Cross-shelf exchange in a model of the Ross Sea circulation and biogeochemistry

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Abstract

Transport of warm, nutrient-rich Circumpolar Deep Water (CDW) onto Antarctic continental shelves and coastal seas has important effects on physical and biological processes. The present study investigates the locations of this transport and its dynamics in the Ross Sea with a high-resolution three-dimensional numerical model. The model circulation is forced by daily wind stress along with heat and salt fluxes calculated from atmospheric climatologies by bulk formulae. All surface fluxes are modified by an imposed climatological ice cover. Waters under the Ross Ice Shelf are not included explicitly, but their effect on temperature and salinity is imposed in a buffer zone at the southern end of the model domain. A simple nutrient uptake is calculated based on the climatological chlorophyll distribution and Monod uptake kinetics.

Model circulation is strongly affected by bottom topography, due to weak stratification, and agrees with schematics of the general flow and long-term current measurements except near the southern boundary. The sea-surface temperature is similar to satellite estimates except that the warmest simulated temperatures are slightly higher than observations. There is a significant correlation between the curvature of the shelf break and the transport across the shelf break. A momentum term balance shows that momentum advection helps to force flow across the shelf break in specific locations due to the curvature of the bathymetry (that is, the isobaths curve in front of the flow). For the model to create a strong intrusion of CDW onto the shelf, it appears two mechanisms are necessary. First, CDW is driven onto the shelf at least partially due to momentum advection and the curvature of the shelf break; then, the general circulation on the shelf takes the CDW into the interior.

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1. Introduction

The Southern Ocean is a heterogeneous system composed of a variety of sub-regimes, each with its own characteristic physical, biological and chemical features (Tréguer and Jacques, 1992). For example, the Ross Sea (typically defined as the waters over the continental shelf from Cape Adare at 170°E to Cape Colbeck at 158°W (Fig. 1)) is the most biologically productive region of the Antarctic, has generally sluggish circulation and is dominated by polynya processes (Arrigo et al., 1998; Jacobs and Giulivi, 1998). In contrast, the West Antarctic Peninsula (WAP) region contains large populations of Antarctic krill (\textit{Euphausia superba}) and has an active circulation, including interactions with the southern boundary of the Antarctic Circumpolar Current (Marr, 1962;...
Hofmann et al., 1996; Hofmann and Klinck, 1998). One approach to understand the causes of these variations is through the analyses of multidisciplinary data sets in conjunction with numerical modeling. In order to address fundamental questions on the controls of phytoplankton productivity and growth in these two systems, a suite of physical and biological models is being developed and used in both locations. Physical forcing, which includes advective circulation, vertical mixing, vertical stratification and irradiance availability, is the primary factor producing the observed vertical and horizontal variability in phytoplankton distribution and primary production in the Ross Sea (Smith et al., 2000). Circumpolar Deep Water (CDW) is a relatively warm, salty and nutrient-rich water mass that flows onto the continental shelf at certain locations in the Ross Sea. This water mass moderates the ice cover through heat flux, provides a relatively warm subsurface environment for some animals and provides nutrients to stimulate primary production. CDW transport onto the shelf is known to be episodic, but persistent, and is thought to occur at specific locations due to bottom topography. This study focuses on the transport of CDW.

The circulation of the Ross Sea is dominated by a wind-driven gyre, whose flow is strongly influenced by three submarine ridges that run in a southwest–northeast direction. Flow over the shelf below the surface layer consists of two anticyclonic gyres connected by a central cyclonic flow (Locarnini, 1994). In spring and winter, flow is significant, although largely dominated by tides (Pillsbury and Jacobs, 1985; Jaeger et al., 1996). The flow below 240 m has been shown by direct
current measurements to be largely barotropic (Picco et al., 1999). The winds are quite variable, with frequent storms due to passing low-pressure systems as well as strong katabatic winds along the coast. The Ross Sea is covered with ice for much of the year and ice concentrations are influenced by winds, particularly in the south-central region where little in situ melting occurs (Arrigo et al., 1998). The ice remains in the western region throughout the austral spring and generally melts in January due to local heating. This leads to extremely strong stratification and shallow mixed layers in the western Ross Sea (Artegiani et al., 1992; Arrigo et al., 1998).

Following Carmack (1977), five separate water masses can be identified on or near the continental shelf, including Ice Shelf Water (ISW), Modified CDW (MCDW) and Antarctic Surface Water. ISW, which emerges from beneath the Ross Ice Shelf primarily in the west-central part of the continental shelf, has subsurface temperatures below the surface freezing point. MCDW, derived from oceanic CDW, has temperatures of 1.0 to $-1.5^\circ$C and is found near the shelf break, but can intrude well onto the shelf. Antarctic Surface Water (AASW) is warm (up to 2.0$^\circ$C) and fresh (33.5–34.5) being produced in summer from deeper shelf waters. AASW typically contains a mixed layer and the underlying Winter Water (WW) which is the deep cold mixed layer from the previous winter.

In this study, a finite difference primitive equation model is used to simulate the circulation of the Ross Sea in order to analyze the cross-shelf break transport of CDW.

2. Model configuration

The Rutgers/UCLA Regional Ocean Model System (ROMS), which is used for this study, is a coarse grain parallel primitive equation ocean circulation model derived from the serial S-Coordinate Rutgers University Model (SCRUM, see Hedström, 1997). The model has a free surface and uses a vertical $s$ (terrain-following) coordinate. The model domain (Fig. 1) covers the Ross Sea and extends well past the continental shelf break as far north as 68$^\circ$S. The southern boundary of the model is a closed solid free slip wall, but includes processes from the Ross Ice Shelf (details below). All the other three boundaries are partially or completely open, depending on the coastline (Fig. 1). The horizontal grid spacing (5 km) was chosen as a compromise between trying to capture all the scales that may be present in the topographic data and the available computing power. There are 24 vertical levels that are concentrated towards the top and bottom of the model domain (Fig. 2). The thickness of the top model layer varied from 2.0 to 33.6 m. The gridded bathymetry is derived from the 1/12$^\circ$ ETOPO5 (NGDC, 1998) and is artificially “straightened” for the last few grid points (isobaths set normal to the boundary) at every open boundary to improve performance of open boundary conditions. The interpolated bathymetry was somewhat smoothed with a modified Shapiro filter that was designed to selectively smooth areas where the changes in bathymetry are large with respect to the total depth (Wilkin and Hedström, 1998). Therefore, most of the smoothing was on the shelf, there was less smoothing over the shelf break and the mean slope over the entire model domain was reduced from 1.16% to 0.72%.

Fig. 2. Bathymetry and layer thickness across the shelf (south on the left side) up to the shelf break at 178$^\circ$E. The vertical lines represent every other grid point.
Laplacian horizontal mixing of tracers is along geopotential surfaces with a diffusivity of 5 m$^2$ s$^{-1}$. The third-order upstream momentum advection scheme (Gamma scheme of Shchepetkin and McWilliams, 1998) is naturally dissipative and therefore only a small background value (0.1 m$^2$ s$^{-1}$) of explicit horizontal momentum mixing was required. Quadratic bottom stress, with a coefficient of $3.0 \times 10^{-3}$ (non-dimensional), was applied as a body force over the bottom layer. The vertical momentum and tracer mixing were handled using the K profile parameterization (KPP) mixing scheme (Large et al., 1994), including the use of a surface mixed layer. The KPP scheme was slightly modified for the presence of sea ice by reducing the solar short-wave radiation contribution to the surface turbulent buoyancy forcing in proportion to the ice cover present. Lytle and Ackley (1996) have shown that solar warming of the mixed layer through ice results in only a small heat gain ($\approx 0.1$ W m$^{-2}$ with a maximum of 0.75 W m$^{-2}$). Also, due to the extreme stabilizing effect of melting ice, the KPP surface mixed-layer depth under stabilizing conditions was set to a minimum depth, equal to the directly wind forced minimum depth under stable conditions in a Kraus/Turner bulk mixed-layer model (Krauss and Turner, 1967; Niiler and Kraus, 1977). No double diffusive mixing was included.

Open boundaries used a two-dimensional radiation scheme combined with adaptive nudging (Marchesiello et al., 2001). The time scale of the nudging is based on inward or outward phase speed of properties at the boundary. The model setup at the boundaries differed from the Marchesiello et al. (2001) US West Coast experiment, only in that no nudging is used for the baroclinic velocities and $\tau_{in}$ (the time scale for nudging for incoming waves) is 10 days instead of 1 day ($\tau_{out}$ is 1 year in both cases). No sponge layer, or region of increased horizontal viscosity near the open boundaries, was used. Temperature and salinity on the open boundaries were relaxed to the World Ocean Atlas 1998 (WOA98) climatology. The barotropic velocities were relaxed to monthly depth-averaged circulation from the OCCAM (Webb, 1996; Webb et al., 1998) global 0.25° resolution model.

The model is initialized with horizontally uniform temperature and salinity and a vertical structure obtained from the average of the WOA98 values in the model domain. In order to lessen the shock from the initial geostrophic adjustment, the horizontal variation is ramped in by running the model for 5 days with the stratification strongly relaxed to the gridded temperature and salinity climatology. The non-linear equation of state is used. Silicate, nitrate and chlorophyll distributions are initialized from a new climatology created for the area (Smith et al., 2003) and then forced to remain constant for the first 60 days of the simulation.

Initial model runs were forced with monthly climatological wind stress from the ECMWF reanalysis (Trenberth et al., 1990) applied as a body force over the top three layers of the model. However, it became apparent from the surface layer structure that better temporal resolution of the wind was necessary. Daily values of wind stress and wind speed were obtained from a blend of NSCAT and ERS-2 scatterometer data and NCEP analyses to give a repeatable annual cycle from August 1996 through July 1997 (Milliff et al., 1999). All the results presented here will be from the model forced with the daily winds.

Instead of using a fully dynamic sea-ice model, ice concentrations from a climatology derived from the Special Sensor Microwave Imager (SSM/I) are imposed. The model surface heat flux is calculated as a linear combination of heat flux due to ice cover and the open-water heat flux with the ratio determined by the ice concentration in that grid cell (Markus, 1999). The open-water heat flux was calculated with the COARE bulk flux algorithm (Fairall et al., 1996), with most of the necessary atmospheric data coming from monthly climatologies from either the NCEP reanalysis (air pressure, humidity and air temperature) or the ISCCP cloud climatology. Daily winds were used for the open-water heat flux calculation. The prognostic model sea-surface temperature (SST) was used as opposed to a pre-defined SST climatology. The short-wave solar radiation was computed using the model of Zillman (1972) with the cloud-cover reduction formula of Dobson and Smith (1988) and an assumed oceanic albedo of
0.07. The distribution of solar irradiance in the water is calculated with the two-band model of Paulson and Simpson (1977) assuming a Jerlov (1976) water type I (very clear). The heat flux underneath the ice \( (H_{\text{ice}}) \) is calculated as
\[
H_{\text{ice}} = \rho_w c_p c_h u_* (T_{\text{mix}} - T_f),
\]
where \( \rho_w \) is the water density, \( c_p \) is the specific heat capacity of water, \( c_h = 0.006 \) is the heat transfer coefficient, \( u_* \) is the friction velocity calculated from the wind stress, \( T_{\text{mix}} \) is the mixed-layer temperature, and \( T_f \) is the salt-water freezing temperature (see, Markus, 1999 for details). The model surface fresh water flux (imposed as a salt flux) is also calculated as a linear combination of open-water evaporation minus precipitation and that due to ice melting or freezing (Markus, 1999). The salt flux underneath the sea-ice is a function of the basal melting or freezing which is controlled by the difference between the water-ice heat flux \( (H_{\text{ice}}) \) and the ice-atmosphere conductive heat flux calculated as in Markus (1999) (from Semtner, 1976). The frazil ice contribution to the salt flux in Markus (1999) was removed because the parameterization as presented does not directly rely on the temperature of the water column and thus could have frazil ice formation or melting in conditions that should preclude it. For the open-water salt flux calculation, precipitation was taken from the Xie and Arkin (1997) climatology. There is no relaxation of the surface temperature or salinity to climatology except at the open boundaries and ice shelf edge.

The southern edge of the model, which includes the edge of the Ross Ice shelf, should allow for the cavity under the shelf, but is instead a closed wall. Water properties are maintained by nudging the model temperature and salinity to the monthly climatology (from WOA98) with a time scale varying from 40 days at the end of a seven grid point buffer zone to 5 days at the ice shelf.

After the 5-day spin-up of the stratification, the simulation starts in late austral winter (September 15). The model nitrate and silicate are held to the new climatology until model day 60 (mid-November), while the chlorophyll concentration is always set to the climatology. After day 60 the macro-nutrients are simulated as a passive tracer subject to biological uptake (based on the chlorophyll concentration and Monod uptake kinetics, see Smith et al., 2003 for details), advection and diffusion. Model simulations have been run successfully for well over a year.

3. Results

3.1. Circulation

The circulation in the model has very small vertical shear below the mixed layer primarily due to the weak stratification. This result agrees with current meter measurements in the western part of the Ross Sea shelf that showed the circulation below 240 m to have little vertical shear with a limited seasonal variability (Picco et al., 1999).

Locarnini (1994) analyzed water-mass tracers and used a few long-term current meter records to develop a schematic of the subsurface flow in the Ross Sea suggesting two anticyclonic gyres, with the western gyre ending near 176°W and the eastern gyre beginning at about 172°W. The water properties also suggest a general outflow from the shelf to the deep ocean over troughs in the bathymetry and an inflow over banks. Current meter measurements and particle transport patterns in Jaeger et al. (1996) imply the existence of the western anticyclonic gyre. A circumpolar-coupled sea-ice–ice-shelf–ocean circulation model (Assmann et al., 2003), which includes the Ross Ice Shelf cavity and has moderate (\( \approx 40 \) km) resolution on the shelf, also has a seasonally varying western anticyclonic gyre. A vector plot (Fig. 3) of long-term (approximately 1 year) current meter measurements from several sources (Table 1) shows generally westward flow along the ice shelf (except for a few locations), a northward flow in the extreme west edge of the area and a strong circulation to the northwest along bathymetry on the continental slope.

A temporal average of the model circulation over a year (model days 30–390, Fig. 4) in the western shelf region shows the flow to be similar to the estimated circulation. Much of the model flow is constrained to follow topography and strong currents develop along the shelf break, at the base...
of the continental slope and next to Iselin Bank. An anticyclonic gyre clearly exists with the eastern edge of the gyre in the right location, although it does not quite match the Locarnini schematic, which has the eastern return flow underneath the ice shelf—clearly impossible with this model. Point comparisons of flow vectors are problematic because the small-scale and strongly topographically forced nature of the flow means than small errors in the position can lead to large differences in flow velocity at a specific location. With this in mind, the comparison to actual current meter measurements is quite reasonable. The model flow matches the general westward flow along the ice-shelf edge, although it obviously does not have the flow into or out of the cavity underneath the ice shelf shown at several places. The circulation on the western boundary and on the continental slope match the observations well in direction and magnitude. The small-scale spatial variability in the model current near the location of the three moorings in the interior of the shelf makes a comparison more difficult, but the model flow agrees generally with observations.

The general pattern of cross-shelf flow is bathymetrically controlled (Fig. 4). The ridges running north/south generate cyclonic circulation around the banks and anticyclonic circulation around depressions. At 300 m, the temperature (Fig. 5) and salinity (not shown) show the strong influence of circulation on property distributions. Specifically, warm CDW floods the shelf in the far east and intrudes onto the shelf in the west over and along the western edge of submarine banks. Tongues of cold water flow northward along the western boundary, 173–175°E and 178°W (182°E) over and along the western edge of troughs. These tongues are not present in the initial fields (not shown) but match observations of
minimum temperature and bottom salinity shown in Locarnini (1994).

3.2. Temperature

The accuracy of the model’s surface processes is tested by comparing the model SST to satellite estimates. Since there is no relaxation to a predefined temperature, the calculation of SST is a check on several processes, including the open-water surface heat flux, heat transfer between the ice and the water, penetration of the solar heating below the surface, horizontal advection and vertical mixing. The Ross Sea is ice covered over much of the year and thus the SST in the model is close to the freezing point of water for most of the annual cycle. Typically, reduced ice concentrations first appear in the interior of the Ross Sea, becoming a large open area in December. By January, this area is generally open to the north; by March, the ice cover starts to fill in again from the south to the north.

A climatology of Advanced Very High Resolution Radiometer (AVHRR) satellite SST covering 1985–2000 with 9 km resolution, averaged over 5-day periods (Casey and Cornillon, 1999) and interpolated onto the model grid, is compared to model SST. Temperature increases first in the central Ross Sea low ice regions in December and January. By February the interior cools while the temperature north of the shelf increases. The timing of the SST annual cycle in the model matches very well with observations. In early January (Fig. 6), the location of the warmer ($> -0.5^\circ C$) surface water in the model matches closely with observations, although the model's

| Table 1 | Long-term current meter measurements |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| Latitude | Longitude | Depth (m) | Duration (days) | $u$ (cm s$^{-1}$) | $v$ (cm s$^{-1}$) | Ref. |
| 76°41.0'S | 169°00.2'E | 220 | 362 | 0.6 | −0.5 | 1 |
| 76°41.0'S | 169°00.2'E | 780 | 362 | 1.5 | −0.7 | 1 |
| 74°01.5'S | 175°00.6'E | 220 | 397 | 0.4 | −0.6 | 1 |
| 72°29.6'S | 175°08.1'E | 555 | 397 | 2.8 | 0.1 | 1 |
| 75°06.1'S | 164°13.0'E | 230 | 399 | −2.3 | 6.0 | 1 |
| 75°06.1'S | 164°13.0'E | 425 | 399 | −5.9 | 5.8 | 1 |
| 75°06.1'S | 164°13.0'E | 55 | 331 | −0.8 | 7.7 | 1 |
| 74°01.5'S | 175°00.6'E | 140 | 331 | −1.1 | 3.0 | 1 |
| 74°01.5'S | 175°00.6'E | 402 | 331 | 0.6 | 2.1 | 1 |
| 74°01.5'S | 175°00.6'E | 748 | 331 | 2.6 | 0.9 | 1 |
| 74°01.5'S | 175°00.6'E | 885 | 331 | 4.9 | 0.4 | 1 |
| 78°00.0'S | 177°01.6'W | 244 | 219 | −3.5 | 2.5 | 1 |
| 78°00.0'S | 177°01.6'W | 391 | 256 | −2.2 | 2.0 | 1 |
| 78°00.0'S | 177°01.6'W | 579 | 223 | 2.0 | 4.6 | 1 |
| 75°56.2'S | 177°36.0'W | 283 | 353 | 3.2 | 1.7 | 1 |
| 75°56.2'S | 177°36.0'W | 450 | 353 | 3.8 | 1.5 | 1 |
| 75°56.2'S | 177°36.0'W | 503 | 353 | 2.7 | 1.5 | 1 |
| 75°56.2'S | 177°36.0'W | 597 | 353 | 2.7 | 0.9 | 1 |
| 78°13.6'S | 172°29.4'W | 211 | 355 | 0.4 | −5.0 | 2 |
| 78°13.6'S | 172°29.4'W | 383 | 355 | 0.9 | −3.6 | 2 |
| 78°11.0'S | 174°39.0'W | 237 | 355 | −8.8 | −1.4 | 2 |
| 78°11.0'S | 174°39.0'W | 310 | 354 | −8.2 | 1.8 | 2 |
| 78°05.5'S | 175°30.0'W | 492 | 355 | −7.6 | −1.2 | 2 |
| 78°05.5'S | 175°30.0'W | 253 | 350 | −5.9 | −0.4 | 2 |
| 78°05.5'S | 175°30.0'W | 327 | 354 | −6.2 | −1.4 | 2 |
| 78°01.0'S | 179°46.0'E | 508 | 354 | −2.6 | 2.0 | 2 |
| 78°01.0'S | 179°46.0'E | 444 | 204 | 1.0 | 1.5 | 2 |

The warmest water is somewhat to the east and slightly warmer (0.1°C) than the observations. By late February (not shown), both the observations and model are cooler in the interior (the maximum temperatures are very similar) and have the warmest temperatures north of the shelf. To quantify the SST comparison, the model values are compared to observations at every grid point where AVHRR data were available (low or no sea-ice) every 10 days whenever AVHRR data is available for >90% of the model domain (late November through mid-March). The average rms error over this period was 0.34°C with a maximum of 0.44°C in late January. The errors over the model grid are mostly compensated at any given time leading to a time and space mean error over the period of only 0.010°C (maximum: −0.13°C). The longest model simulation to date is 1.7 years; a comparison of the SST in January of the second year does not show any obvious difference from the first year’s results or the climatology.

The warmest summer SST was higher than indicated by the climatology, although the spatial pattern was correct. Two general problems could produce the excess heating. The open-water heat flux formulas used were derived for tropical regions and may not be appropriate in high latitudes. Note that different bulk algorithms vary significantly in their calculation of heat flux (Zeng et al., 1998). A related problem could be due to the cloud correction algorithm. Different cloud-cover algorithms have a wide variety of effects in the Antarctic (Kim and Hofmann, in preparation).

The second problem is the magnitude of vertical mixing, which is estimated from a turbulence parameterization. Other than the daily cycle of solar radiation, the external forcing of the surface
fluxes for the initial run is based on monthly climatologies with a coarse (up to 2.5°) spatial resolution. Just switching to daily winds with 0.5° horizontal resolution increased the vertical mixing and improved the comparison to observed SST (average r.m.s. error reduced from 0.44°C to 0.35°C). The true winds in the area, however, include strong coastal katabatic wind surges and winds generated by the frequent passage of low-pressure systems (Bromwich and Stearns, 1993). The lack of forcing from individual high-energy atmospheric events limits mixing, especially in austral summer when mixing driven by wind energy competes against the stabilizing effect of the solar radiation. Furthermore, the coefficients in the turbulence model may need to be adjusted. Finally, double diffusive mixing, important at high latitudes (Smith and Klinck, 2002), was not used.

A cross-section of the model temperature in mid-January (Fig. 7a) and mid-July (Fig. 7b) shows the seasonal cycle of the mixed layer. In January, the SST is highest over the continental shelf due to a shallow and relatively warm surface layer. In July, the mixed layer over the shelf has cooled considerably and has deepened to greater than 100 m. This compares reasonably with the few observations of the mixed layer underneath sea-ice in austral winter in the Southern Ocean near the continent (Gordon and Huber, 1990; Klinck, 1998; Gouretski, 1999).

The model temperature sections show the strong gradient at the shelf break below the mixed layer between the shelf waters and the warmer CDW off-shelf that has been referred to as the Antarctic Slope Front (Ainley and Jacobs, 1981; Jacobs, 1986, 1991). This front supports a westward flow along the shelf margin (in the model and observations) and often is characterized by a strong “V” shape of isotherms at the shelf break (Fig. 7b). A mid-depth intrusion of warmer water onto the shelf is visible in the −0.4°C and −0.6°C isotherms in the winter section.

4. Discussion

A primary motivation for this model is the study of exchange of water masses across the shelf break and its effect on biological processes. For example, CDW is shown to be important, due to its
Fig. 6. (a) Model top layer temperature (°C) at t = 110 days (January 2) and (b) Satellite (AVHRR) SST (°C) climatology for the period 1–5 January.
temperature, to the reproductive cycle of Antarctic krill (Hofmann et al., 1992). The specifics of cross-shelf nutrient transport, as part of CDW intrusions, are important for biological production (Prézelin et al., 2000). Also, the rate and timing of vertical diffusion providing heat and nutrients to the surface layer affect annual cycles of sea-ice and summer production.

Fig. 7. (a) Model temperature (°C) section across the shelf and shelf break at 178°E at t = 120 days (mid-January). Temperatures below 0°C are represented with dashed lines and the contour interval is 0.2°C and (b) Model temperature (°C) section across the shelf and shelf break at 178°E at t = 300 days (mid-July).
4.1. Shelf budgets

In order to study exchange across the shelf break, an enclosed “shelf” is defined by the 1000 m isobath from Cape Adare to 158°W and then to land at each end. The variability of velocity and tracers along the shelf break was strong enough that it was necessary to calculate running totals of volume and tracer fluxes during a model simulation. A budget of the volume flux onto the shelf (Fig. 8a) computed from 5-day averages shows the mean total transport onto the shelf to be extremely small (<100 m$^3$ s$^{-1}$ on average, Table 2). Note that the total horizontal transport onto the shelf does not have to be exactly zero due to the free surface. Although the volume flux is basically zero, there is a small net on-shelf transport of heat and small off-shelf transports of salt, nitrate and silicate. After the initial spinup, there is not much of a seasonal signal in the net transport of heat or salt. However, since the vertical gradient of nitrate along the shelf break is small except (sometimes) in the surface layer, the net volume transport onto the shelf is almost zero and the transport in the surface layer tends to be onto the shelf (see below), there is more net off-shelf transport of nitrate in austral fall when the surface nitrate is depleted due to biology.

The annual mean uptake of nitrate by the biology in the top 100 m is 11.7 kmol s$^{-1}$. Since the cross-shelf break nitrate transport is slightly off-shelf, this indicates that cross-shelf transport alone cannot close the annual nitrate budget on the shelf and some other important process, such

Fig. 8. (a) Advective total transport onto the shelf of volume (Sv), heat (TWatt), salt (psu-Sv), nitrate (kmol s$^{-1}$) and silicate (kmol s$^{-1}$) over the period $t = 30–390$ days. Positive values indicate transport onto the shelf and (b) Advective transport onto the shelf of volume (Sv), heat (TWatt) and salt (psu-Sv) for different water types over the period $t = 30–390$ days. Positive values indicate transport onto the shelf. Nitrate and silicate fluxes (not shown) are very similar to the salt flux. See text for definition of the water types.
as nitrification or remineralization, must be missing in the model. This is discussed in more detail in Smith et al. (2003).

The volume flux onto the shelf in the top three model layers (Fig. 8b), which are directly forced by the wind stress, is 0.30 Sv and the timing and amount are consistent with the surface Ekman flux (not shown). Prevailing winds at the shelf break are primarily from the south and south-east. This flux brings little heat onto the shelf since the surface water has the same temperature on and off the shelf and is near the freezing point much of the year (heat content here is calculated with respect to freezing); but it does bring some salt and nutrients. There is considerably more off-shelf volume flux in a relatively strong bottom Ekman layer. The bottom friction in the model is applied as a body force over the bottom layer which, in 1000 m deep water, is only 11.7 m thick. However, this bottom layer has a net off-shelf transport of 0.76 Sv, and also transports heat, salt and nutrients off the shelf.

The cross-shelf break transport of CDW, defined as water below 100 m with a temperature above 0.5°C, is time variable but averages 0.37 Sv onto the shelf. This water brings in substantially more heat, and comparable amounts of salt and macro-nutrients as the water near the surface.

The cross-shelf break transport of bottom water was calculated for water below 400 m with a temperature less than −0.5°C and a salinity greater than 34.6. The annual average volume flux is small (−0.66 Sv) and includes two pulses starting in early February (model day 140) and early August (model day 320). The direction is always off-shelf. The timing of the two pulses is interesting. The model is initialized from the WOA98 database that has surface water that is too warm. The imposed winter ice cover at the beginning removes heat from the model and creates brine by freezing. In the 30 days between model days 180 and 210, the ice cover over the western half of the shelf changes from 30–70% to almost 100% also resulting in the model creating cold dense water. The time lag of ≈130 days is the time required for this dense water to drain off the shelf. However, the model still does not create much bottom water. The volume of bottom water created in the Ross Sea is uncertain, but Locarnini (1994) estimated that 33% of the production of Antarctic Bottom Water (AABW) comes from the Ross Sea which, assuming 8–12 Sv of total AABW production (Orsi et al., 1999) results in 3–4 Sv coming from the Ross Sea. It should be noted that the Orsi et al. estimate is of newly formed bottom water sinking past the 2500 m isobath, and there is still some entrainment of environmental waters into the bottom water plumes off-shore of the 1000 m isobath (the depth used for the model estimate).

Several model limitations explain this deficiency. The model is forced by climatological atmospheric conditions that are temporally and spatially smoothed. Several intense polynyas along the western side of the Ross Sea (for example in Terra Nova Bay), which are not included in the model, are known to produce very dense water. Additionally, there are no strong storms to provide extra ice production, which will produce colder saltier water. Finally, the imposed ice does not fully represent the effects of ice freezing on water-mass properties. In addition to all of these shortfalls, the model water properties have not fully adjusted to the imposed surface fluxes during this short simulation.

<table>
<thead>
<tr>
<th>Water type</th>
<th>Vol. (Sv)</th>
<th>Heat (TW)</th>
<th>Salt (psu-Sv)</th>
<th>Nitrate (kmol s⁻¹)</th>
<th>Silicate (kmol s⁻¹)</th>
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</thead>
<tbody>
<tr>
<td>Total</td>
<td>−8.1 × 10⁻²</td>
<td>4.6</td>
<td>−0.72</td>
<td>−3.1</td>
<td>−7.2</td>
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<td>Sfc. Ekman</td>
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<td>0.21</td>
<td>10.2</td>
<td>7.1</td>
<td>18.3</td>
</tr>
<tr>
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<td>12.8</td>
<td>11.9</td>
<td>38.2</td>
</tr>
<tr>
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<td>−2.1</td>
<td>−23.1</td>
<td>−20.3</td>
<td>−54.2</td>
</tr>
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</table>
4.2. Shelf break exchange processes

Transport of CDW across the shelf break is thought to be tied to certain bathymetric features. For example, a linkage has been proposed between topographically induced upwelling of nutrient-rich CDW (primarily at sites of shallow topography near the shelf break) onto the shelf and the presence of diatom dominated phytoplankton assemblages on the WAP continental shelf (Prézelin et al., 2000). A cross-section (Fig. 9) of the mean cross-slope velocity along a section in the northwest part of the shelf shows a relatively strong bottom Ekman layer with water flowing off the shelf and generally barotropic flow above that with alternating sections of on- and off-shelf transport. The cross-slope velocity at each individual grid point along the shelf break is calculated as

\[
U_{\text{cross}} = \bar{U} \cdot \left( \frac{\nabla H}{\nabla H} \right),
\]

where \(H\) is the bathymetry.

Even though the total model cross-shelf volume transport is very close to zero, there are several locations that have relatively large and variable transport consistently on or off the shelf. It appears that many of the locations of across shelf transport are related to the curvature of the shelf break (Fig. 1). This curvature can be numerically computed as the change along the shelf break (where \(s\) is the distance along the shelf break) of the unit gradient of the bathymetry

\[
d \left( \frac{\nabla H}{\nabla H} \right).\]

The curvature is defined to be positive when, going from east to west, the shelf break rotates clockwise. Since the flow along the shelf break is primarily from east to west, positive curvature rotates the bathymetry so that momentum advection would tend to drive flow onto the shelf. When the annual mean cross-shelf break volume flux as a function of position along the shelf break is compared to the curvature (Fig. 10), there is an obvious correlation between the volume flux and the curvature with the on-shelf flow lagging.

**Fig. 9.** Cross-section of the mean cross-slope velocity averaged over the period \(t = 30–390\) days for shelf break indices (20–140). The contour interval is 0.5 cm s\(^{-1}\), dashed lines indicate negative values and positive values indicate flow onto the shelf.

**Fig. 10.** Cross-shelf break volume flux (Sv), shelf break curvature (normalized units \(\times 0.25\)) and cross-shelf break momentum advection term for the entire grid cell volume of water (10\(^3\) m\(^4\) s\(^{-1}\)). These are averaged values over the period \(t = 30–390\) days. The x-axis is grid number along the defined shelf break.
by a couple of grid points. The correlation is at a maximum at a lag of either 2.5 \( (r = 0.520) \) note that the curvature locations are one-half grid point removed from the cross-shelf flow locations) or 3.5 \( (r = 0.497) \) grid points corresponding to 12.5–17.5 km in along shelf break distance. The large-lag standard error (Sciremammano, 1979) is computed to be 0.111, so that the correlation is easily statistically significant at the 99% level \( (2.6\sigma) \). Physically, this indicates that circulation crosses the shelf break in places where the tendency of the flow to maintain a given direction would have it cross a bathymetric contour that is rapidly changing direction. The maximum cross-shelf break flux occurs about 15 km after the maximum change in the shelf break direction. No significant correlation was found between the cross-shelf break flux and a quantitative value (either the magnitude of the slope or the depth of the terrain several grid points back) of whether the terrain near the shelf break is “shallow” or “deep”.

The terms in the momentum acceleration equation were calculated from time-averaged quantities at each layer and summed over the entire water column and then rotated at each grid point into a cross-shelf break and along shelf break component. Over most of the shelf break length, an annual average shows the dominant terms in the momentum balance to be the Coriolis term, the pressure gradient force term, and the horizontal momentum advection term. The surface wind stress and bottom drag, while important in the top and bottom boundary layers (Fig. 8b), do not have a large impact on the cross-shelf acceleration of the entire water column. The momentum advection term is greater than the residual (sum of Coriolis, pressure gradient force and horizontal momentum advection) overall of the shelf break edge except between shelf edge points 240–275. In this area (Fig. 1), the temporal and spatial variability is too great to accurately calculate a term balance from 5-day averages. The momentum advection term (Fig. 10) is very strongly correlated with the shelf break curvature \( (r = 0.682) \) and is significantly correlated with the cross-shelf break transport with a lag of either 3 \( (r = 0.412) \) or 4 \( (r = 0.422) \) grid points (large-lag standard error 0.119). In the numerical model, momentum advection helps to force flow across the shelf break in specific locations due to the curvature of the bathymetry.

Along the shelf break where there is mean on-shelf flow, the deeper CDW that is driven upward by the shallowing bathymetry should raise nutrients that could potentially support a localized increase in biological production.

However, how many of these locations of strong on-shore flow lead to CDW being transported well onto the shelf? It was noted previously that the north/south ridges on the shelf generate cyclonic circulation around banks and anticyclonic circulation around depressions. Warm CDW was shown to intrude onto the shelf in some areas (e.g., along 178°E; Fig. 5) over the western edge of a submarine bank or eastern edge of a trough. For example, the on-shelf flux around shelf edge grid indices 77, 92 and 104 (Fig. 1) can be carried into the interior of the shelf by the general circulation of the shelf. Conversely, the on-shelf flux around indices 61 or 236 does not intrude very far (Fig. 5) because of the mean circulation on the shelf. For the model to create a strong intrusion of CDW onto the shelf, it appears two mechanisms are necessary. Much of the CDW first appears on the shelf due to momentum advection and the curvature of the shelf break. Then, the general circulation on the shelf, which in this case is strongly influenced by bathymetric variations, pushes the CDW into the interior or back off the shelf.

4.3. Model deficiencies and future plans

There are several improvements that we hope to make to this model in the near future. Circulation near the Ross Ice Shelf can be improved to better represent flow under the shelf. This can be done by forcing flow in the buffer zone or by expanding the domain to include part or all of the ice cavity. The imposed temperature and salinity in the southern buffer zone need to be more realistic to better represent the character of water (ISW) produced in the ice cavity. A model of the Weddell Sea that included sub-ice shelf cavities (Beckmann et al., 2001)
has shown that water modified in the cavities contributes significantly to the water-mass formation along the continental slope.

Most of the energy in the current meter measurements is at tidal periods (Pillsbury and Jacobs, 1985; Jaeger et al., 1996; Picco et al., 1999) and barotropic and baroclinic tides can have a large impact on mixing processes (Muench et al., 2002). Tidal circulation, driven both at the oceanic boundaries and the entrance to the Ross Ice Shelf cavity, will be added based on a regional (e.g., Padman et al., 2002; Robertson et al., 2003) or global tide model.

The model surface salinity is too low in austral summer (figure not shown), which is due to the ice masked surface flux that does not account for a finite volume of ice, thereby allowing melting to create fresh water long after the ice should be completely melted. This problem will be fixed with a dynamic sea-ice model although that will raise other difficulties in getting the ice model to behave realistically.

Weak vertical mixing in the surface layer indicates a need for better representation of the mechanical forcing (wind stress) and perhaps some changes to the KPP parameterization. Using higher spatial and temporal resolution wind stress from a blended satellite/analysis product helped significantly. There are also some coastal locations with automated wind stations that could be used to give more realistic winds in some areas. However, none of these are likely to pick up all the details of the wind forcing, especially the very important katabatic winds near the coast. Coupling the ocean model to a regional mesoscale atmospheric model, such as Polar MM5 (Bromwich et al., 2001), may be the solution to this problem.

Finally, brine rejection during the formation of ice in polynyas is considered to be very important in the creation of bottom water (Zwally et al., 1985); lack of polynyas may be the cause of the small amounts of bottom water created in the model. Properly simulating coastal polynyas, or at least mimicking the effect they have on the underlying water, will also require better resolution in the forcing data and a true dynamic sea-ice model (Van Woert, 1999).

Acknowledgements

We thank the Office of Computing and Communications Services at Old Dominion University for the use of the Sun HPC 10000 on which the simulations were run. Dr. Herman Arango provided the Beta 1.0 release of ROMS. The AVHRR SST data were provided by Dr. Kenneth Casey. The OCCAM model output fields are courtesy of the Southampton Oceanography Center. Dr. Aike Beckmann suggested looking at the effect of curvature on the cross-shelf break transport in an extremely helpful review. Computer facilities and support were provided by the Commonwealth Center for Coastal Physical Oceanography. This work was supported by the U.S. National Science Foundation Grant OCE-99-11731.

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