong dependency of density on salinity for waters below the surface freezing point, forcing the ISW plume to stratify at the appropriate density/salinity horizon and thus loosing any dominant freshwater signal in the water column.

The amount of noble gases released and the strength of the supersaturation signal in the outflows both are proportional to the melting at the ice shelf base. Although the model is forced with a recurring annual cycle, basal melting beneath FRIS varies between $119 \text{ km}^3/\text{a}$ (cubic meter per year) and $143 \text{ km}^3/\text{a}$ around a mean value of $130 \text{ km}^3/\text{a}$ (Fig. 6). This mean is larger than the maximum net melt rate beneath FRIS of $118 \text{ km}^3/\text{a}$ derived from remote-sensing data (Joughin and Padman, 2003) but smaller than the $157 \text{ km}^3/\text{a}$ estimated from $\delta^{18}\text{O}$ in the ISW plume off Filchner (Schlosser et al., 1990); both values are based on an ice density of 917 kg/m^3 . The inter-annual variability (despite a constant annual surface forcing) results from the exchange between outflowing and inflowing water masses in front of the ice shelf (Nicholls, 1997) and the time a signal imprinted on the continental shelf needs to traverse the subice cavern. The amplitude of the annual melt cycle is smaller than the total variability of melting (Fig. 6). Therefore, we assume that the 50-year mean represents the noble gas distribution in front of FRIS fairly well, taking into account that the seasonal change of the hydrographic properties below 100 m affects the noble gas saturation in the meltwater plume by less than 0.1%.

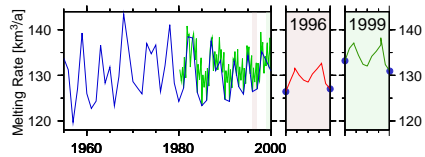


Figure 6: Melt rate for Filchner Ronne Ice Shelf for a 45-year period (left). The blue line connects values of 1 January of each year, the green line connects monthly means of the second half of this period. Two different arbitrary chosen annual melt rate cycles (right). The unit is km^3/year with $100 \text{ km}^3/\text{a} \approx 3 \text{ mSv}$.

In the following, we compare measured with modeled salinity/temperature vs. helium supersaturation in front of FRIS (Fig. 5). Due to the different oceanographic regimes on the southern Weddell Sea continental shelf, the results are separated for Filchner (east) and Ronne (west).

3.1 Salinity vs. helium supersaturation

The salinity–helium supersaturation values are clustered according to their origin (Fig. 5a) with higher salinities (higher helium supersaturation) representing the continental shelf in front of Ronne (Filchner) Ice Shelf. Considering the whole water column, three endmembers can be identified in the salinity-helium supersaturation (S - ΔHe) space:

1. low S and low ΔHe ,
2. high S and low ΔHe ,
3. low S and high ΔHe .

The first endmember represents the summer surface water masses of the upper 100 m which, freshened by sea ice melting, contribute to the stabilization of the water column. The observed salinity covers a wider range because of a high interannual variability of the sea ice cover, not captured by the model due to its time invariant surface forcing. Since the light surface water does not seem to mix with the ISW outflow (Fig. 5a), it will not be considered in the later discussion. The second endmember characterizes HSSW which results from deep convection driven by salt accumulation during sea ice formation. This process shifts the salinity (of endmember 1) to higher values but keeps the helium supersaturation low due to gas exchange across the ocean-atmosphere interface. The third endmember is the result of the continuous modification of HSSW at the ice shelf base (Fig. 1). The entrainment of glacial melt water decreases the salinity while the complete solution of the gases released by melting increases the helium supersaturation.

The modeled values along the FRIS edge qualitatively agree with the observations with the upper 100 m of the water column covering a wider range than further below. The salinity–helium supersaturation range, however, is smaller, and the helium maximum in front of Filchner Ice Shelf is half the observed value (see also Fig. 4e–f). This difference can be explained by the narrowness (15–50 km) of the ISW plume in the Filchner Trough (Schlosser et al., 1990) which cannot be captured in detail by a model resolution of ~ 24 km at the FRIS front. In contrast, the lower maximum supersaturation related to the broader ISW flow out of the Ronne cavern (west of Berkner Island) is better represented by the simulation.

3.2 Temperature vs. helium supersaturation

In front of Filchner Ice Shelf (Fig. 5b) observed temperature vs. helium relations below 100 m are grouped along a straight line with a minimum temperature of -2.19°C and a maximum supersaturation of $\Delta\text{He} = 21.5\%$ (Figs. 2 and 5 in Schlosser et al. (1990)). Such straight line is common for temperature vs. salinity distributions in front of ice shelves representing the melting of meteoric ice into sea water (Gade, 1979). Deviations from this line mainly represent water of the upper 100 m which includes a significant amount of warmer surface waters. Since the model resolution is coarse and it's driven with a smooth annual cycle, the observed relations scatter more around the regression line (Fig. 8). The observed correlations are furthermore

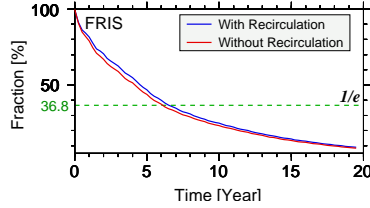


Figure 7: Temporal evolution of an artificial tracer concentration within FRIS. A water parcel is considered “recirculated” if it retains its initial signature after leaving and reentering the cavern. The residence time (e -folding time) is 6.5 or 6 years by considering or neglecting recirculation.

10%–50% lower than for the simulation (Tab. 1). In front of Ronne Ice Shelf the supersaturations are smaller with a maximum of $\Delta\text{He} = 9.70\%$ and temperatures are higher with a minimum of -2.03°C close to Berkner Island (Fig. 4d, 5b).

3.3 Calculation of properties of the melt-free component

The Foldvik-process considers newly formed bottom water as a mixture of helium and neon supersaturated ISW with nearly 100% ($\Delta\text{He}, \Delta\text{Ne} = 0$) saturated ambient water masses. In order to specify the ambient waters in the southern Weddell Sea we determine the temperature and salinity of the glacial melt-free component by calculating the regression line based on observed and modeled data points of the temperature/salinity–helium supersaturation spaces below 100 m. The line’s intercept with the ordinate ($\equiv 0\%$ supersaturation) provides the temperature/salinity of the ambient water. For the analysis we used the samples at the locations depicted in Fig. 3 and the temporal means of all open ocean grid points adjacent to the FRIS edge calculated from the data of 1 January of model years 30–80 (since the release of the tracer). The standard deviation for the temporal means is considered as uncertainty. For the data set of each year (years 30–80 since release of the tracer) we calculated a regression line. The envelopes of these regression lines represent the model’s uncertainty for the properties of the glacial melt-free component.

The simulated salinity/temperature vs. helium supersaturation values are located within the cluster of measured values (Figs. 8a, b). The maximum standard deviation for supersaturation, considering all model grid points for the year 30–80 period, of 1.06% exists in front of Filchner Ice Shelf at 218 m depth. For temperature and salinity, maximum standard deviations of 0.05°C and 0.019 occur at depths of 183 m and 163 m, respectively, in front of Ronne Ice Shelf. This spatial separation of the variability for different characteristics is evidence for the Ronne regime being dominated by shelf waters while Filchner is mainly influenced by the outflow of glacial meltwater. In addition, ocean/ice shelf interaction in the cavern seems to damp the variability of the inflow characteristics (temperature and salinity) and to convert the fluctuations to a variability of noble gas saturation in the outflowing meltwater plume.

Along the FRIS front the model tends to overestimate temperature and salinity of the glacial melt-free component penetrating onto or already on the continental shelf

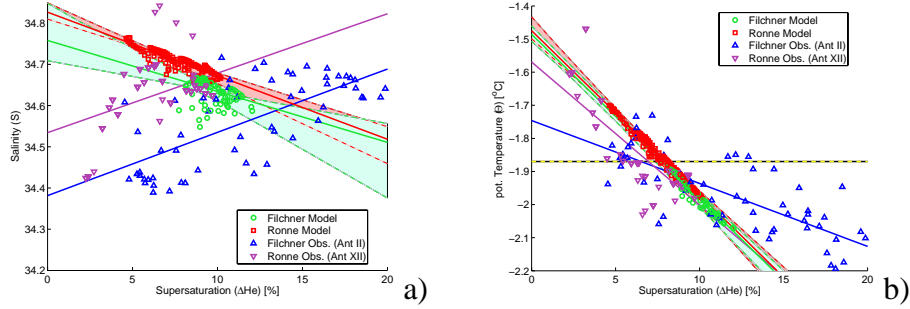


Figure 8: Salinity vs. helium supersaturation (left) and temperature vs. helium supersaturation (right) diagrams for the water column below 100 m on the southern Weddell Sea continental shelf separated into the Filchner (east) and Ronne (west) regimes. The x-coordinate was extended to zero supersaturation to determine the salinity and potential temperature of the melt-free component using a linear regression. Measurements are represented by triangles while circles and squares represent means of model years 30–80 since tracer release (see legend). The colors of the linear regression lines correspond to the colored data points in the legend. The yellow line in the right figure marks the surface freezing temperature T_f .

as shown in Table 1. This is mainly caused by the model resolution insufficient to capture the narrow cold ISW outflows and the topography at the continental shelf break which blocks warm open-ocean waters from access to the continental shelf, and by the forcing with the output of a stand-alone sea-ice model. The latter simulates a permanent sea-ice cover in front of FRIS reducing the salt release during sea-ice formation with consequences for deep convection and the cooling of the shelf water column.

On the other hand, Southern Ocean observations are affected by long-term variability. The measurements in front of Filchner Ice Shelf (*ANT III*, 1985) were conducted during a phase of the Antarctic Circumpolar Wave (ACW) with positive temperature and negative sea-ice coverage anomalies in the southwestern Weddell Sea related to weaker southerly winds (White and Peterson, 1996). In contrast, data off Ronne Ice Shelf (*ANT XII/3*, 1995) were collected during negative temperature and positive sea-ice coverage anomalies related to stronger southerly winds which enhanced the salt input and deep convection. The latter increases at depth the amount of water that recently was in contact with the atmosphere at temperatures equal to the surface freezing point (T_f), but decreases the relative amount of ISW with temperatures below surface freezing. Therefore, for a continental shelf influenced by ISW outflow enhanced convection may lead to a warming of the lower water column. Whether such mechanism caused the smaller temperature difference between model results and observations in front of Ronne Ice Shelf remains speculative.

The correlations for salinity and helium supersaturation show a non-uniform picture, being positive for the observations and negative for the simulation. In addition, both correlations for Ronne are higher than those for Filchner (Tab. 1). The latter is additional evidence for the regional separation between main inflow and main outflow for

Area	Intercept		Δ	Slope /100		Correlation Coefficient [%]	
	Observation	Simulation		Observation	Simulation	Observation	Simulation
<i>Salinity-supersaturation relation</i>							
Filchner	34.38 ± 0.05	34.76 ± 0.00	-0.38	1.5 ± 0.45	-1.24 ± 0.02	70	-43
Ronne	34.53 ± 0.04	34.83 ± 0.00	-0.3	1.4 ± 0.5	-1.542 ± 0.004	80	-90
All	34.57 ± 0.03	34.86 ± 0.03	-0.29	0.51 ± 0.32	-2.137 ± 0.005	23	-83
<i>Temperature-supersaturation relation</i>							
Filchner	-1.75 ± 0.06	-1.486 ± 0.001	-0.26	-1.90 ± 0.43	-4.954 ± 0.007	-76	-97
Ronne	-1.57 ± 0.05	-1.475 ± 0.001	-0.1	-4.3 ± 0.6	-4.962 ± 0.003	-46	-99
All	-1.70 ± 0.03	-1.465 ± 0.000	-0.24	-2.4 ± 0.3	-5.129 ± 0.003	-76	-99

Table 1: Characteristics of the glacial-melt free component as intercepts at zero supersaturation, and slopes of the linear regression through the pairs of temperature/salinity and helium supersaturation as depicted in Figs. 8a) and b) for all data (“All”) and separated for the “Filchner” and “Ronne” regimes. The uncertainties represent the 95% confidence interval, while Δ is the intercept difference with $\Delta = \text{observation} - \text{simulation}$.

the FRIS cavern. The correlation between temperature/salinity of the inflow (Ronne) and helium supersaturation of the outflow (Filchner) is also low (not shown). However, since results prior to model equilibrium indicated the subice transit of a long-term trend, we suggest a “low-pass” filtering within the cavern, supported by a multi-year residence time (Bayer and Schlosser, 1991; Mensch et al., 1996) in combination with mixing between water masses of different age (Fig. 7).

The addition of melt water, characterized by high supersaturations, reduces the salinity of the outflow. This combination is reproduced by the model even when the correlation is weak in front of Filchner (Tab. 1). The observations instead show that higher supersaturations are correlated with higher salinities. However, if we consider all data in front of FRIS, the correlation is very weak (Tab. 1). In contrast, the combination of cooling due to melting and higher supersaturations in the ISW plume is clearly detectable by observed and modeled Θ and ΔHe everywhere along the FRIS front (Fig. 4). One possibility for this discrepancy could be that the observations did not cover the saltiest shelf waters predominantly located in the westernmost corner of the southern continental shelf close to the Antarctic Peninsula. However further study needs to be done to understand the observed positive correlation between salinity and helium supersaturation.

4 Spreading of ISW

A meltwater plume carries a higher helium than neon supersaturation signal due to a lower solubility for helium. Far from the source the glacial-melt helium signal is, however, difficult to separate from a helium signal originating from other sources (Roether et al., 1998, 2001), like crust (crustal) and mantle (primordial) helium. Therefore, neon

is a more suitable tracer to pursue glacial melt in the ocean since the atmosphere/ocean exchange (Ozima and Podosek, 1983) and ice shelf melting (Schlosser, 1986) are the only sources. In the following, we analyze the spreading of water masses containing ISW in the Southern Ocean by simulating circumpolar helium and neon distributions. In addition, we discuss the importance of neon as an oceanic tracer.

The ISW spreading path, assuming an accuracy of the measurements of 0.2%, is shown in Fig. 9. The largest bottom patch with a significant supersaturation signal occurs in the southwestern Weddell Sea. Sources are the Brunt-Riiser-Larsen- (BRILIS), Filchner-Ronne, and Larsen (LIS) ice shelves (from east to west). The BRILIS signal spreads westward with the coastal current towards the sill of the Filchner Trough where it encounters the stronger supersaturation signal from FRIS. In agreement with the observations in front of Ronne Ice Shelf (Sec. 3) and from the southern Weddell Sea continental shelf (Gammelsrød et al., 1994; Foldvik et al., 1985; Mensch et al., 1996; Nicholls and Makinson, 1998; Nøst and Østerhus, 1998) an additional supersaturation signal appears west of Berkner Island (Fig. 9). Although in Ronne Basin close to the Antarctic Peninsula higher helium and neon supersaturations also exist, the signals are smaller than in the Filchner Trough (helium 6–7% vs. 9–10%, neon 4–5% vs. 7–8%). On the way to the northwestern Weddell Sea, as part of the cyclonic gyre circulation, Larsen Ice Shelf adds a tracer signal which is, however, not strong enough to overwhelm the supersaturation signal from FRIS. Helium and neon rich waters pass the northern tip of the peninsula to flow either westward in the southern Drake Passage as part of a slope current (Nowlin, Jr. and Zenk, 1988) or eastward within the northern limb of the Weddell Gyre (Fig. 9). In the Ross Sea, ISW rich water follows the circulation of the Ross Gyre (e.g., Assmann and Timmermann (2005)) as far as Cap Adare (western tip) where the signal splits into two branches (Fig. 9). One advects with the Ross Gyre to the east slowly eroding due to mixing with lower saturated waters. The other branch flows westward with the coastal current being detectable up to Prydz Bay in front of Amery Ice Shelf (AIS). The Amery signal is clearly distinguishable from the RIS signal as it dominates the supersaturation in Prydz Bay. Since it mixes also into the Antarctic Coastal Current, it can be traced as far as the Greenwich Meridian (eastern Weddell Sea) (Fig. 9).

The isotope laboratory at the University of Bremen (Institute for Environmental Physics) measures neon data with a precision (internal consistency) of 0.1%–0.3% (Well, 1995; Well and Roether, 2003). For a currently supposed detectable supersaturation of 0.5–1%, ISW can be only found in the bottom layer on the continental shelf in front and next to the major caverns (Fig. 9); i.e. northwest of Amery Ice Shelf, in a narrow band along the Ross Ice Shelf front, and in the Weddell Sea where the large supersaturated patch completely erodes as it reaches the northern tip of the Antarctic Peninsula. An enhanced detectability would improve the representation of ISW flows in the Southern Ocean especially for those originating from Ross and Amery ice shelves.

4.1 Melt water fraction along the ISW spreading path

The fraction of meltwater a_{ice} within the water column along the ISW spreading path can be calculated based on helium and/or neon concentrations assuming the water to

be formed by a two component mixture – one part supersaturated ISW and the other part ambient water saturated by 100%:

$$c_{\text{obs}} = a_{\text{ice}}\hat{S}c_0 + (1 - a_{\text{ice}})c_0$$

$$a_{\text{ice}} = \frac{1}{15} \left(\frac{c_{\text{He}}}{c_0} - 1 \right) \quad (4)$$

$$a_{\text{ice}} = \frac{1}{10.6} \left(\frac{c_{\text{Ne}}}{c_0} - 1 \right). \quad (5)$$

Here, c_0 is the saturation concentration, and c_{He} and c_{Ne} are the (observed) concentrations of helium and neon. The constants on the right hand side result from the saturation (\hat{S}) of helium and neon of 1500% and 1160% (100% saturation and supersaturation of pure glacial meltwater of 1400% and 1060%) (Schlosser, 1986) due to ice shelf melting. We expect that the saturation concentrations of the ambient water masses are identical, which is valid for water masses at nearly the same temperature. For the assumption of a two component mixture, a supersaturation in helium (neon) of 10% is equivalent to a meltwater fraction of 0.714% (0.943%) (see color bar in Fig. 9). High meltwater fractions of $a_{\text{ice}} \approx 0.7\%$ were observed in the water column directly in front of Filchner Ice Shelf (Schlosser et al., 1990). The fraction a_{ice} , however, reduces to 0.01% on its way from the source area to the South Orkney Islands. Enhanced turbulent mixing along the rough topography of the South Scotia Ridge continues to erode the meltwater signal such that an improved accuracy of the measurements might only have a limited effect on a better identification of ISW spreading within the northern limb of the Weddell Gyre.

5 Conclusions

For the first time a regional ocean circulation model, which includes the interaction between ice shelves and ocean, is used to determine the release of noble gases, helium and neon, due to ice shelf basal melting. Noble gases are ideal for tracing the meltwater embedded in Ice Shelf Water which contributes to the formation of bottom water in the Southern Ocean and, thus, to the ventilation of the global abyss. Although helium has a stronger supersaturation signal, we recommend neon to be used for this procedure due to its simple input function.

The comparison between modeled and observed temperature/salinity vs. helium supersaturation relations in front of FRIS shows that the model reproduces the characteristic features on the southwestern Weddell Sea continental shelf, i.e., increasing salinity from east to west, lowest temperatures combined with highest supersaturations in the Filchner Trough, and a cavern circulation with main inflow at the Ronne and main outflow at the Filchner ice fronts. The discrepancy between the observed and modeled characteristics of the glacial melt-free components might be caused in part by the fact that the observations were conducted at different phases of the ACW. However, model deficiencies with regard to resolution of bottom topography on the continental shelf and the forcing with the output of a stand-alone sea-ice model might have even a bigger impact on the model's deviation from the observations. Although forced with a

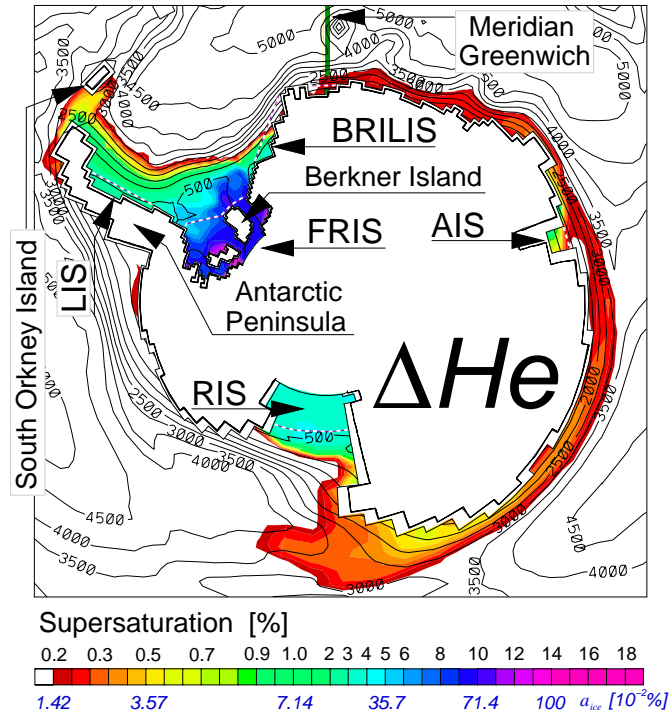


Figure 9: Modeled (year 50 since tracer release) supersaturations of helium in the bottom layer of the Southern Ocean together with the freshwater fraction (lower legend) from ice shelf melting deduced from the helium supersaturation (right) according to equation (4). The figures include the ice shelf fronts as represented in the model (white/magenta lines), bottom topography (black lines) with contour intervals of 500 m, and ice shelf names. FRIS: Filchner-Ronne Ice Shelf; BRILIS: Brunt-Riiser-Larsen Ice Shelf; AIS: Amery Ice Shelf; LIS: Larsen Ice Shelf; RIS: Ross Ice Shelf.

recurring annual cycle, the water mass characteristics on the continental shelf as well as the net basal mass loss fluctuate during the 60-year integration, caused by a feedback mechanism on time scales of several years between the ice shelf cavern and the continental shelf. The freshwater flux originating from FRIS has an impact on shelf water salinity and sea ice thickness (Hellmer, 2004) and, therefore, on the composition of the water masses feeding the meltwater plume from beneath the ice shelf.

Our model results support the general perception that ocean-ice shelf interaction has a significant influence on the characteristics of deep and bottom water. However, to provide a more comprehensive picture of the deep and bottom water formation in Antarctic marginal seas gas exchanges that occur in the open ocean also have to be considered. We therefore prepare an analysis of the results of a transient chlorofluorocarbon simulation (Rodehacke et al., 2005b).

In general, noble gas measurements are an excellent tool to study the cavern-continental shelf interaction because they are less susceptible to environmental changes which occur on interannual timescales. The current accuracy of measurements allows for the detection of ISW only close to the fronts of the major Antarctic ice shelves. Our model results indicate that with an enhanced detectability of 0.2% the spreading paths of water masses containing traces of glacial melt/ISW could be easier determined beyond the source areas. The meltwater signal from Filchner-Ronne Ice Shelf could be traced within the northern limb of the Weddell Gyre with access to the Scotia Sea. Traces of ISW originating from the Ross Ice Shelf would be detectable as far as Prydz Bay. Here the outflow from Amery Ice Shelf dominates the meltwater signal which advects with the coastal current towards the Weddell Sea with still a clear signal at the Greenwich Meridian. This occurrence provides additional evidence for a flow of freshly ventilated water masses into the Weddell Sea at 3000–4000 m depth (characterized by high CFC concentrations; Klatt et al. (2002)), influenced by the processes in front of Amery Ice Shelf as already suggested by the model results of Schodlok et al. (2001).

Our model approach neglects the influence of sea-ice formation on the saturation of noble gases in the mixed-layer. It is still debatable whether a reduction of the saturation due to sea-ice formation has an impact on the neon concentration in freshly ventilated water masses of the deep ocean. Hood (1997) suggests that sea ice formation erodes the neon supersaturation originating from ocean/ice shelf interaction, though Well and Roether (2003) argue that such strong neon reduction is not consistent with tracer observations in the Southern Ocean. Therefore, we further have to speculate about the processes which cause the observed slight supersaturation of neon in the abyssal ocean. One process might be the partial pressure change during the solution of noble gas in the mixed-layer and its subduction (Hamme and Emerson, 2002). Our model results suggest that the influence of ocean-ice shelf interaction on the neon supersaturation in newly formed deep and bottom water cannot be further neglected. However, models with higher resolution and improved physics regarding the parameterization of the ocean-sea ice interaction and vertical mixing, and an adequate representation of descending plumes at the continental slope are necessary to satisfactorily address the importance of the glacial meltwater for the water mass characteristics of the Southern Ocean.

Acknowledgement:

We thank W. Roether and his former group at University Bremen for providing the noble gas data and supporting the completion of the manuscript in various ways. Thanks also to three reviewers, whose invaluable criticism helped to clarify and improve the manuscript. We thank R. Timmermann and C. Lichey for providing model forcing fields. C. Rodehacke thanks B. Klein and the team of the Institute for (Tracer) Oceanography at University of Bremen for their contribution to the data processing. This project was funded by the Deutsche Forschungsgemeinschaft (DFG) Nr. Ro 318/43.

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