Influence of tides on sea ice in the Weddell Sea: Investigations with a high-resolution dynamic-thermodynamic sea ice model

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The effects of tides on the seasonal fluctuations and regional differences of the sea ice cover in the Weddell Sea are investigated using a dynamic-thermodynamic sea ice model operating at high temporal and spatial resolution. Tidal currents add a highly variable stress force at the ice-water interface. They are derived from an eight-constituent barotropic, ocean-only, tidal model. Thermodynamics are calculated for seven ice classes plus open water; dynamics consider an averaged ice class and open water only. Results from runs of the sea ice model with and without the additional tidal stress demonstrate that tidal currents alter the local and, to lesser extent, basin-wide evolution of sea ice. The tidal currents reduce the expansion of the ice cover, mostly cause a speed up of the retreat, and lead to a smaller minimum ice extent. The most significant local differences between the two model simulations are found in tidally active regions over the continental shelf breaks of the western Weddell Sea, in front of Filchner-Ronne Ice Shelf, and offshore Dronning Maud Land.


1. Introduction

The Antarctic sea ice cover is considered a key factor for high southern latitude interactions between atmosphere and ocean, thus also being relevant to the global climate system [Kreyoscher et al., 2000]. Its overall extent and ice concentration has been routinely observed with satellite sensors for several decades [Zwally et al., 1983; Barber et al., 2001; Fowler et al., 2001; Zwally et al., 2002]. Ice drift was observed in selected areas by, for instance, upward looking sonars [e.g., Strass and Fahrbach, 1998; Harms et al., 2001], and the large-scale velocity fields have, until recently, mainly been measured in situ by ice-mounted buoys [Kottmeier et al., 1997].

More information about the basic features of the sea ice cover can also be obtained from dynamic-thermodynamic sea ice models driven by atmospheric and oceanic forcing [Hibler, 1979; Beckmann et al., 1999; Timmermann et al., 2002; Pereira et al., 2002, and references therein], which have been extended to large spatial scales. Most existing ice models are run with atmospheric forcing only: if ocean effects are included, it is typically through simple inclusion of mean flows associated with the wind stress and the thermohaline forcing due to surface buoyancy fluxes. Such models are capable of qualitatively reproducing the general properties of the sea ice distribution. However, several studies have suggested that nonatmospheric forcing can also play a significant role in setting the distributions of sea ice motion, concentration, and thickness, for instance by ocean heat flux associated with convection and upwelling [e.g., Martinson et al., 1981; Middleton and Humphries, 1989; Robertson et al., 1995; Muench et al., 2001].

To date, most studies of ocean influence on sea ice have been based on observations of ice variability in the Weddell Sea [Foldvik et al., 1990; Geiger et al., 1998; Padman and Kottmeier, 2000; Geiger and Drinkwater, 2005]. These authors highlight the role that strong tidal currents can play in setting local ice characteristics. Tides can influence the sea ice cover in two ways. (1) Lateral shear and strain associated with the small spatial scales of tidal current variability may be sufficient to greatly increase the average oceanic energy loss to the atmosphere, the rate of ice formation, and associated salt input to the upper ocean [Eisen and Kottmeier, 2000]. The internal stresses within the sea ice cover presumably produce a spatial ice thickness distribution that is quite different from the case without the action of tides. (2) Tidal mixing due to stress at the base of the sea ice and within the water column may also enhance the heat flow from deeper water masses to the surface mixed layer, causing reduced bottom freezing.

While it is fairly clear from the previously cited studies that tidal currents can have a local effect on...
sea ice, no attempt has yet been made to assess the effect of tides on the regional distribution of ice properties around Antarctica. Kowalik and Proshutinsky [1994] ran a simple ice model coupled to a barotropic ocean tide model of the Arctic Ocean to show that tides by themselves could increase annual mean ice formation rates by up to 10–100 cm a−1. Recently, Kwok et al. [2003] have used RADARSAT sequential imagery to show that high-frequency (roughly semidiurnal) ice divergence in the high Arctic Ocean could generate a significant lead fraction and associated ice formation, even in winter when the mean sea ice concentration is very high. These authors attribute most of the semidiurnal divergence signal to near-inertial oscillations forced by the wind, a result that is consistent with the findings of Padman and Kottmeier [2000] for the deep central basin of the Weddell Sea. The picture that emerges from all of these studies is that high-frequency periodic divergence, whether due to tides or the inertial response of the ice to wind stress, can be an important contributor to the ocean heat loss to the winter atmosphere in polar regions, and the associated ice formation rates and salt flux to the upper ocean.

With this background in mind, our goal in this paper is to investigate whether the effect of tides on sea ice properties around Antarctica is only significant in locations where tidal currents are large, or if tidal effects can be significant on regional scales. As this is primarily a proof-of-concept study, we consider only the Weddell Sea. As indicated above, there are several prior studies from the Weddell Sea demonstrating at least local effects of ocean tides on sea ice, and sufficient ocean current data exist to indicate that recent models of tidal currents [Padman et al., 2002] are sufficiently reliable for this study. We use the comparison of numerical simulations with and without tidal currents to demonstrate the role of tides. Both simulations include realistic atmospheric forcing, thermodynamic processes, and mean ocean currents, so that advection of tide-induced anomalies within the ice cover can occur. This is a significant improvement on the Kowalik and Proshutinsky [1994] Arctic model, which was forced only by ocean tides and did not consider atmospheric and thermodynamic processes, so that the effects of tides could not be distributed around the basin.

The most significant operational impact of adding ocean tidal effects to atmospherically forced sea ice models is the need for higher temporal and spatial resolution. Atmospherically forced models are usually run at a resolution of about 1° latitude and longitude with temporal resolution of about 6 hours to 1 day. In contrast, tidal currents have spatial scales of variability that are similar to the scales of bathymetry changes, and currents oscillate through a complete period with timescales of ~12 and ~24 hours. Much of the tidal kinetic energy in the Weddell Sea is in the form of diurnal topographic vorticity waves (TVWs) that have cross-slope scales of about 20–50 km and along-slope wavelengths of 100–300 km [Padman and Kottmeier, 2000]. We choose a model resolution of 0.1° meridionally, 0.3° zonally, and 2 hours in time to overcome the limitations in spatial and temporal resolution of large-scale models. Tidal forcing is provided by a barotropic tidal model that prescribes the oceanic forcing fields for a dynamic-thermodynamic numerical sea ice model. Feedback from the ice cover to the ocean tides and mean circulation is ignored in the present study. Thermodynamics are considered for seven ice thickness categories and open water, ice dynamics are evaluated for a single, averaged ice thickness class.

Further details on the modeling approach, physics, and forcing fields are provided in the following section. The temporal evolution of the most important sea ice properties resulting from the model runs over the 2 year period 1995/96, with and without tidal currents, are presented in section 3. A discussion and conclusion follow in sections 4 and 5, respectively.

2. Modeling Approach and Realization

We use a modified version of the sea ice model described by Harder [1996] and Kreysscher et al. [2000]. It consists of three parts: ice dynamics, ice thermodynamics, and the mixed layer module. The different modules mainly incorporate the work of Hibler [1979] (ice dynamics), Parkinson and Washington [1979] (thermodynamics), Owens and Lemke [1990] (effects of snow cover), and Lemke [1987] (mixed layer temperature evolution). Tidal influence enters through the oceanic forcing fields calculated from the barotropic, ocean-only, Circum-Antarctic Tidal Simulation (CATS) [Padman et al., 2002].

In this section, we first present the components of the sea ice model, its governing equations and modeling parameters. Next, a short description of the barotropic tidal model is given, followed by specifications of the different sea ice forcing fields. The final subsection describes our concept for model runs and derived quantities of the model output.

2.1. Sea Ice Model

The sea ice model treats the ice cover as a two-dimensional continuum consisting of a single ice thickness category plus open water for dynamic calculations and seven ice thickness classes plus open water for thermodynamic calculations. Sea ice drift \( \mathbf{u} \) is derived from the momentum equation [Hibler, 1979]

\[
m \frac{d \mathbf{u}}{dt} = \mathbf{\tau}_a + \mathbf{\tau}_w - \mathbf{F}_C + m \mathbf{g} \nabla H + \mathbf{F}_l,
\]

where \( m \) is the ice mass per unit area, \( m(d\mathbf{u}/dt) \) is the acceleration term, \( \mathbf{\tau}_a \) is the wind stress, \( \mathbf{\tau}_w \) is the oceanic stress, \( \mathbf{F}_C \) is the Coriolis force, and \( m \mathbf{g} \nabla H \) is the force due to the sea surface tilt \( \nabla H \).

The internal force \( \mathbf{F}_l \) due to sea ice stress and strain is described by a viscous plastic rheology with a truncated elliptical yield curve [Hibler, 1979], where the ice strength \( P \) depends linearly on the ice thickness \( h \):

\[
P(h, A) = P^* h e^{-C(1-A)}.
\]

The exponential factor \( e^{-C(1-A)} \) accounts for the fact that internal stresses mainly become effective for high ice concentrations, \( A \rightarrow 1 \). Truncation avoids unrealistically large viscosities for very small strain rates. The empirical model parameters are chosen as \( P^* = 2 \times 10^4 \text{ Nm}^{-2} \) and \( C = 20 \), obtained from model comparisons with buoy data [Kottmeier et al., 1997] and satellite-based ice drift data.
Table 1. Empirical Modeling Parameter Values

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C$</td>
<td>rheology parameter</td>
<td>20</td>
</tr>
<tr>
<td>$P^s$, Nm$^{-2}$</td>
<td>ice strength</td>
<td>$2 \times 10^4$</td>
</tr>
<tr>
<td>$c_w$</td>
<td>atmospheric drag coefficient</td>
<td>$1.6 \times 10^{-3}$</td>
</tr>
<tr>
<td>$c_d$</td>
<td>oceanic drag coefficient</td>
<td>$4.5 \times 10^{-3}$</td>
</tr>
<tr>
<td>$c_h$</td>
<td>heat transfer coefficient</td>
<td>$1.3 \times 10^{-3}$</td>
</tr>
<tr>
<td>$c_a$</td>
<td>latent heat transfer coefficient</td>
<td>$1.3 \times 10^{-3}$</td>
</tr>
<tr>
<td>$k_e$, W(mK)$^{-1}$</td>
<td>heat conductivity of ice</td>
<td>2.17</td>
</tr>
<tr>
<td>$f_{cm}$, W(mK)$^{-1}$</td>
<td>heat conductivity of snow</td>
<td>0.31</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>albedo of open water</td>
<td>0.1</td>
</tr>
<tr>
<td>albedo of melting ice without snow</td>
<td>0.66</td>
<td></td>
</tr>
<tr>
<td>albedo of ice without snow</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>albedo of melting snow</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>albedo of snow</td>
<td>0.8</td>
<td></td>
</tr>
</tbody>
</table>

[Harder and Fischer, 1999]. Several model test runs have shown that both the sea surface tilt and the acceleration term can be neglected for timescales from hours to days.

The oceanic stress $\tau_w$ is parameterized quadratically, as commonly used for the Weddell Sea region, as

$$\tau_w = \rho_w c_w (u_w - u)(u_w - u),$$

with $\rho_w$ being the density of sea water, $c_w$ the oceanic drag coefficient, $u_w$ the total ocean current (sum of tidal and annual mean currents (see section 2.3) in our application) and $u$ the ice drift velocity (Table 1). Atmospheric stress is defined analogously.

The ice concentration is calculated from a continuity equation with a source term $S_{th}$ accounting for thermodynamic melting and freezing, plus an additional source term $S_d$ for contributions of dynamic processes. The thermodynamic source term is

$$S_{th} = \frac{1 - A}{h_0} \frac{\partial h_f}{\partial t} + \frac{A}{10h} \frac{\partial h_m}{\partial t},$$

where the freezing and melting rates are $\partial h_f/\partial t$ and $\partial h_m/\partial t$, respectively (see below). The empirical parameter $h_0 = 0.5$ m weights the fraction of ice being formed in open water versus the formation at the bottom of existing floes [Hibler, 1979]. The factor $A(10h)$ describes the potentially significant assumption that 1/10 of the melting occurs at the sides of the floes. The dynamic contributions are given by

$$S_d = \frac{1}{2} \left( \sqrt{\epsilon^2 + e^{-2\epsilon^2}} - |\nabla \cdot u| \right) e^{-C(1-A)},$$

where $\epsilon$ is the deformation rate tensor in its principal axis representation and $e$ is the eccentricity of the elliptical yield curve. The norm $\sqrt{\epsilon^2 + e^{-2\epsilon^2}}$ describes changes of the ice concentration by deformation, while $|\nabla \cdot u|$ expresses changes that occur through divergence. The exponential factor places larger weight on dynamic processes for high ice concentrations. This source term leads to a different behavior in cases of convergent and divergent drift.

The thermodynamic module assumes seven evenly distributed ice thickness classes and one class representing open water. For each ice class the thermodynamical evolution is calculated independently of all other classes in each time step. Except for the surface temperatures of each class, which are memorized for the next time step, all other properties are averaged to produce a mean thickness ice class used in the dynamic calculations (see above). Freezing and melting rates are derived from the energy balance of the mixed layer according to Parkinson and Washington [1979]. The change in energy of the mixed layer is related to the sum of the net energy fluxes at the upper and lower margins. In the presence of sea ice, the mixed layer temperature remains at the freezing point of sea water and the freezing rate of sea ice is derived from the net energy exchange with the atmosphere and the mixed layer base.

The oceanic heat flux is derived from the two-layer boundary layer model of Lemke [1987], based on the work of Kraus and Turner [1967]. It treats the base of the mixed layer as a dynamic boundary and prognostically calculates entrainment of heat and salt from the water mass underneath. The thickness of the mixed layer and the entrainment fluxes depend on the mechanical energy budget, i.e., on stability, turbulence, and current velocity at the surface, thus also allowing for tidally induced upwelling.

Wind stress, oceanic stress and internal forces due to deformation are the three most important forces controlling the ice drift. In our sea ice representation, all three involve empirical parameters: atmospheric and oceanic drag coefficients and ice strength. Only a few measurements of these parameters are available [Mau et al., 1996; Andreas, 1995; McPhee et al., 1999; Richter-Menge et al., 2002]. In particular, ice strength cannot be determined directly. The free parameters are usually chosen through variation of the parameters and comparison of the results with observed sea ice data. We adopt the values of the different parameters from previous model versions, summarized in Table 1. Possible effects of the spatial grid size on the parameter values have been neglected.

The model domain comprises the Weddell Sea region from 78.5°S to 52°S and from 69°W to 3°W, with a meridional and zonal resolution of 0.1° and 0.3°, respectively. To assure numerical stability according to the Courant-Friedrich-Levy criterion, the time step has to be reduced to 2 hours. The boundaries are chosen such that the error from open boundary conditions is minimized. The maximum winter ice extent is 55°S, thus there is no meridional ice advection across the northern boundary. Across-boundary advection has been minimized by choosing the Weddell Sea as the modelling domain. The mask distinguishing grid points on land and sea is interpolated from the coastline data set of the GEBCO Digital Atlas of the British Oceanographic Data Centre. At all boundaries not specified by land, the ice thickness of inflow is assumed to be the same as the ice thickness present at the first grid point. Boundary grid points just outside the model domain are assigned a ice thickness which corre-
sponds to an averaged over several neighboring grid points inside the model domain.

### 2.2. Barotropic Tidal Model

[20] The barotropic, ocean-only, Circum-Antarctic Tidal Simulation, version 0.02 (CATS00.2) that we use here is an updated version of the CATS model used by Padman and Kottmeier [2000] and Rignot et al. [2000]. The governing equations (depth-integrated momentum equations plus continuity) are presented by Robertson et al. [1998]. The CATS00.2 domain includes the entire ocean south of 56°S. For the sea ice model region outside the CATS domain, between 52°S and 56°S, tides were set to zero. (As only a very small fraction of this area is ice covered in winter, the influence on the results is completely negligible.) The grid spacing is 1/4° × 1/12°, which gives ~10 km resolution near the Antarctic coast. For the ocean cavities under the ice shelves, water depth is replaced by “water column thickness,” i.e., the vertical distance between the base of the glacial ice and the seabed. The water column thickness values for the Filchner-Ronne Ice Shelf region are obtained from thickness values for the Filchner-Ronne Ice Shelf region are obtained from Johnson and Smith [1997]. For all other ice shelves in the Weddell Sea the water depth is taken unmodified from ETOPO-5 since very few glacier thickness data are available. The model incorporates depth data for the southwestern Weddell Sea that were acquired during the 1998 Ronne Polynya Experiment (ROPEX-98) [Nicholls et al., 2003].

[21] Four diurnal (O1, K1, P1, and Q1) and four semiannual (M2, S2, N2, and K2) tidal constituents are simulated by CATS00.2. These are the primary constituents in the global tidal model (TPXO.5) [Egbert, 1997] that we use for the open boundary condition along 56°S. Model forcing consists of TPXO.5 height fields along the northern boundary and direct astronomical forcing. The CATS00.2 model is run nonlinearly: the model solves the shallow water equations by forward time stepping with boundary conditions specified as the superposition of all eight constituents. However, as nonlinear effects are generally small, the prediction of tides using just the summation of the 8 major tides is sufficiently accurate for the purpose of this study. The amplitudes and phases of the major tides are determined by analyses of the last 180 days of the model run, using methods following Foreman [1977, 1978], i.e., a least squares fit of phase-shifted sine waves with the frequencies of the eight primary constituents. Sea ice is not included in CATS00.2 and there is no feedback from the sea ice model to the barotropic tidal model.

### 2.3. Forcing of the Sea Ice Model

[22] Major external factors influencing the sea ice model are wind, air temperature, ocean mean currents and tidal currents. These enter via time-dependent forcing fields, interpolated to the model resolution using cubic splines. [23] Data from the Reanalysis Project of the European Centre for Medium-Range Weather Forecasts (ECMWF) prescribe time-dependent atmospheric forcing fields. The ECMWF output includes air temperature at a height of 2 m and surface winds at a height of 10 m [European Centre for Medium-Range Weather Forecasts, 1992]. The ECMWF output data are available at 1.125° resolution every 6 hours. The relatively high spatial resolution of ECMWF is the major reason for our use of this data set for atmospheric forcing. The ECMWF data are linearly interpolated between the 6 hour intervals to adjust to the 2 hour time step of our model. Humidity at a height of 2 m and precipitation data are derived from ECMWF output as climatological monthly means [Harder, 1996]. Monthly mean cloud cover at a resolution of 1.5° is calculated from the data of the International Cloud Climatology Project (ISCCP) [Rasov and Garder, 1993].

[24] Time-dependent oceanic stress is calculated from the ice relative current, which is the sum of the annual mean background current and time-varying tidal currents, minus the ice drift velocity. Annual means of surface currents derived from the OCCAM global circulation model [Webb et al., 1997] specify the invariant part of the currents in the water drag term of the sea ice model. The output of the CATS00.2 tidal simulation is used to specify tidal current data at 2 hour intervals. Numerical stability for a time step of 2 hours requires that the maximum tidal currents (Figure 1) are limited to 0.6 m s⁻¹. Only 0.3% of all grid point values (at all times) of the instantaneous current velocity exceed this velocity limit, only 0.06% exceed 1 m s⁻¹. Model runs with different threshold velocities have shown that the sea ice is insensitive to this cutoff.

[25] Temperature and salinity of the water mass at the lower boundary of the mixed layer module are required as input to compute the rate of entrainment at the base of the mixed layer. They are derived from data of the Hydrographic Atlas of the Southern Ocean (SOA) at a depth of 250 m [Olbers et al., 1992]. From these two-dimensional data fields temperature and salinity at the depth of the lower boundary of the mixed layer, which rarely exceeds 150 m, are calculated.

### 2.4. Concept for Model Runs

[26] The sea ice model calculates time-dependent fields of ice thickness, ice concentration, net ice growth, ice drift velocities and various energy fluxes. To evaluate the difference in outputs from the two different types of model runs on a regional scale, monthly means of these ice properties are calculated from the 2 hourly model output. The difference between monthly means of the two runs, with and without tidal forcing, then provides information about differences in the evolution of the sea ice cover linked to tidal currents.

[27] The initial condition for all sea ice model runs is an ice-free ocean. A cyclostationary ice cover is reached by calculating a 7 year spin-up period forced with 1995 atmospheric data. To investigate the influence of tidal currents on the sea ice cover, two types of model runs are performed: a biennial run under atmospheric forcing conditions from 1995 and 1996 including tidal currents, and a biannual run with the same atmospheric forcing but with no tidal currents. Mean annual ocean currents [Webb et al., 1997] (see above) are included in all runs.

[28] The modifications applied to the original model version, like increased temporal and spatial resolution, extended interaction with other modules, etc., should still enable the full functionality of the model. We demonstrated the reliability of the high-resolution sea ice model by conducting a control run with the circum-Antarctic version of the model [Pohlmann et al., 2000], with a resolution of...
only 1/5 of our sea ice model, for a 1 year run with 1995 forcing [Köntopp, 2000]. The results of the control run are generally in good agreement with our model output without tidal forcing. Both models predict a similar annual cycle for the ice cover in the Weddell Sea. The general structure of the ice cover and ice extent in both models only show small differences. These are an increased maximum ice extent farther to the north and slightly higher ice thickness in the low-resolution circum-Antarctic version, which can be attributed to differences in forcing fields. Nevertheless, the overall agreement indicates that our sea ice model provides reasonable results and can therefore be used for the proposed studies.

3. Regional Results

3.1. Local Evolution

3.1.1. Tidal Influence on Drift

[30] At first we investigate time series of different quantities at a grid point in a tidally active region to gain insights into the local effects of tidal forcing on drift, related energy fluxes, and the development of ice thickness. The subsequent presentation of regional results for the Weddell Sea focuses on ice drift, ice concentration, ice thickness, and ice formation rate\(^1\). The role of the different processes responsible for the interaction between these quantities will be analyzed in more detail in the discussion (section 4).

Figure 1. Study area with 1000 m bathymetry intervals (shaded contours) and maximum tidal current velocity (color code) for the year 1996 calculated by the Circum-Antarctic Tidal Simulation (CATS) barotropic tidal model in m s\(^{-1}\). Outside the CATS region north of 56°S, tides are set to zero. Included tidal constituents are \(O_1\), \(K_1\), \(P_{1}\), \(Q_{1}\), \(M_{2}\), \(S_2\), \(N_2\), and \(K_2\). For easier referencing we define five regions: the continental shelves in front of the Antarctic Peninsula (APS) and the Filchner-Ronne Shelf (FRS), the shelf break (SB), stretching from the southeastern Weddell toward the Peninsula, then northward and, finally, to the east, the Dronning Maud Land coast (DMLC), and the central Weddell Sea (central WS). Furthermore, we will use the South Shetland Islands (61°S, 55°W), the South Orkney Islands (61°S, 45°W), and the South Sandwich Islands (56°S, 26°W) for location of features described later.

Tidal effects on the energy exchange from ocean to atmosphere and on the formation of new ice have been estimated in different studies for nearby regions [Lytle and Ackley, 1996; Eisen and Kottmeier, 2000; Padman and Kottmeier, 2000]. The additional net wintertime oceanic heat loss in tidally dominated regions is 4–10 Wm$^{-2}$ [Padman and Kottmeier, 2000], caused by the induced periodic divergence and lead formation. Formation of new ice can be increased on aerial average by up to 12%, and is associated with an increased salt rejection of some 23% in the western Weddell Sea [Eisen and Kottmeier, 2000]. For the grid point and period under consideration here, the tidal model run generally provides increased instant atmospheric fluxes of 2–30 Wm$^{-2}$ (8 Wm$^{-2}$ on temporal average), primarily due to lower ice concentrations than in the no tides model run. Obviously, pronounced tidally induced fluctuations of ice concentration during days 179–184 lead to a large local increase in heat loss from ocean to atmosphere, resulting in amplified local ice growth rates (becoming less negative to positive). Especially during this period the strong oscillations in ice concentration are also reflected in the ice growth rate, with the same period of $\sim$1 day and a magnitude of some 0.5 mm hr$^{-1}$.

The short local description given above provides guidance to the interaction and relevance of different processes coming along with tidal motion. Basin-wide estimates of important sea ice properties will be presented next.

### 3.2. Ice Drift

In general, the dominant feature of the ice drift matches the cyclonic Weddell Gyre in both model runs, with strongest monthly mean velocities up to 0.25 m s$^{-1}$, northward advection along the APS, and parallel drift along the southeastern Weddell Sea coast (Figure 3). The drift is strongest during the austral winter, when ECMWF 10 m wind analyses show stronger monthly mean winds and more frequent and more intense cyclones [Kottmeier et al., 1997]. The stronger wind motion can overcome the larger ice internal stress forces resulting from thicker and more compact ice. In tidally active areas, the high-frequency ice motion can include a significant component of semidiurnal and diurnal variability, although the ability of the ice to respond to ocean tidal currents depends strongly on the ice concentration and so has a seasonal component to it [Padman and Kottmeier, 2000].

Under conditions of new ice formation in winter, strong tidal currents result in an asymmetric response of the sea ice cover to applied stresses (both atmospheric and oceanic). When ice motion is divergent the free drift approximation is valid. During periods of convergence, however, internal stresses additionally affect ice motion, resulting in smaller velocities and deformation of the ice cover. Because of this asymmetric response to stress, one might expect some rectification of the tidal forcing, and thus a significant change in mean ice drift between the runs without and with tides. Our results indicate that this seems not to be the case. The effect of tidal forcing on the monthly mean ice drift is very small. In the western Weddell Sea near the APS-FRS-SB junction, where the mean ice drift is strong and modelled tidal velocities are up to 0.5 m s$^{-1}$ year round (Figure 1), maximum differences are below 0.03 m s$^{-1}$. Elsewhere, differences in the drift speed between the model runs usually stay well below 0.01 m s$^{-1}$. Owing to the monthly averaging submonthly modulations of the tidal motions, e.g., the spring neap variability, are smoothed out in our results of ice drift. However, their cumulative effect on other parameters like ice growth still remains and leaves a mark especially on shorter timescales (see Figure 2).

### 3.3. Ice Concentration

The annual cycle of ice extent is modified by tidal forcing. In winter, the model predicts a closed ice cover with ice concentrations near 100% and only minor differences
between both runs. (Ice concentration values referred to in the text and displayed in the figures are determined after the thermodynamic calculations are completed and before the next cycle of dynamic calculations sets in.) Both simulations calculate the same maximum ice extent in September, with the ice edge being $\sim 61^\circ$S at the Antarctic Peninsula and $\sim 57^\circ$S in the eastern Weddell Sea (Figure 4).

The similar stable equilibrium reached in both runs by September is most likely related to a strong and closed ice cover, which mostly suppresses tidal effects. At this time, concentration differences stay in general below 1%. Exceptions are the marginal ice zone, especially from the APS along the South Shetland Islands toward the South Orkney Islands. Here, tidal influence reduces ice concentrations by more than 25% (see auxiliary material 01). A second feature of slightly reduced ice concentrations, which is most developed in September, extends from the South Orkney Islands over a width of 200–300 km in an arc toward the APS and then south along the coast, with ice concentrations reduced by up to 15%, and then continues with differences of some $-3$ to $-5$% eastward in the FRS region around $73^\circ$S.

Excluding winter months, tidal currents reduce the ice concentration in large areas by 3–10% on average. In a region along the coast in the southeastern Weddell Sea, a large coastal polynya develops in both runs in November, extending from $-30^\circ$W to the eastern model boundary. In austral spring, differences in the melting behavior of the two runs become visible in the area of maximum ice extent. Whereas the tidal influence causes a retardation of the retreat by 2$^\circ$ to the east in December, it speeds up the retreat in the northwest. After the ice edge reaches the shelf break in January, tides cause an overall faster retreat of the ice cover which is especially pronounced in tidally active regions.

The minimum ice extent occurs in February and March. When tides are included, the ice covers an area of 593 km$^2$. Without tides, the remaining ice cover in the western Weddell Sea in summer is about 12% larger, comprising 665 km$^2$, and extends $\sim 1^\circ$–2$^\circ$ farther to the east than in the run with tides. Ice concentrations in both runs are around 90% in most of the remaining area, although the fraction of the area with $\sim 100$% ice concentration is lower in the run with tides. These differences are attributed to the faster ice retreat in spring.

During the onset of sea ice formation, ice production rates in the eastern and northwestern Weddell Sea are significantly reduced by tides. Two separated areas with ice concentration differences below $-20$% develop out of the area of reduced ice concentrations on the eastern side of the remaining ice cover. One appears off the DMLC in April and extends toward the northeast, increasing in size (Figure 5). In June, it reaches the Weddell-Enderby Basin when maximum differences are still in excess of $-20$%, centered around $64^\circ$S, $5^\circ$W. It finally disappears in July. The second strong feature is present along the western shelf break north of $70^\circ$S along $50^\circ$W and moves northward, reaching the South Orkney Islands in June and blends into general lower ice concentrations along the marginal ice zone (see auxiliary material 03). The movement of both areas of lower ice concentrations is accompanied by higher heat loss to the atmosphere in the tidal run. During May and June, in large parts of the eastern and northern Weddell Sea ice concentrations are reduced by tidal effects by 3–5%. In the central and the southern Weddell Sea concentration differences remain small (Figure 5).

### 3.4. Ice Thickness

The effects of tidal currents on the regional sea ice thickness distribution tend to be more pronounced than for
ice concentration and drift. (Here, ice thickness is defined as the height of the ice layer spread evenly over the grid cell, i.e., the grid cell mean ice thickness.) Generally, in both model runs the ice is thickest in the southwestern Weddell Sea near the Antarctic Peninsula and along the Filchner-Ronne Ice Shelf, ranging from \( \sim 1.5 \) m to \( \geq 3 \) m close to the coast throughout the year. In winter, ice thickness gradually decreases to 0.4–0.6 m toward the eastern model boundary and varies between 0.8–1.5 m in the central Weddell Sea (Figure 6).

A clear effect of tidal forcing on the mean ice thickness coincides with the region of reduced ice concentration along the shelf break and in the marginal ice zone in the Bransfield Strait south of the South Shetland Islands, as described above. A band of higher ice thickness in the tidal run extends from the eastern side of the Antarctic Peninsula at 70°S to the southeast along the FRS toward 75°S, 25°W at the DMLC (Figure 7). Differences in this region become apparent in May, grow up to 0.3 m, and decay again in February (see auxiliary material 06). Along the Antarctic Peninsula this feature is closer to the coast than the region of strongest tides, which is located roughly above the shelf break about 5° (\( \sim 150 \) km) off the coast. Maximum differences in ice thickness in this area reach up to 0.5 m in October and finally disappear in February.

Apart from this persistent band the ice thickness differences undergo clear seasonal variations in extent and intensity. During summer and early fall a strong coastal anomaly is seen along the northern Antarctic Peninsula from about 70°S to 63°S. The development of this stage starts in October, when the region of thicker ice is established around 70°S. During summer it extends and progresses northward, passing the northern tip of the Peninsula in May. From there it continues toward the east, reaching the South Orkney Islands around September, and disappearing in November near the South Sandwich Islands. South of \( \sim 65° \)S, the ice thickness anomaly starts to decrease in June to values below 0.15 m. The direction and velocity of the progression of this thickness anomaly is in agreement with the general ice circulation of the Weddell Gyre. This suggests that advection by the mean ice drift might contribute to the spatial shift of the pattern.
A second prominent feature of the thickness difference is the highly reduced ice thickness in the southern Weddell Sea along the southern Peninsula and the FRS. Tidal forcing in this region leads to a decrease of some 0.2–0.5 m in a large area throughout the whole year. Seasonal changes are limited and mostly visible in the extent of the negative anomaly, which covers a minimum area in spring and reaches the maximum in fall, obviously out-of-phase with the other two features representing thicker ice. The negative thickness anomaly is probably a dynamic effect caused by interaction of tidal currents with the Filchner-Ronne Ice Shelf. Tidal currents lead to an increased northward flow from underneath the ice shelf and export ice produced in this area farther north. As also known from other areas along the Filchner-Ronne Ice Shelf northward flow leads to enhanced formation of polynyas and thus reduced ice concentration, as well as a reduced ice thickness close to the ice shelf front.

In the remaining parts of the Weddell Sea, differences are generally small throughout most of the year, varying from −0.05 m to +0.05 m. A seasonal exception is the southern summer, when the stronger melting in the tidal run leads to a reduction by 0.05–0.15 m in most of the remaining ice-covered areas. In early fall, differences in the eastern Weddell Sea with up to 0.3 m thinner ice in the tidal run can be tracked back to larger melting and a reduced extent during summer, both leading to a delayed spreading of the ice cover as discussed for the ice concentration above.

### 3.5. Net Freezing Rate

The net freezing rate is the seasonally averaged thermodynamic ice production, a useful representation of the effects of sea ice development on the local salt distribution of the upper ocean. The regions of net freezing predicted by our model in the southern Weddell Sea and near the coast are associated with salt rejection and thus densification of the surface waters, while calculated net melting in the northern Weddell Sea supplies fresh water to the ocean.

The map of differences of net freezing rates between the tidal and nontidal runs is very patchy, especially in tidally energetic regions and in the marginal ice zone (Figure 8). Most features reflect differences that originate at the ice margins during expansion and decline of the ice cover in previous months and are subsequently conserved in the yearly signal because of their strength.

Over the central Weddell Sea the pattern consists of many small-scale features, with differences around 100 cm a⁻¹, corresponding to a relative change in ice production of ±5% (Figure 9). A clear net tidal effect on a regional scale with increased ice formation exists in the FRS-APS region south of 70°S (Figure 8). Here, the location of a band of higher differences of up to 400 cm a⁻¹ is comparable to the features representing lower ice concentrations and especially higher thicknesses described above. The most severe relative change in ice production, however, with an increase of more than 100%, occurs in the tidally active region around 65°S, 52°W (Figure 9). While the absolute ice production in this area is comparable to the FRS-APS region, the generally small ice growth rates make this area very sensitive to tidally induced changes, though small on an absolute scale. Along the DMLC, ice production is reduced when tides are included, because of a reduced ice formation in fall and winter and increased melt in spring. North of the Antarctic Peninsula, a strip about 100 km wide of
increased ice formation ($\sim+300$ cm a$^{-1}$) over the continental shelf in the Bransfield Strait is adjacent to a band of equal width of reduced freezing (approximately $-200$ cm a$^{-1}$) toward the shelf break and the deeper ocean. The strong contrast in ice growth is probably related to the interaction of tidal currents with the Antarctic Circumpolar Current, which approaches the shelf in this area. This feature is also related to the differences in ice extent and thickness of the tides versus no tides run presented above.

Keeping in mind that these values represent the additional local ice growth due to tides, it is evident that the impact on the actual ice thickness differences are not as dramatic (Figure 7). The maximum values of 100–400 cm a$^{-1}$ are comparable to the known ice production in areas of frequently open water, such as the Ronne Polynya. Other factors, like advection, are also taken into account for the development of the total ice thickness.

4. Discussion

The interplay of local processes and advection leads to strong modifications in the spatial distribution of sea ice properties by tidal forcing. The dominant local process is the reduction of the mean ice concentration in tidally active areas by increased divergent motion. A direct consequence of lower mean ice concentration in the presence of tides is an altered ice-ocean interaction via the atmospheric heat flux changing the energy made available in the mixed layer. Likewise, surface currents change the mixed layer depth and can cause entrainment of heat from below. All of these processes act during different seasons in various ways on the sea ice cover and cause the observed anomalies, the relation of which we try to separate now.

Periodic currents cause short time fluctuations of up to 5% in ice concentration, with periods of $\sim1$ day, in extreme cases even more than 20%. Reduced ice concentrations, manifested as small leads formed by tidal currents, are responsible for an accelerated melting process in spring and a retarded expansion of the ice cover in early fall. What causes the movement and expansion of anomalies observed in most properties? One possibility is that the sea ice cover is capable of “memorizing” imprinted properties, which are then advected with the mean ice drift over large distances. Or are the anomalies caused instantaneously on the spot by local processes?

In the western Weddell Sea, ice concentration and thickness anomalies move along the flow direction of the Weddell Gyre northward along the APS and finally to the South Sandwich Islands. In the central Weddell Sea and along the DMLC, anomalies in concentration and heat flux move with the marginal ice zone, caused by an expanding sea ice cover in fall and winter. What causes the movement and expansion of anomalies observed in most properties? One possibility is that the sea ice cover is capable of “memorizing” imprinted properties, which are then advected with the mean ice drift over large distances. Or are the anomalies caused instantaneously on the spot by local processes?

The combination of lower ice concentration, i.e., mainly enhanced absorption of radiation, and additional entrainment of warmer waters into the mixed layer by tidal currents, are responsible for an accelerated melting process in spring and a retarded expansion of the ice cover in early fall. What causes the movement and expansion of anomalies observed in most properties? One possibility is that the sea ice cover is capable of “memorizing” imprinted properties, which are then advected with the mean ice drift over large distances. Or are the anomalies caused instantaneously on the spot by local processes?
tide-induced anomalies is related to a combination of ice drift velocity and the propagation velocity of the ice margin. Their northward propagation and eastward turning is first supported by the Weddell Gyre and then by the Antarctic Circumpolar Current. The observed propagation velocity of anomalies in this region is too rapid to be accounted for solely by passive advection of the tidally modified ice properties. The propagation of ice concentration anomalies observed along the DMLC, however, are not supported by currents. On the contrary, they even progress against the mean current. They can therefore only be caused by local processes, and not by processes occurring during minimum ice extent. Considering the location of the DMLC anomalies over time, they are related to an interplay of local tidal influences near the shelf break and the marginal ice zone. With expanding ice cover toward the northeast, lower ice concentrations in the tidal run appear in the area where the shelf break intersects with the marginal ice zone, i.e., where tidal currents are strong and the sea ice cover is still weak. The position of the anomalies along the APS are also coincident with the shelf break and the marginal ice zone. Therefore the same interaction between tidal currents and the marginal ice zone is also likely responsible for the anomaly propagating northward along the APS, however supported by mean currents, and not the conditions imprinted in the sea ice cover during minimum ice extent.

The present study demonstrates the importance of including tides in models of Antarctic sea ice. The model we have used here does not, however, fully couple the ocean to the changing ice properties. Some small but possibly nonnegligible change in ocean tides might be expected from the added friction at the ocean-ice interface [see, e.g., Kowalik and Proshutinsky, 1994]. The simplified mixed layer parameterization, by which the heat from the warmer water underneath the mixed layer is made available to the ice cover, is also a potential source of error. The most significant limitation of our model approach, however, is the use of only one ice thickness class in respect to rheology and dynamic behavior. Although the thermodynamic module considers several thickness classes, averaging properties of all classes to obtain a single mean thickness class likely has some influence on the behavior ice cover during convergence, especially on the formation of pressure ridges and their distribution. Nevertheless, even with an improved treatment of multicategory ice dynamics the principal mechanisms will likely remain the same and produce results in comparison to model runs without tides which are at least qualitatively on the same order of magnitude.

Our use of a barotropic tidal model (CATS00.2) excludes the potential for baroclinic tides to contribute to the influence of tides on ice properties. Muench et al. [2002] suggested that tides along the south Scotia Ridge (the northern boundary of the Weddell Sea) could provide significant additional heat flux from the Warm Deep Water to the surface layer and ice cover. A recent model of baroclinic tides for the northern Weddell Sea and Scotia Sea (S. Howard, personal communication, 2004) indicates that the root-mean-square divergence of ice-water stress associated with near-surface baroclinic tidal currents could be considerably higher than for barotropic-only currents. That is, the true influence of tides on ice properties may be even greater than our model, forced with barotropic tides, suggests.

5. Conclusions

On short timescales on the order of hours and days, tides lead to highly variable heat fluxes, sea ice concentration, ice production rate, and ice thickness. Tidal currents also cause significantly different distributions of these properties on small spatial scales even when considering monthly means. Key regional characteristics of the sea ice cover, like maximum ice extent and ice volume, are also altered by the tidal influence, however, to a much smaller degree than local characteristics. The addition of tidal forcing does significantly modify the temporal development of the expansion and retreat of the sea ice cover, and needs to be considered in an operational model of sea ice distribution. Especially in regions showing large differences, tidal forcing could also influence the temporal evolution of regional and larger-scale coupled model outputs.

It is, furthermore, likely that the addition of baroclinicity to the ocean tidal forcing will lead to significantly larger tide-induced signals in sea ice properties. If this is true, tidal effects on ice will be important not only operationally but also in regional and larger-scale coupled ocean-ice-atmosphere climate models. Thus although there is a need for significant improvements in our ice model and the coupling with the upper ocean, developing the skills to routinely including small-scale ocean forcing in sea ice predictive models is clearly worthy of further effort.

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