Water properties on the west Antarctic Peninsula continental shelf: a model study of effects of surface fluxes and sea ice

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Abstract

A vertical- and time-dependent numerical mixed-layer and sea-ice model is used to analyze processes responsible for sea-ice and surface mixed-layer water properties on the continental shelf on the west side of the Antarctic Peninsula. Atmospheric observations from Faraday and Palmer stations along with satellite sea-ice observations and shipboard water observations (four hydrographic cruises between January 1993 to February 1994) are used for forcing and verification. The focus of this study is the year 1993 during which the best observations exist. However, a 16-year simulation is completed to analyze interannual variations of ice thickness and mixed-layer depth. This model study shows that surface waters of the west Antarctic Peninsula are heated in the summer by solar radiation and cooled in the winter by sensible heat losses. Diffusive-convection is important for upward heat flux across the pycnocline. Ice melt in the spring is due to solar warming of open water, which then melts ice; the direct melting for ice by solar heating is negligible. The near closure of surface heat and salt budgets over 1 year supports the minor importance, or at least the compensation, of near-surface lateral exchanges. Intrusion of Upper Circumpolar Deep water from the Antarctic Circumpolar Current onto the subpycnocline shelf is a critical element of both salt and heat budgets. A 16-year simulation (1978–1993) reproduces most high and low ice years as observed by satellite microwave measurements, thus supporting the major contribution of thermodynamic (local) processes in creating sea-ice and mixed-layer properties.

1. Introduction

The atmosphere of the Southern Hemisphere affects strongly the waters of the Southern Ocean continental shelves through exchanges of heat, fresh water, and momentum. Many of these exchanges give rise to dense shelf waters that eventually become the bottom waters of the world ocean. In other regions, however, only the near-surface waters respond to atmospheric forcing and dense shelf water is not produced. The details of atmospheric exchange in one such region is analyzed in this paper.

The continental shelf waters on the western side of the Antarctic Peninsula (WAP, Fig. 1) participate in a vigorous exchange with the atmosphere, but do not become dense enough to convect to the bottom (Hofmann and Klinck, 1998a). The upper
100 m of water (the surface mixed layer) warm and cool in an annual cycle, while the deeper water maintains a relatively constant, oceanic character that derives from the Antarctic Circumpolar Current (ACC). Upper Circumpolar Deep Water (UCDW), which is carried northeastward along the shelf break by the ACC, episodically spills onto the shelf. Previous studies of the heat and salt changes on the shelf (Klinck, 1998; Smith et al., 1999) propose that heat from the warmer UCDW passes vertically into the near-surface layers warming the Winter Water (WW), which is the deep cold mixed layer from the previous winter. These vertical exchanges change the character of water in the upper few hundred meters of the ocean and thereby determine the type of water created in this region and seen down-stream (east) of the tip of the Antarctic Peninsula.

The DOVETAIL experiment (Muench and Hellmer, 2002) focused on the circulation of dense water from the Weddell Sea past the tip of the Antarctic Peninsula into the World Ocean. The ACC also flows through the DOVETAIL region (Matano et al., 2002) after exchanging water with the WAP. While the WAP shelf does not produce dense water to add to the outflow from the

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Fig. 1. Geography of the west Antarctic Peninsula. Black shading is land. The solid and dashed lines are the 500 and 1000 m isobaths, respectively. Initials indicate Marguerite (MaB) and Dallman (DaB) Bays; Adelaide (AdI), Renaud (ReI), Anvers (AnI), South Shetland (SSI) and Elephant (El) Islands and Gerlache Strait (GeS). Palmer (P) and Faraday (F) Stations are shown in white letters. Triangles show stations sampled during the March–May 1993 LTER cruise.
Weddell Sea, it does have the potential to modify water along the southern side of the ACC and thereby influence near-surface conditions in the DOVETAIL region.

The purpose of this paper is to use a model to quantify the physical processes that cause the observed hydrographic structure, mixed-layer dynamics and sea-ice processes over the WAP shelf. We focus on ocean exchange with the atmosphere, including freezing and melting of sea ice, in addition to processes associated with oceanic heat and salt fluxes. Although surface fluxes drive the seasonal changes, the water structure is controlled mainly by vertical diffusive fluxes and horizontal, sub-pycnocline exchange of water from ACC, specifically, UCDW onto the shelf. An earlier study (Klinck, 1998) used temperature and salinity changes from hydrographic observations to show the importance of heat exchange across the permanent pycnocline. Finally, the interannual variability of sea ice and mixed-layer depth can be explained, for the most part, by local processes including heat and salt exchange with the atmosphere and vertical exchange with water from the ACC that intrudes on the shelf. Although the calculations are specific to this shelf, they indicate how oceanic exchange onto shelves affects ice characteristics and water properties; processes that might be important in the near-surface in other parts of the Antarctic.

We calculate heat and salt fluxes with a vertical- and time-dependent mixed-layer and sea-ice model forced by observed atmospheric conditions. The numerical model combines a thermodynamic–dynamic ice model with the vertical form of the Princeton Ocean Model (POM), which uses a turbulence closure scheme to provide time dependent turbulent vertical mixing. This model has been shown to reproduce realistic mixed-layer dynamics in other regions (Blumberg and Mellor, 1987).

2. Data and methods

2.1. Oceanic and atmospheric observations

Between January 1993 and February 1994, portions of the WAP shelf (Fig. 1) were sampled during four cruises (details in Table 1). The fall (March–May) 1993 cruise provided the most extensive, quasi-synoptic hydrographic observation for the region to date. The other three cruises sampled near the center of the shelf. On all cruises, conductivity–temperature–depth (CTD) measurements were made with a station spacing of 20 km across-shelf using a SeaBird CTD mounted on a Bio-Optical Profiling System (BOPS) (Smith et al., 1984). Data were obtained to within 20 m of the bottom, or to 500 m in water deeper than 500 m. On the fall 1993 cruise, a SeaBird 911+ was used to obtain 10-km station spacing and samples below 500 m. Full details on data collection and processing for each cruise are given in Lascara et al. (1993); Smith et al. (1993a, b); Klinck et al. (1994). Data are available online at http://www.i-cess.ucsb.edu/lter/lter.html.

Atmospheric observations from Faraday Station, made by British Antarctic Survey (BAS) personnel (Jones and Limbert, 1987), span 1957 to 1993 and include dry- and wet-bulb air temperatures, cloud cover, wind speed, wind direction, relative humidity, and atmospheric pressure at sea level at 3-h intervals (Fig. 2). Surface radiation, including hourly records of total short-wave radiation, are available from BAS for the years 1985 through 1993. Data are archived at the University of Wisconsin (http://uwamrc.ssec.wisc.edu/aws/).

Table 1

<table>
<thead>
<tr>
<th>LTER Cruise ID</th>
<th>Season</th>
<th>Dates (dd-month yy)</th>
<th>Lines sampled</th>
</tr>
</thead>
<tbody>
<tr>
<td>93A</td>
<td>Summer</td>
<td>08-Jan 93 to 07-Feb 93</td>
<td>200–600</td>
</tr>
<tr>
<td>93B</td>
<td>Fall</td>
<td>25-Mar 93 to 15-May 93</td>
<td>000–900</td>
</tr>
<tr>
<td>93C</td>
<td>Winter</td>
<td>23-Aug 93 to 30-Sep 93</td>
<td>200–600</td>
</tr>
<tr>
<td>94A</td>
<td>Summer</td>
<td>11-Jan 94 to 07-Feb 94</td>
<td>300–600</td>
</tr>
</tbody>
</table>
Estimated daily sea-surface temperature (SST) for the waters surrounding Palmer Station for the years 1992–1994 was constructed from ship observations near Palmer Station (50 km radius) and an Automatic Weather Station (AWS) located at Palmer Station. A quadratic curve was fit to the spring, summer, and fall observations to produce a continuous daily record (Fig. 2E).
Winter SST is assumed to be freezing ($T_f = -1.83^\circ$C) when satellite observations indicate ice cover of at least 20%. The surface heat budget is not sensitive to the details of the ocean SST (details in Section 3.1).

A 16-year record of sea-ice concentrations for the west Antarctic Peninsula shelf was obtained from the National Snow and Ice Data Center (NSIDC) archives (Cavalieri et al., 1997). The NSIDC data combine three previous data sets that have been cross-calibrated at NASA Goddard Space Flight Center (GSFC) (Comiso et al., 1997). The total ice area was calculated from gridded ice observations for a region covering about 45 × 30 pixels (625 km$^2$ per pixel) over the WAP for each day. Additionally, ice fractional area was estimated within 100 km of Palmer Station (30 pixels or about 19,000 km$^2$) for comparison with the sea-ice model (Figs. 2F and 3).

### 2.2. Surface heat flux calculation

The total heat and salt budgets for surface waters of the WAP include fluxes through the atmosphere-ocean interface and the ice-ocean interface as well as fluxes through the permanent pycnocline. A preliminary surface heat budget assumes negligible advective and through-ice heat flux compared to the open-water flux (justified later).

Terms in the heat budget are calculated with commonly used bulk formulas (e.g., Parkinson and Washington, 1979), but some details are given below. The heat flux through open water ($Q_{\text{open}}$) is estimated as

$$Q_{\text{open}} = (1 - A_i)(1 - \alpha_s)Q_{\text{sw}} + \epsilon(Q_{\text{lw}}^{\text{up}} - Q_{\text{lw}}^{\text{down}}) + Q_{\text{sens}} + Q_{\text{lat}},$$

(1)

Fig. 3. (A) Air temperature from Faraday Station. (B) Sea ice area fraction near Palmer Station extracted from SMMR/SSMI observations. Vertical dotted lines are January 1 of each year.
where $Q_{sw}$, $Q_{lw}$, $Q_{sens}$ and $Q_{lat}$ represent heat flux due to short-wave radiation, long-wave radiation, sensible heat and latent heat, respectively. $a_i$ is the surface albedo, and $A_i$ is the fraction of the area that is ice covered. All constants not given in the text are in Table 2. The short-wave radiation ($Q_{sw}$) is estimated from the geometric model of Zillerman (1972) with the cloud correction by Laevastu (1960). This calculation was validated against measurement at Faradays (Smith, 1999). The upward long-wave radiation ($Q_{lw}^{up}$) is calculated from the Stefan–Boltzmann relationship

$$Q_{lw}^{up} = \sigma (SST + 273)^4,$$

where $\sigma$ is the Stefan–Boltzmann constant, and SST is sea-surface temperature from Palmer Station. Incoming long-wave radiation ($Q_{lw}^{down}$) is estimated from air temperature, $T_{air}$ (°C) and cloud cover, $C_i$ (Maykut and Church, 1973).

$$Q_{lw}^{down} = \sigma (T_{air} + 273)^4(0.7855 + 0.2232 C_i^{0.75}).$$

(3)

Sensible heat exchange ($Q_{sens}$) is estimated from

$$Q_{sens} = \rho_a C_{pa} C_{HO} W (SST - T_{air}),$$

(4)

where $C_{pa}$, $\rho_a$ and $C_{HO}$ are defined in Table 2, and $W$ is the wind speed at 10 m (m s$^{-1}$). Latent heat exchange ($Q_{lat}$) is calculated from

$$Q_{lat} = \rho_a L_v C_{HO} W (q_o - q_A),$$

(5)

where $q_A$ and $q_o$ are the measured specific and the saturated specific humidities, respectively.

### 2.3. Thermodynamic ice-ocean model

A thermodynamic ice-ocean model is used to examine the role of heat and salt fluxes on the seasonal evolution of the mixed layer. The ocean model is based on the vertical portion of the POM (Mellor, 1993; Blumberg and Mellor, 1987), which includes prognostic equations for current velocity, temperature, and salinity as a function of depth and time. The Mellor–Yamada level 2.5 turbulence closure scheme (MY2.5) (Mellor and Yamada, 1974, 1982) provides an estimate of time and vertically dependent vertical diffusivities for momentum and scalar properties ($K_m$ and $K_h$), respectively. To these coefficients are added two other estimated diffusivities as described next.

Turbulence closure schemes are unable to represent realistically mixing in highly stratified conditions (Large et al., 1994; Kantha and Clayson, 1994), which is a significant problem for melting ice and results in an unrealistic, very thin (1 grid interval) layer of fresh water that then warms with surface heating, and stabilizes the top of the water column. Observations indicate a 5- to 10-m thick seasonal mixed layer under melting conditions. A mixing scheme based on the gradient Richardson Number (Kantha and Clayson, 1994) is used to estimate a diffusivity at each depth under highly stratified conditions; this value is added to the other coefficients.

The differential transfer of heat and salt through the permanent pycnocline by double diffusion may be important along the west Antarctic Peninsula shelf (Smith et al., 1999). The density ratio ($R_p$) indicates the importance of the double diffusive instability.

$$R_p = \frac{\beta S}{\alpha T_0},$$

(6)
where $T_z$ ($^\circ C \text{ m}^{-1}$) and $S_z$ (psu $\text{ m}^{-1}$) are the large-scale vertical gradients of temperature and salinity, respectively, and $\alpha$ ($^\circ C^{-1}$) and $\beta$ (psu$^{-1}$) are the thermal expansion and haline contraction coefficients. Values of $R_p > 1$ are favorable to the diffusive–convective instability (Kelley, 1984). To estimate the density ratio on the WAP shelf, a typical value for $\beta/\alpha$ is 16 and typical temperature and salinity differences across the 100-m thick pycnocline are 1.3$^\circ$C (0.013$^\circ$C m$^{-1}$) and 0.2 (0.002 psu m$^{-1}$), respectively; the resulting ratio is 2.5. Detailed calculations from average profiles of temperature and salinity for each cruise result in a larger density ratio of 3 (winter) and 4 (fall and summer) from 100 to 350 m (figures not shown).

An example CTD cast from the fall 1993 cruise (Fig. 4a) is used to calculate the density ratio and diffusive–convective heat flux (Fig. 4b) as well as the diffusivities (Fig. 4c). Note that the heat flux is typically $3 \text{ W m}^{-2}$ with peaks up to $8 \text{ W m}^{-2}$. The average temperature and salinity profile for each of the four cruises (Fig. 5) shows that the example station is typical. Fedorov (1988) developed an empirical formula for layer thickness as a function of stratification and the Prandtl number; conditions on the WAP give 5.7 m as the double diffusive layer thickness. Observations from the four cruises discussed here indicate 3–10 m layer thicknesses, when layered convection is active.

Diffusive–convective heat flux is included in the model by increasing vertical diffusivities by parameters $K^{T}_{dd}$ and $K^{S}_{dd}$, which are calculated from estimated diffusive–convective heat flux. Various models exist to calculate the flux. A review of the double diffusive process is presented in Robertson et al. (1995). We use the following formulation from Large et al. (1994) which is based on Marmorino and Caldwell (1976) and Fedorov (1988).}

\[
Q_{dd} = 0.00859 \rho_0 C_{po} \alpha^{-1} (gk_T^2 \gamma^{-1})^{1/3} \\
\times \exp\{4.6 \exp[-0.54(R_p - 1)](\alpha \Delta T)^{4/3}\},
\]

\[\text{(7)}\]

Fig. 4. Diffusive–convective heat flux and diffusion coefficients calculated from a station in the middle of the shelf (600.140) during the March–May 1993 cruise. (A) Salinity (solid line) and temperature (dotted line). (B) Diffusive–convective heat flux (solid line) and density ratio (dotted line) calculated from station profiles. (C) Estimated vertical diffusion coefficients for heat (solid line) and salt (dotted line) due to diffusive–convective fluxes from the station profiles (after Fedorov, 1988).
where parameters are defined in Table 2. The temperature difference ($\Delta T$) is calculated over adjacent vertical grid points (2.6 m spacing).

This heat flux is converted to a diffusivity by

$$K_{dd}^T = \frac{Q_{dd}}{\left(\rho_o C_{po} \frac{\partial T}{\partial z}\right)}.$$  

(8)

The salt diffusivity is found from heat diffusivity by (Large et al., 1994)

$$K_{dd}^S = \begin{cases} (1.85 - 0.85 R_p) K_{dd}^T R_p^{-1} & \text{when } 1 > R_p > 2, \\ 0.15 K_{dd}^T R_p^{-1} & \text{when } R_p > 2. \end{cases}$$

(9)

Typical values on the WAP shelf result in a diffusive–convective heat flux of about 5 W m$^{-2}$, which requires a diffusivity ($K_{dd}^h$) of $15 \times 10^{-5}$ m$^2$ s$^{-1}$. A typical coefficient for salinity is much smaller (i.e. $K_{dd}^h \approx 8 \times 10^{-6}$ m$^2$ s$^{-1}$ for $R_p = 2.5$), which is characteristic of double diffusive fluxes (Fig. 4b). Note that these diffusivities are at least a factor of 10 larger than the background diffusivities described next.

Minimum vertical diffusivities are specified ($10^{-5}$ and $10^{-6}$ m$^2$ s$^{-1}$ for momentum and scalars, respectively) to provide a small amount of smoothing for the numerical solutions. The effect of these weak, background diffusivities was determined from average temperature and salinity profiles for each cruise from which diffusive fluxes were estimated (figures not shown). The largest heat flux was at the base of the pycnocline (about 100 m) with values ranging from 1 (summer) to 3 (winter) W m$^{-2}$. Salt fluxes were largest above 100 m with values ranging from 0.012 to 0.017 mg salt m$^{-2}$ s$^{-1}$. These background fluxes are 1 to 2 orders of magnitude smaller than those fluxes estimated here or in Klinck (1998).

Water properties below 200 m on the WAP shelf are maintained by lateral intrusion of UCDW onto the shelf from the ACC (Klinck, 1998; Smith et al., 1999). The vertical model represents this process by nudging the deep temperature and salinity to the average of all hydrographic observations on the shelf during the March–May 1993 cruise. This average covers the widest area of all the cruises. The average deep structure from all observations is the same as that from the March–May cruise since the deep water variability on this shelf is small (Fig. 5). The nudging time scale below 300 m is the model time step, which ensures that the deep properties are maintained. The nudging parameter is smoothly reduced to zero from 300 to 200 m. The heat and salt flux due to
nudging is calculated to estimate the lateral flux across the shelf break.

2.4. Ice model

The sea-ice model is based on four papers (Mellor and Kantha, 1989; Kantha and Mellor, 1989; Håkkinen and Mellor, 1990, 1992) that are referred to collectively as MKH. Only details that differ from these papers will be presented here.

The ice model has four prognostic variables: ice thickness \( h_i \), local ice area fraction \( A_i \), and the horizontal components of ice velocity \( U_i \) and \( V_i \). Changes in the ice thickness result from growth/melt rates at the air–ice, air–ocean, and ice–ocean interfaces, which are produced by imbalances in the local heat fluxes. The freezing point is a linear function of salinity \( T_f = -0.05435 \times S \).

Thermodynamics determines how the volume of ice \((A_i h_i)\) changes at each time step. An independent equation, taken from Håkkinen and Mellor (1990), is used to calculate the ice fractional area \( A_i \). Two empirical parameters control how area coverage changes as ice freezes or melts. Freezing increases both ice volume and area fraction. Melting preferentially reduces area fraction. We use the same values for the empirical constants as in Håkkinen and Mellor (1990) except we choose the flow divergence to be zero. We limit the fractional area to be no more than 85% to maintain leads that are common in this area (Fig. 3B).

The primary difference between the model used here and MKH is the treatment of thermodynamics within the sea ice. MKH includes one internal ice temperature point (i.e. Semtner 1-level ice model) while the model presented here uses the Semtner 0-level ice model (Semtner, 1976), which treats the ice as a uniform layer with thickness \( h_i \) and neglects internal ice thermodynamics. The temperature distribution within the ice is uniform for relatively thin ice (thicknesses of 0.5–1 m at winter maximum), which is the case along the western Antarctic Peninsula.

The model is forced by short-wave radiation, air temperature, wind speed (to provide energy for mixing), and precipitation. At Palmer Station, precipitation is about the same in each month, so the annual average precipitation divided by the total time steps in a year is added to the model at every time step. If ice exists, then \( A_i \) partitions the precipitation between ice and ocean. Over ice, the model accumulates snow; otherwise, precipitation enters the ocean as water.

2.5. Numerical details

The ocean model is integrated in time using an implicit scheme (Mellor, 1993; Blumberg and Mellor, 1987) to avoid constraints imposed by small vertical spacing and large vertical diffusivities. The ice equations are solved with a forward in time (Euler) scheme. To avoid a singularity, the ice thickness tends to a small number \((10^{-3} \text{ m})\) and \( A_i \) vanishes under conditions of no ice cover. The time step \((\Delta t)\) for all model simulations is 900 s. The model depth is 400 m and is resolved with 150 levels, giving a vertical grid spacing of 2.6 m, which resolves the observed vertical structure near the surface.

3. Results

The easiest calculation uses only observed atmospheric conditions, observed ice cover, and bulk formulas to estimate fluxes through the ice-free surface. This necessitates ignoring heat flux through the ice and ocean advective fluxes. If this initial heat budget has a small annual average heat flux, then the ignored processes either compensate or are small. If the budget does not integrate to zero, then a more detailed calculation is required that represents sea-ice and ocean processes. The next section presents the simple heat budget, which has a non-negligible annual mean flux. Section 3.2 discusses the model based results.

3.1. Surface heat budget

Atmospheric conditions are used with the bulk aerodynamic formulas to calculate a surface heat budget (Fig. 6), which accounts for average ice cover. In this budget, heat loss through ice in winter is set to zero (later calculation show that it is small). The dominant terms in the heat budget
are short-wave solar radiation (heating) during the summer and sensible heat loss (cooling) in the winter. The total budget and the individual terms averaged over the three 1993 cruises (Table 3) agree with this result.

In the summer and fall months (days 300–365 and 0–100), short-wave radiation is the dominant source of heat (Fig. 6A) with values of 175 W m$^{-2}$, although maximum heating can be up to 300 W m$^{-2}$. Net long-wave radiation is a persistent oceanic loss throughout the year, with an annual average of 30 W m$^{-2}$ and a maximum of 80 W m$^{-2}$. Other terms are negligible, resulting in net heating of 100 W m$^{-2}$.
During the winter months (days 100–250), the short-wave flux diminishes to zero, while the net long-wave loss persists. The latent heat flux, due to evaporation, is small throughout the year but becomes a modest loss of 20 W m\(^{-2}\) in the late fall because higher winds favor evaporation. Sensible heat loss (about 50 W m\(^{-2}\)) dominates, but is larger during events (up to 150 W m\(^{-2}\)), which are due to modest winds from the south bringing in much colder air (e.g., Fig. 2A and B, day 180). Note, however, that even though winter is a time of net cooling, warming events are possible due to air temperatures above \(-2^\circ C\) (Fig. 6C, e.g., days 135 and 210–220).

Estimated sea-surface temperature (SST) was used in the above calculation (a curve fit to sporadic measurements, Fig. 2E), but due to a small surface temperature range, it has little effect on the final result. The 1-year heat budget was repeated with the SST set to a constant 0°C; the resulting change was about 10% (Smith, 1999). The largest difference occurs in winter when sensible heat loss is reduced by the higher SST. Other fixed values for SST between 2°C and \(-2^\circ C\) indicated a weak effect of SST on the heat budget.

The time-integrated heat budget (Fig. 7) shows a net cooling over 1993 of about 2 \times 10^8 \text{Jm}^{-2}, which is equivalent to a persistent annual loss of 6 W m\(^{-2}\). Heat transfer by double diffusion across the permanent pycnocline can be this large (Fig. 4B) so the heat budget is recalculated with a persistent warming of 5 W m\(^{-2}\) due to heat from below. This heat flux causes a net warming, but it illustrates the effect, and potential importance, of heat flux to the surface layer from below. Finally, heat does escape to the atmosphere though the ice in winter, so a third calculation added 10 W m\(^{-2}\) cooling through the ice during times of ice cover. With this extra cooling, the annual heat budget nearly balances.

The surface heat budget nearly closes over 1 year. Heating across the pycnocline and the cooling rate through ice are small but make minor, but important, contributions. Therefore, it is necessary to pursue a more complete analysis that includes oceanic and ice processes.

### 3.2. Results from a thermodynamic sea-ice, mixed-layer model

The coupled, dynamic–thermodynamic, sea-ice, mixed-layer model is initialized with the observed temperature and salinity from August 1993, an idealized area-averaged ice thickness of \(h_i = 0.5\) m and run for 8 years. The 1993 atmospheric observations from Faraday were repeated to evaluate model spin-up and stability.

The simulated temperature and salinity of the mixed layer reproduced several key features of the ice-ocean system along the west Antarctic Peninsula (Fig. 8), including the timing of the ice cycle (i.e. timing of ice advance/melt and maximum extent), mixed-layer depth (defined in the model as the depth at which the turbulent kinetic energy crosses \(5 \times 10^{-7} \text{m}^2 \text{s}^{-2}\)), and the depth of the permanent pycnocline. The simulated distribution of ice thickness adjusts from its initial winter condition (\(h_i = 0.5\) m) to 0.3 m by the second winter, developing a consistent seasonal cycle (maximum winter thickness of 0.35 m). For the remainder of the simulation, maximum ice thickness increases by less than 1 cm yr\(^{-1}\) (Fig. 8A).

The seasonal temperature variation develops in the first model year (Fig. 8B), although there is a slight deepening of the winter mixed layer to about
160 m over the first three model years. After the fourth year, the deepening rate is less than 1 m yr$^{-1}$.

The seasonal salinity structure takes longer (about 4 years) to stabilize (Fig. 8C). After model year 4, there is little deepening of the salinity structure, most evident in the 33.9 isohaline (heavy solid line on Fig. 8C). The summer freshening and the winter salt rejection are evident in the annual excursion of the mixed-layer depth (heavy dashed line on Fig. 8C).

A 1-year record (Fig. 9) is extracted from the 8-year simulation (starting in the middle of model year 5, 1 January) for comparison with 1993 hydrography from four cruises (Figs. 5 and 10) spanning January 1993 to February 1994 (Klinck, 1998; Smith et al., 1999) and ice conditions from SMMR/SSMI for 1993 (Fig. 11).

![Figure 7: Surface heat budget for 1993 integrated over the year (solid line). The cumulative heat budget with net heating from below of $5 \text{ W m}^{-2}$ (dashed line). The cumulative heat budget with both net heating from below of $5 \text{ W m}^{-2}$ and net loss through ice of $-10 \text{ W m}^{-2}$ (dotted line).](image-url)

The winter (days 200–300) model surface layer is approximately 140 m thick, with temperature at freezing ($T = -1.83 \degree C$) and salinity of 33.8, respectively (Fig. 9B and C). These well-mixed conditions persist when ice cover is present. Beginning around day 300, surface heating warms the water, which melts the ice resulting in a very small surface temperature change. After the ice melts, heating warms the surface waters. The mixed-layer shoals over a month and a relatively warm surface layer develops over the, now isolated, remainder of the winter mixed layer. The WW below 50 m continues to warm both from above and below. In the fall (beginning on day 435), the surface layer erodes over a span of two months due to surface cooling and increased winds (deeper mixing), eventually exposing the deep, cold WW layer.
The annual cycle of salinity involves freshening in the spring with the melt of the ice cover, and the release of all the snow that fell on the ice over the winter (Fig. 9C). Slight freshening occurs over the summer due to precipitation, which counter-balances a small upward salt flux. The simulated summer surface layer is between 40 and 80 m thick, with temperature and salinity...
of 2°C and 33.65, respectively. There is a progressive deepening of the surface layer in the fall and an increase in mixed-layer salinity as ice starts to form.

Compared to observations, the simulated summer surface layer thickness is about right (50 m, Fig. 10A and B) but temperature is about 1°C too warm (Fig. 5A). The simulated winter mixed layer
extends to about 150 m (Fig. 10), which is 1.5 to 2 times the observed depth of 80–100 m (Fig. 5A and 10D). The WW layer in summer is colder than observed by about 0.5°C. Water properties below 250 m match, but are forced to do so by the strong nudging.

The simulated summer surface-layer salinity is below 33.8, in agreement with observations (Fig. 5B), although the model salinity at the surface is a few tenths higher than observed and lacks the linearly decreasing salinity near the surface. The winter surface salinity increases to 33.9, in agreement with observations. The simulated winter mixed-layer thickness is greater than the observed 100 m (Figs. 5B and 10D). The penetration of the simulated mixed layer to the top of the permanent pycnocline in winter is in agreement with observations.

The excess winter mixed-layer depth (Fig. 10) could be due to excess mixing, likely caused by the
large ice velocities (of order 0.5 m s$^{-1}$), which is common for horizontally unbounded one point ice models (Ikeda, 1989). Since the winter mixed layer controls the depth of the permanent pycnocline, this results in the top of the pycnocline being deeper than observed. The high simulated summer surface temperatures are due to insufficient vertical mixing in the surface layer once the stratification develops in the spring. This effect causes the heat to be confined to the near surface, creating warmer surface temperatures and colder Winter Water temperatures (near 100 m).

The simulated ice cover lasts for 165 days (5.5 months) of the 365 day simulation (Fig. 9A). The ice grows rapidly over a few weeks beginning around day 150 (early May). After the initial growth, the ice alternately advances and retreats but reaches a relatively stable thickness. The ice area remains near the maximum (0.85), with two large declines to 0.6 during times of warm air and ice melt. Around day 290 (mid-August), a rapid spring melt begins that lasts about 3 weeks.

The simulated ice cover is compared to the SMMR/SSMI area observations for the region near Palmer Station (Fig. 11A). The model area cover develops at the time shown in the observations but grows very rapidly to 0.85. The ice thickness has a slower development but peaks at about the same time as the observed area. Two large (days 180, 200) and two small (days 240, 270) events in ice area match between observations and model, but the model area changes more than indicated by the observations. Model thickness increases at times that observations indicate increased areal coverage. The model area declines very rapidly at day 300, with a similar decline in model thickness. Observations have a late spring expansion of ice area (days 320–340), which is not included in the model.

The model solution allows analysis of ice processes (freeze and melt). During the rapid, initial freezing of ice (days 150–170, Fig. 11), ice grows at the ice–ocean and the atmosphere–ocean interfaces, which combine to produce a total growth rate of $7 \times 10^{-7}$ m s$^{-1}$ (5 cm d$^{-1}$). After day 170, the bottom of the ice begins to melt due to the heating from below and the diminished heat conduction through the ice. The period just after the initial freeze is a time of relatively warm air temperature (Fig. 2B). Open-water freezing becomes small, except during events with durations of about a week, due to cold air temperature.

Throughout the 1993 simulation, the principal ice balance is between melting at the ice–ocean interface and growth at the atmosphere–ocean interface (Fig. 11B). Surface melting is significant only for a short time during spring (after day 290). Mid-winter ice retreat is caused by insufficient freezing (due to warm air temperatures) to counter the persistent melting at the ice–ocean interface. Ice growth is episodic and driven by high freezing rates ($1–5 \times 10^{-7}$ m s$^{-1}$ or 1–4 cm d$^{-1}$) while melting at the ice–ocean interface is small but persistent. The spring melt begins around day 290 (mid-October) with rates of $3–8 \times 10^{-7}$ m s$^{-1}$, which is sufficient to remove the ice cover within 20 days.

The simulated net heat budget (Fig. 12A) is similar to that estimated (Fig. 6), with strong cooling in winter and warming in summer. The
The winter surface ocean looses heat to sea ice at a rate of 5–10 W m\(^{-2}\) (Fig. 12B). Heat flux to ice in spring is about 100 W m\(^{-2}\), which occurs by solar warming of open water, which increases melt at the base of the ice.

Positive salt flux to the surface water (Fig. 12C) occurs only with ice cover. Over a year, the salt flux due to ice processes is nearly closed (integrates to zero). The additional salt flux to the surface layers due to diffusion across the permanent
pycnocline is balanced by the annual average precipitation (about 0.5 m yr$^{-1}$).

Mixing in the MY2.5 turbulence closure scheme is driven by brine rejection due to freezing in winter and surface winds in spring (Smith, 1999). Mixing associated with diffusive-convection occurs throughout the year, providing a heat flux of about 5 W m$^{-2}$ at the permanent pycnocline (Fig. 4B).

In this vertical-only model, deep water would cool and freshen over a multi-year simulation due to heat and salt transport upward to the surface layer (figures not shown). On the WAP shelf, sub-pycnocline water properties are maintained by episodic intrusions of oceanic water (UCDW) that is warmer and slightly saltier than shelf water. Mass is conserved below the pycnocline so onshore intrusions are balanced by expulsion of sub-pycnocline water elsewhere on the shelf. This influence of oceanic water is included in the model as nudging at depth to specified temperature and salinity, which appears in the model as a deep source of heat and salt. Typical heat flux due to nudging is 8–12 W m$^{-2}$ while salt flux is 0.5–0.6 mg salt m$^{-2}$ s$^{-1}$; these fluxes vary little over the 1 year of simulation being analyzed (Smith, 1999).

3.2.1. Sensitivity to sub-pycnocline temperature

The sub-pycnocline temperature determines the heat flux across the pycnocline. The effect of different choices is analyzed with a series of experiments, with the deep temperature shifted by $-0.75^\circ$C, $-0.50^\circ$C, $-0.25^\circ$C, $0.25^\circ$C, $0.50^\circ$C and $0.75^\circ$C. These temperatures range from colder than any temperature observed on this shelf (Hofmann et al., 1996) to that of pure oceanic UCDW (Smith et al., 1999). Each experiment was run exactly as the previously described case, and model year 5 was analyzed. Increasing the deep temperature reduces ice thickness by as much as half for the warmest temperatures (Smith, 1999); decreasing the temperature produces somewhat thicker ice (3–5 cm thicker). Thicker ice, due to colder deep temperatures, produces thinner mixed layers. The depth of the permanent pycnocline is unaffected by these deep temperature changes. The observed salinity range is very small, so sensitivity to deep salinity is not analyzed.

3.2.2. Sensitivity to diffusive–convective flux

Formulas and coefficients for diffusive–convective heat and salt fluxes vary; sensitivity is examined by changing these diffusive coefficients by factors of 0.0, 0.5, 1.5, 2.0, and 3.0. All other model processes, parameters, and analyses are unchanged. Removing diffusive-convective increases average winter ice thickness by 50%. Doubling coefficients causes nearly 50% reduction in ice thickness. Tripling coefficients produces thin ice to the point that ice-free conditions occur in mid-winter (Smith, 1999). A final test removes the differential flux of heat and salt by setting the double diffusion coefficient to a constant $5 \times 10^{-5}$ m$^2$ s$^{-1}$ for both heat and salt, which is larger than the background values. This choice produces reasonable mixed-layer temperatures, but excessive winter salinities, which argues for different heat and salt diffusion rates, as occur with diffusive-convective, to produce realistic surface conditions.

3.3. A 16-year simulation

Encouraged by the similarity of the model to conditions in 1993, we ran a 16-year simulation for the years 1978–1994 during which atmospheric conditions from Faraday Station and sea-ice observations from SMMR/SSMI are available.

The modeled ice thickness (Fig. 13) reproduces many of the high and low ice years, defined by departure from the 16-year average (Smith et al., 1996; Stammerjohn and Smith, 1996) observed in the GSFC SMMR/SSMI ice field (Fig. 3B). In particular, the model reproduced the record low and high ice years in 1989 and 1987 (Jacobs and Comiso, 1993). It also produced high ice in 1980, 1981 and 1982 and low ice in 1983, 1984 and 1985. However, the very large ice area for 1980 was not produced by the model (Figs. 3B and 13B).

The mixed-layer depth is shallow (10 to 30 m) in the spring and deep (100–150 m) over the winter (Fig. 13B). Winter mixed layers are deeper during high ice years. The permanent pycnocline is nearly
50–70 m shallower in low ice years and deepens in high ice years.

4. Discussion

The warm, salty character of the sub-pycnocline water on the west Antarctic Peninsula shelf is distinct from other Antarctic shelves (Hofmann and Klinck, 1998b) because the ACC flows along the shelf break. While atmospheric and solar processes dominate the heat budget of the surface layer, contributions of heat and salt from the UCDW play an important role. Specifically, the thickness of winter ice is affected by upward heat flux from below. Because this flux is due to diffusive-convection, it is mainly heat that is transported and not salt; thus, the stratification of the system is maintained. This process controls the character of the water on the shelf by quickly removing heat from UCDW intruding on the shelf, but leaving the salt. Hydrographic observations (Smith et al., 1999) clearly show recent intrusions of this distinctive water onto the shelf; but even a few months later, it is not possible to locate this water through temperature differences.

4.1. Surface heat budgets

The heat budget in summer is dominated by short-wave radiation while the winter budget is dominated by episodic sensible heat losses driven by low air temperature. The annual average for long-wave radiation is approximately 30 W m\(^{-2}\) with modest variations created by changes in cloud cover and air temperature. The latent heat flux due to water evaporation is the least significant term, being 4–10 times smaller than sensible fluxes.
Latent heat flux is low for the region because of high (80–90%) relative humidity. Low evaporative heat losses agree with observations (Cullather et al., 1998) that the west Antarctic Peninsula region is characterized by net precipitation over evaporation.

Despite heat loss from the surface ocean in 1993 (Fig. 7), estimated net heating from observed temperature changes between 1993 and 1994 was typically 5 W m\(^{-2}\) and never more than 25 W m\(^{-2}\) (Klinck, 1998). Heat lost to the atmosphere must be replaced by another source, which could include horizontal processes, vertical heat flux through the permanent pycnocline, and heat exchange between the surface layer and sea ice. Estimates for vertical heat flux across the pycnocline (Klinck, 1998; Smith et al., 1999) are between 5 and 10 W m\(^{-2}\). A persistent heat source of 5 W m\(^{-2}\) is enough to balance the net loss to the atmosphere (dashed line in Fig. 7).

Closure of the surface heat budget with only vertical processes does not rule out horizontal fluxes of heat (and salt). However, they do suggest that horizontal fluxes should approximately balance over the year and that much of the local variability is explained by vertical processes. A further indication of the validity of the budgets can be found from average, net heating between individual 1993 cruises (i.e., the slope of the curves in Fig. 7, about ±50 W m\(^{-2}\)), which agree with observed changes in heat content of surface waters (Klinck, 1998).

The importance of advective heat flux is evaluated from \(\rho_c C_{po} Hu(\partial T/\partial x)\), where the layer thickness (\(H\)), advective speed (\(u\)), and spatial temperature gradient are estimated from observations. Both mixed layer and sub- pycnocline layers have thickness of about 100 m. An estimated advection speed of 0.02 m s\(^{-1}\) is appropriate for across-shelf or alongshelf directions. The sub-pycnocline, across-shelf temperature gradient is 0.4°C per 100 km. Under these choices, advection can produce 40 W m\(^{-2}\), which indicates the importance of sub- pycnocline, lateral heat flux to this system. Near the surface, the temperature gradient is much smaller, say 0.01°C per 100 km, resulting in 1 W m\(^{-2}\). These estimates are bigger for faster speed or higher gradients, but advective heat flux near the surface is modest compared to the vertical diffusion–convection flux (Fig. 4B) and small compared to the net surface heating (Fig. 6C).

Modeled heat fluxes agree with the heat budgets calculated using atmospheric data from Faraday Station; an expected result, given that the same equations were used in both calculations. The only real difference between the fluxes calculated from station data and the simulated heat fluxes is the use of derived SST in the simple budget. As already discussed, the overall heat budget is not sensitive to SST and therefore similarities between the calculations are not surprising. The model estimated heat flux between surface waters and the ice is 5–10 W m\(^{-2}\), while the ice-free heat flux is between 30 and 50 W m\(^{-2}\) with maximum heat loss on the order of 100 W m\(^{-2}\) (Fig. 12).

### 4.2. Ice processes

The simulated ice cycle reproduces well the GSFC SMMR/SSMI record, capturing key features such as the timing of initial growth and the rapid spring melt (Fig. 11A). The modeled ice cycle also reproduces much of the higher frequency variability observed in the ice field due to rapid atmospheric changes. The model agrees with our heat flux calculation (Fig. 6) that the timing of the ice cycle is largely determined by net solar radiation while the ice extent is determined by sensible heat loss to the atmosphere.

The model shows that UCDW on the shelf influences overall ice thickness. Once the summer surface stratification is eroded (due to surface mixing, cooling, and brine rejection), heat diffusing vertically through the permanent pycnocline is available to melt the ice. Similarly, increasing heat flux through the pycnocline reduces ice concentration, as was found for the Weddell Sea (Martinson, 1990).

The simulations demonstrate that the two dominant ice processes are freezing at the ice edge in leads (driven by sensible heat loss) and melting at the ice–ocean interface. When winter air temperature rises, surface freezing diminishes,
allowing persistent melt at bottom of the ice to reduce ice volume at these times.

4.3. Mixed layer and pycnocline characteristics

The depth of the mixed layer and the hydrographic character of the upper water column result from complex interaction between surface thermodynamics and momentum flux associated with wind- and ice-induced motion. Surface cooling during the fall, and the subsequent brine rejection associated with ice formation, drive the bottom of the winter mixed layer to the permanent pycnocline. Additional mixing due to momentum flux under the ice also deepens the mixed layer.

In the spring, short-wave radiation through open leads warms the surface water, which melts the ice over a few weeks. This modeled, rapid ice melt reproduces observations (Stammerjohn and Smith, 1996) and is a strong indication that the input of solar radiation dominates the timing of the ice cycle. Ice melt generates a 5–20-m-thick fresh layer at the surface, which warms (once ice has melted), generating a distinct seasonal pycnocline.

The free drift dynamics, which lack the retarding effect of internal ice pressure, tend to overestimate ice velocity and therefore the mixed-layer depth. The roughness parameter for ice (Mellor and Kantha, 1989) is a maximum of 0.05 m for ice 3 m thick and decreases linearly with ice thickness to zero, so frictional coupling with the water is not sufficient to slow the ice. Ice velocities in this study are typically 0.5–0.75 m s\(^{-1}\), which are high but consistent with results from other 1-D ice models (Ikeda, 1989). The winter mixed layer is somewhat deeper than observations due to this excess speed, but other aspects of the water properties match with observations, indicating that in this region, thermodynamic, not mechanical, processes control the surface layer.

The depth of the permanent pycnocline is controlled by the depth of the surface mixed layer in winter, which is controlled by the stratification. The proposed weak upward flux of salt from the UCDW on the shelf balances precipitation at the surface, maintaining the salinity contrast across the permanent pycnocline. If there were larger net ice export, more ice production or any other process to increase the salinity of the surface layer, then deeper winter mixed layers, perhaps to the bottom, might be possible. Similarly, a decrease in the intrusion of UCDW onto the shelf, either by an offshore shift or a slowing of the ACC, would allow a reduction of the sub-pycnocline salinity and a reduction of stratification. These differences would fundamentally change the structure of the water in this region, perhaps allowing the production of dense shelf water.

4.4. Trends from the 16-year simulation

The winter ice cover to the west of the Antarctic Peninsula is characterized by an interannual variability of high and low ice years, which correlates with atmospheric conditions (Stammerjohn and Smith, 1996; Jacobs and Comiso, 1993; Smith et al., 1996). This 1-D, thermodynamic sea-ice model, forced with atmospheric conditions from Faraday, is able to reproduce most of this high/low ice cycle (Fig. 13). Of particular interest is the model’s ability to reproduce the highest (1987) and lowest (1989) ice years on record. In addition to the extremes, the model reproduces the observed trends such as the consecutive high ice years of 1979–1982, which were immediately followed by three low ice years from 1983 to 1985. Therefore, these results support an atmospheric (that is, thermodynamic) control on the ice cover, as suggested by Smith et al. (1996).

5. Conclusions

The DOVETAIL experiment investigates dense water formation in the southern Weddell Sea and its transport past the tip of the Antarctic Peninsula. The western side of the Antarctic Peninsula does not play a role in deep-water formation nor does it affect the flow of dense water to the abyssal ocean. However, due to the close proximity of the southern side of the Antarctic Circumpolar Current to the WAP continental shelf break and the nearness of a rather warm, by Southern Ocean standards,
water mass (UCDW) to the surface, water properties upstream of the DOVETAIL region are affected by the WAP shelf. This warm water mass limits ice formation and prevents dense-water formation.

The analysis presented here has identified that surface waters of the west Antarctic Peninsula are heated in the summer by solar short-wave flux and cooled in the winter by sensible heat losses. Heat flux through sea ice is small (by a factor of 10 or more compared to flux in leads, Fig. 12a), but heat flux in winter through leads, even with small spatial coverage (10–20% leads), can be significant. Ice melt in the spring is due to solar warming of open water, which then melts ice. The near closure of the surface-layer heat budget argues for minor influence of lateral exchanges near the surface on the heat and salt budget.

The winter mixed layer is deep (about 100 m) and is capped in summer by a fresh warm layer about 20 m thick. The sub-surface WW layer stores heat that enters from above during the summer and from below throughout the year. This heat is released suddenly to the atmosphere when the seasonal pycnocline breaks down in the fall. Diffusive-convection is important in allowing heat flux across the thermocline without removing the halocline. Such preferential heat flux compared to salt flux was calculated from hydrographic observations (Klinck, 1998).

The depth of the permanent pycnocline is controlled by the depth of deep winter mixing, which is controlled by the stratification. The salinity contrast is sufficient to prevent convection to the bottom. Salinity increase in winter due to ice formation is not sufficient to break down the stratification. The amount of ice production required to increase the surface-layer salinity (33.8) to the bottom value (34.65) by freezing is 2.9 m, for a surface-layer thickness of 100 m and assuming a new ice salinity of 5. Such ice production, while not outlandish or impossible, is 3–6 times that seen on the WAP shelf. Such production would require considerable ice export or the presence of large and persistent polynyas.

A numerical model with only vertical processes can reproduce observed ice and mixed-layer properties when forced by observed atmospheric conditions. Thus, thermodynamic processes control sea-ice and mixed-layer properties, and lateral transport of heat and salt play small roles in this region. However, the intrusion of UCDW onto the shelf from the ACC is critical in maintaining observed conditions (Klinck, 1998). Furthermore, heat flux due to the intruding deep water (UCDW) affects winter ice thickness and prevents dense-water formation on this shelf.

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