Internal tide generation along the South Scotia Ridge

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Received 8 April 2005; accepted 19 July 2005

Abstract

The generation of semidiurnal internal tides along the South Scotia Ridge (SSR)—the boundary between the northern Weddell Sea and the Scotia Sea—is investigated using a fully three-dimensional primitive equation numerical model. The sensitivity of the model to changes in horizontal and vertical grid spacing, stratification, and choice of bathymetric data set is discussed. The highest resolution model (1/16° × 1/25° × 41 levels) with realistic stratification produces an average along-ridge lateral baroclinic $M_2$ energy flux magnitude $|J(M_2)| \approx 160 \text{ W m}^{-1}$ for the section of the SSR between the tip of the Antarctic Peninsula and the South Orkney Plateau. This value is <1% of the flux obtained with a lower-resolution global baroclinic tides model run with stratification representative of mid-latitudes (Simmons et al., Internal wave generation in a global baroclinic tide model. 2004. Deep-Sea Research II 51, 3043–3068). Our results indicate that baroclinic tidal energy will not be an important contributor to mixing in the pycnocline in the SSR region. However, the addition of baroclinic tides significantly increases the predicted root-mean-square horizontal divergence of the surface tidal current fields ($\sigma(V_H \cdot u)$), with consequences for the annual cycle of sea-ice formation and melt and the upper-ocean thermohaline structure. Peak values of $\sigma(V_H \cdot u)$ along the ridge, for $M_2$ only, correspond to a tidal contribution of >10% to time-averaged open-water fraction. Thus, baroclinic tides may play a significant role in setting mean heat and freshwater exchange and ice formation in the northern Weddell Sea.

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Keywords: Baroclinic tides; Tidal modeling; Turbulent mixing; Sea ice

1. Introduction

The South Scotia Ridge ("SSR") trends eastward from the tip of the Antarctic Peninsula and forms a leaky topographic boundary separating the northwestern Weddell Sea from the southern Scotia Sea (Fig. 1). Dense water formed by winter cooling and ice formation farther south on the southern and western Weddell Sea shelf must circulate around or over the SSR in order to enter the World Ocean's deep circulation. This region was the focus of the Deep Ocean Ventilation Through Antarctic Intermediate Layers (DOVETAIL) program, with field work conducted in the
late 1990s. Components of DOVETAIL and related studies are described in a set of papers in Deep-Sea Research, Part II, Volume 49(21).

Previous data-based and modeling studies for this region raised the possibility that tides might contribute to the thermohaline structure around the ridge. Two mechanisms have been proposed. First, it is now well known that the conversion of barotropic to baroclinic tides through tidal current interactions with steep and rough topography contributes a significant fraction of the total power available to mix the global ocean; see, for example, Munk and Wunsch (1998), Wunsch (2000), and Egbert and Ray (2000, 2001). Barotropic tidal currents have speeds of several centimeters per second over the SSR (Fig. 2), and the topography is steep; thus, it is reasonable to expect that locally generated internal tides might be energetic. Indeed, the results from a recent global baroclinic tide model (Simmons et al., 2004: hereinafter denoted SHA) suggest that the SSR might be a “hot spot” for internal tide generation; yet it is a region that has not been carefully studied. Second, recent results obtained from a sea-ice model coupled to ocean circulation, with and without the addition of barotropic tides, demonstrate that the spatial variability of tide-induced stress on the sea ice plays an important role in mean sea-ice properties and the associated exchanges of heat and fresh water between the ocean and atmosphere (Koentopp et al., 2005). This work supports an earlier study of sea-ice response to tides in the Arctic (Kowalik and Proshutinsky, 1994) in which it was shown that tides could increase annual averaged ice growth rates by up to 1 m a\(^{-1}\) in tidally active regions, mostly along the continental shelf break where tidal current speeds and phases change rapidly.

Our interest in these processes along the SSR stems from the desire to model the water mass transformation experienced by waters of Weddell Sea origin as they flow around the southern flanks of the ridge.
of the SSR before entering the World Ocean. Deeper water masses can contribute to the global inventory of Antarctic Bottom Water, while the interactions between the sea ice and surface layer determine the ultimate fate of the fresh water that is exported from the Weddell Sea under the influence of wind stress. Diapycnal mixing through the main pycnocline, centered near 200 m depth, provides a mechanism for fluxes of the underlying dense water properties into the surface layer, where surface Ekman dynamics and other processes such as mesoscale eddies can effect cross-ridge transport. Muench et al. (2002) estimated a typical diapycnal scalar diffusivity, based on a survey of CTD profiles and vessel-mounted acoustic Doppler current profiler (ADCP) data.
collected during the DOVETAIL program, of $K_v \approx 10^{-4}$ m$^2$ s$^{-1}$, resulting in an upward heat flux in the main pycnocline of $F_H \approx 4$ W m$^{-2}$. These values were broadly consistent with values estimated from analytical equations for internal tide generation (Morozov, 1995), applied to idealized ridge geometry, with dissipation occurring over an assumed baroclinic tide propagation length scale of $\sim 1000$ km. The recent global baroclinic tide modeling study by SHA suggested that baroclinic tides produced along the SSR might be much more energetic than estimated by Muench et al. (2002), raising the possibility that tide-induced mixing in the pycnocline might be very strong. One goal of the present study therefore is, to determine local rates of baroclinic tide generation along the SSR for comparison with the Muench et al. (2002) and SHA studies.

Our second goal is to determine whether the addition of baroclinic tides to coupled ocean/sea-ice models should significantly alter the influence of tides on the temporal evolution and spatial distribution of sea-ice properties. The SSR region is normally ice-covered during the austral winter. Tidally forced, periodically divergent stress at the ice/water interface leads to higher mean open-water (“lead”) fraction, with consequently greater rates of ocean-to-atmosphere heat exchange and ice formation (in fall and winter) and freshwater production through ice melt (in spring and summer). Lead fractions due to the periodic divergence of depth-averaged tidal currents (see, e.g., Padman and Kottmeier, 2000) can exceed 5% in tidally active areas such as those along the shelf break of the southern and western Weddell Sea. Heat loss through leads in winter exceeds the loss rate through adjacent multi-year, snow-covered ice by a factor of 10–100 (Wadhams, 2000). Thus, even tide-induced lead fractions of a few percent can significantly increase, and may sometimes dominate, the area-averaged ocean-to-atmosphere heat losses and subsequent ice production rates. The thermohaline and wind-forced circulation then distributes tide-induced thickness anomalies generated locally over a much broader region (Koentopp et al., 2005). We hypothesize that the addition of baroclinic tides would further increase the tide’s influence on sea-ice properties, by increasing maximum surface currents and decreasing the spatial scales of variability.

2. The baroclinic tide model

We use the Princeton Ocean Model (POM) (Blumberg and Mellor, 1987) for our analyses. Our application of POM is similar to studies of internal tide generation along the Hawaiian Ridge by Merrifield et al. (2001). POM is a three-dimensional, nonlinear, hydrostatic, primitive equation model which uses terrain-following (“sigma”) vertical coordinates. The model employs a simplified Mellor-Yamada 2.5 turbulence closure scheme (Mellor and Yamada, 1974, 1982) for vertical viscosity $\nu_z$. In our runs, we set the background (minimum) vertical viscosity of 0.2 m$^2$ s$^{-1}$. This parameterization is used for numerical stability rather than to provide realistic estimates of $\nu_z$ and $\nu_0$ is smaller than the mean value of $0.8 \times 10^{-4}$ m$^2$ s$^{-1}$ estimated for the Powell Basin main pycnocline by Muench et al. (2002). Vertical scalar diffusivity $K_z$ was set to zero so that the area-averaged density profile did not drift from its initial specified profile. Lateral mixing also was set to zero except in the open-boundary flow relaxation region (sponge layer), described below, where a value of 500 m$^2$ s$^{-1}$ was selected for horizontal viscosity $A_H$.

The model includes only ocean dynamics: there is no parameterization of sea-ice effects. The only significant ice shelf in the model domain is the Larsen C Ice Shelf on the eastern side of the Antarctic Peninsula, and this has been treated simply as a reduction of water depth associated with the ice draft. The generation rate for baroclinic tides near the Larsen C is not significant, nor does significant energy from this region propagate to the SSR to affect our primary results.

Several configurations of the model were run to explore model sensitivity to the bathymetry grid, model vertical and lateral resolution, and stratification. The parameter sets for models discussed in this paper are summarized in Table 1.

Our primary model domain covers a box from 58.12°S to 67.84°S and 65.8°W to 34.6°W, including sponge layers (Fig. 1). A smaller domain (with limits of 63.84°S to 58.12°S, and 55.5°W to
The large and small domains are indicated in Fig. 1. Normal-flow open boundary conditions (“OBC”) were obtained from the CATS02.01 (CATS) 1/4° × 1/12° model or the Antarctic Peninsula (AP) 1/30° × 1/60° model. Constituents (“Constit”) were either $M_2$-only or $M_2 + S_2$ (Run#8). Stratification (“Strat”) was either: PB, a mean profile for Powell Basin, from DOVETAIL measurements; or SHA, the global mean profile used by Simmons et al. (2004). Horizontal grid spacing (“Grid”) and the number of sigma levels (“Levels”) are indicated. Bathymetry (“Bathy”) was either: E5, ETOPO-5 modified according to Padman et al. (2002); or E2, ETOPO-2 with modifications to improve coastline accuracy around the Antarctic Peninsula. Run length ($R$) and analysis length ($A$) are in days. The along-ridge mean energy flux $\mathbf{J}$ (in kW m⁻²) for the $M_2$ constituent was evaluated as the sum of the mean northward and southward fluxes crossing the two diagonal dashed lines shown in Fig. 4 (lower panel). The mean energy flux divergence ($\langle \nabla \cdot \mathbf{J} \rangle$) for the $M_2$ constituent for the small box shown in Fig. 4 (lower panel) is in mW m⁻². The barotropic-to-baroclinic energy conversion rate ($C$) from the SHA model, averaged over the same small box, is also shown. The value of $\mathbf{J}$ based on the SHA model, and formulas of Morozov (1995) applied to idealized ridge topography, are included on the last two rows.

*Uses exactly the same bathymetry as in SHA (ETOPO-2, nearest-neighbor, not smoothed).

*Value shown is conversion rate $C$, for comparison with $\langle \nabla \cdot \mathbf{J} \rangle$. 

45.5°W) surrounding the central SSR, between the tip of the Antarctic Peninsula and the South Orkney Plateau, was used for computationally intense runs with multiple constituents and for tests of model sensitivity to different bathymetry grids. Horizontal grid spacing was chosen as either 1/8° × 1/8° (consistent with SHA), or 1/16° longitude × 1/25° latitude (mean spacing ~3–4 km). We used either 15 or 41 sigma levels, which were evenly spaced in the interior and more closely spaced in the benthic and surface boundary layers. The lower value provided roughly the same interior resolution as the SHA model. The higher value gives a typical vertical spacing in the interior of ~100 m in deep water and ~50 m over the ridge crests, where water depth is ~1500 m. Each model run was forced at all open boundaries with the component of the barotropic tidal current normal to the boundary. Normal-flow forcing produced depth-averaged current fields that were more consistent with our barotropic-only tide models than those obtained with elevation forcing, as determined by comparison between runs using the two methods. The complex amplitudes for the normal flow velocities were obtained from either the 1/4° × 1/12° Circum-Antarctic Tidal Simulation version 02.01 (CATS02.01) (Padman et al., 2002), or our newer high-resolution (1/60° × 1/30°) Antarctic Peninsula model, AntPen04.01. A sponge layer was used on the baroclinic components at all open boundaries, following Martinsen and Engedahl (1987). Baroclinic $u$ and $v$ velocities were reduced to zero over the outermost 10 grid points to prevent internal wave reflection at the boundaries.

Stratification was derived from DOVETAIL CTD measurements in Powell Basin, just south of the SSR (see Fig. 1), and was horizontally uniform. This condition misrepresents some spatial variability in barotropic-to-baroclinic energy conversion, but allows shorter runs since spatially varying stratification would require equilibration.
with geostrophic flows. The potential density profile $\sigma_0(z)$ (Fig. 3, upper left panel) consists of a thin surface layer above a sharp pycnocline centered near 200 m, overlying deep water that is only weakly stratified. In order to allow direct comparison with SHA, additional runs were made using their global-model stratification; see Fig. 2 in SHA.

The possible effects of bathymetry grids on the generation of baroclinic tides were explored by running models using either a modified form of ETOPO-5 bathymetry (representing a fairly smooth dataset) or ETOPO-2 bathymetry (showing rougher topography). For Run#1 (see Table 1), we used exactly the same bathymetry as used by SHA; a nearest-neighbor interpolation from the 2-min ETOPO-2 data onto the $1/8^\circ \times 1/8^\circ$ SHA grid. For all other runs we first smoothed ETOPO-2 (or ETOPO-5) to the resolution of our model grid. As SHA discussed in their Appendix A.2.3, baroclinic generation rate is sensitive to bathymetric roughness, and nearest-neighbor interpolation preserves the roughness of the originating high-resolution grid much better than filtering to the model grid.

For the single-constituent $M_2$ runs on the larger domain, the model was usually run for 15 days with tidal analyses applied to the last 5 days. The exception was for two runs (Run#1 and Run#2) with SHA-based grid spacing and stratification: these runs required $\sim$50 days for domain-integrated baroclinic kinetic energy to approach a stable value. For the $M_2 + S_2$ run on the smaller domain (Run#8), the model was run for 30 days,
with tidal analyses on the last 15 days in order to encompass one spring-neap cycle of the semidiurnal tides. We ignore the complicating effects of diurnal tides since diurnal-frequency waves are sub-inertial at this latitude. However, a recent study by Stashchuk and Vlasenko (2005) suggests that the nonlinear superposition of tidal harmonics might lead to much more energetic internal waves than would be expected from a simple linear sum of the generation due to each harmonic. Future studies will test whether the addition of diurnal tides to model runs has a significant impact on baroclinic wave generation around the SSR.

We use the modeled, depth-integrated baroclinic energy flux vector, $\mathbf{J}$ (W m$^{-1}$) and a spatial average of the energy flux divergence $\langle \nabla \cdot \mathbf{J} \rangle$ (W m$^{-2}$) to characterize the energetics of baroclinic tide propagation and generation. Values of $\mathbf{J}$ were calculated following the appendix in Holloway and Merrifield (1999). Values of $\langle \nabla \cdot \mathbf{J} \rangle$ were calculated as the line integral of the outward normal component of $\mathbf{J}$ around a closed box. These values can be compared to an area-average of the depth-integrated barotropic-to-baroclinic energy conversion rate, $C$ (W m$^{-2}$) described by SHA (their Eq. (12)). However, $\langle \nabla \cdot \mathbf{J} \rangle$ and $C$ are not identical since $\langle \nabla \cdot \mathbf{J} \rangle$ is modified by energy dissipation within the domain (associated in our model runs with $A_z$ and benthic stress).

3. Results

In the following, we refer to the total ($u_t$), barotropic ($u_{bt}$; depth-averaged), and “baroclinic” ($u_{bc} = u_t - u_{bt}$) velocities. Note that $u_{bc}$ includes not only true baroclinicity (i.e., the response to density stratification) but also frictional benthic-boundary-layer structure that would be present in some form regardless of stratification. Components of $u_t$ are represented by $u_t$ (positive eastwards) and $v_t$ (positive northwards); similarly for $u_{bt}$ and $u_{bc}$.

3.1. General structure of $M_2$ baroclinic velocities

Baroclinicity increases the maximum values of $U_{\text{mag}}(M_2)$ relative to the barotropic model values (Fig. 2), extends the region over which tidal currents are significant, and decreases the spatial scales of velocity gradients. The spatial complexity is actually much greater than suggested by Fig. 2 (lower panel), since the ellipse phase of $u_{bc}$ varies more rapidly than the magnitude. This is apparent is the rapid spatial variability of $v_t(z = 0)$ and $\nabla \cdot u_t(z = 0)$ shown in Fig. 3. The model determines amplitude and phase coefficients of the baroclinic tidal currents on the assumption that they are temporally constant for given ($x,y,z$). Baroclinic tides have long been viewed as highly incoherent in time and space (see, e.g., Wunsch, 1975), in response to the waves’ sensitivity to changing background conditions such as stratification and mean flows. Nevertheless, Ray and Mitchum (1996) found a strong, temporally coherent surface height signal in TOPEX/Poseidon altimetry, associated with low mode baroclinic tides radiating off the Hawaiian Ridge. Under the idealized conditions of our own models (time-invariant mean stratification and no mean flows), our modeled baroclinic tides are also coherent. We expect, however, that at a minimum there would be a seasonal cycle in baroclinic tide coefficients in response to the seasonal cycle in sea-ice and the density of the surface mixed layer.

A transect of the north component of $u_{bc}(M_2)$ across the SSR at a specific time towards the end of Run#6 illustrates some general features of internal tides in this region (Fig. 3, upper right). The wavelength of the $M_2$ internal tide in this region is about 100 km over deep water away from the ridge crests, consistent with our earlier ray-tracing study (Muench et al., 2002). Velocity shear in the main pycnocline is most intense near the ridge crests, but with some shear being found in the southern Scotia Sea and in Powell Basin.

3.2. Baroclinic $M_2$ lateral energy fluxes

The distribution of $\mathbf{J}(M_2)$ for Run#6 (Fig. 4) shows maximum magnitudes of $\sim 200$ W m$^{-1}$, with the largest values in a fairly small portion of the domain extending from $\sim 60^\circ$S (near a fracture ridge north of Bransfield Strait) to $\sim 47^\circ$W (the western end of the South Orkney Plateau). Bathymetry for this model was based on ETO-PO-5 interpolated onto the model’s finer grid. As
expected, fluxes are generally northward to the north of the SSR, and southward into the Powell Basin. The depth-integrated, baroclinic tidal energy flux associated with the central SSR, expressed as a mean value per unit length of the ridge, was calculated as the sum of the mean northward and southward fluxes crossing two lines, north and south of the SSR, respectively (see Fig. 4, lower panel, for locations of lines). The average value of $|J|$ for this section of ridge, for
Run#6, is \( \sim 50 \text{ W m}^{-1} \) northward and \( \sim 10 \text{ W m}^{-1} \) southward for a total of \( \sim 60 \text{ W m}^{-1} \). For ETOPO2-based bathymetry filtered to the resolution of the model grid (Run#7), the mean value of \( |\mathbf{J}| \) is \( \sim 110 \text{ W m}^{-3} \) northward and \( \sim 40 \text{ W m}^{-3} \) southward (\( \sim 150 \text{ W m}^{-3} \) total). The modeled values of \( |\mathbf{J}| \) are consistent with peak-to-peak isopycnal displacements of \( O(10) \text{ m} \) in the main pycnocline near 200 m (not shown). Larger displacements are found in the weakly stratified water below the pycnocline, although sometimes these displacements are barotropic in nature as a simple response to water-column compression and expansion during advection over steeply sloping topography.

The values of \( |\mathbf{J}| \) in Table 1 for our model runs identify the important factors affecting modeled baroclinic tidal energy fluxes. Changing the vertical grid spacing makes little difference to \( |\mathbf{J}| \) when SHA stratification is used; compare Run#3 with Run#4. (Run#3 is the same as Run#2, but the tidal analyses are applied to days 10–15 rather than 45–50.) For Powell Basin stratification, increasing both vertical and horizontal resolution significantly increases \( |\mathbf{J}| \); compare Run#5 with Run#6. Choice of stratification is also critical. Changing stratification from the realistic local profile used by us to the “global” profile used by SHA has a profound effect (factor of \( \sim 300 \)) on \( |\mathbf{J}| \); compare Run#3 with Run#5. Finally, the source of the bathymetric grid is important: the increase from \( |\mathbf{J}| \approx 0.06 \) to \( \sim 0.15 \text{ kW m}^{-1} \) from Run#6 to Run#7 is primarily due to the differences in large-scale ridge structure between the ETOPO-5 and ETOPO-2 bathymetric data sets. As discussed by SHA in their Appendix A.2.3, apart from the large-scale differences between gridded bathymetry data sets, it is also important to represent the small-scale roughness of the bathymetry. SHA maximize small-scale roughness, for their \( 1/8^\circ \times 1/8^\circ \) grid, by interpolating from ETOPO-2 using a nearest-neighbor approach rather than filtering to the spatial scales of the coarser model grid. In general, we filter bathymetry products such as ETOPO-2 to the same resolution as the model grid; this usually results in a grid for which the baroclinic pressure gradient errors in POM, resulting from sloping sigma-surfaces over topography, are negligible; see Beckmann and Haidvogel (1993) and Barnier et al. (1998). However, for Run#1 we used bathymetry exactly as constructed by SHA.

The effect of changed stratification (a factor of \( \sim 300 \) difference in \( |\mathbf{J}| \) between Run#3 and Run#5) is two orders of magnitude greater than predicted by the WKB-like scaling used by SHA (their Eqs. (14) and (15)) to account for spatial variations in stratification. One reason for this is apparent from a comparison plot of snapshots of the surface values of \( v_{bc} \) for Run#2 and Run#6 (Fig. 5). Changing the stratification alters not only the vertical distribution of baroclinic energy, but also the horizontal wavelength of the baroclinic waves. The change in wavelength leads to different patterns of both radiation and interference. For example, the Powell Basin is about 3 wavelengths wide for local stratification, but less than 1 wavelength across for SHA stratification. This leads to a distinctly different structure of baroclinic energy in the Basin resulting from interference of baroclinic waves generated along the SSR and from the ridge extending out from the tip of the Antarctic Peninsula along \( \sim 63^\circ \text{ S} \) in the northern Weddell Sea.

Run#1–Run#4 used horizontal and vertical grid spacings and stratification that are consistent with the SHA global baroclinic model; however, energy fluxes for these runs are only \( \sim 2\% \) of the values in the SHA model. The reasons for this discrepancy are not known, although some possibilities are listed in the discussion and conclusions section. For the moment, however, we note that the limited amount of oceanographic data collected during the DOVETAIL program (Muench et al., 2002) provide stronger support for our modeled fluxes (of order 0.1–1 kW m\(^{-1}\)) than for the \( \sim 45 \text{ kW m}^{-1} \) evaluated for the same section of ridge in the SHA model.

3.3. Spring-neap cycling in an \( M_2+S_2 \) model run

Model Run#8 solved for \( M_2 \) and \( S_2 \) open-boundary forcing for the small domain at high resolution (\( 1/16^\circ \times 1/25^\circ; \sim 4 \text{ km} \)). The magnitudes of the two constituents are very similar; hence, the barotropic forcing drops to almost zero at neap
tide ($t \approx 23.5$ days in Fig. 6, upper panel). The overall energy of the baroclinic mode, represented by the domain-integrated internal kinetic energy, is much less variable (Fig. 6, upper panel). The “age” of the baroclinic tide, i.e., the lag of the peak spring internal tidal energy relative to the peak spring barotropic energy, is $\sim 3$ days. This age is comparable to values reported by Holloway and Merrifield (2003) for the Hawaiian Ridge. These authors noted that the age increases away from the regions of generation, a direct response to the relatively slow group speed for baroclinic tides. We see a similar effect in our models: for example, the spring baroclinic tide at a point close to the SSR...
The divergence of tidal stress acting on the sea-ice (the so-called “ice accordion”) is essentially periodic, and can be estimated from our ocean-only models assuming no feedback from this stress to the ocean tidal currents. Padman and Kottmeier (2000) showed that, in the absence of significant compressive forces within the ice (i.e. mean ice concentration significantly less than 100% and/or thin ice), the ice velocity strongly resembles the underlying ocean velocity at tidal periods. Asymmetries in ice dynamics for compression and tension, and thermodynamic processes that favor sensitivity to ice thickness for thin ice types, increase the mean open-water (“lead”) fraction and the associated ocean-to-atmosphere heat flux and ice formation and melt rates (Kowalik and Proshutinsky, 1994; Geiger et al., 1998a,b; Padman and Kottmeier, 2000; Kwok et al., 2003; Koentopp et al., 2005; Geiger and Drinkwater, 2005). A useful measure of divergence for these studies is the root-mean-square (rms) of the current divergence (denoted \( \sigma(\nabla_H \cdot \mathbf{u}) \)) averaged over a tidal cycle. In the region of high \( |\nabla_H \cdot \mathbf{u}| \) (roughly, along-transect distance of 500–900 km in Fig. 3), \( \sigma(\nabla_H \cdot \mathbf{u}) \approx 10\sigma(\nabla_H \cdot \mathbf{u}_{bt}) \).

The rms divergence can be scaled to produce an estimate of the tidal contribution to the time-averaged open-water fraction \( (C_L) \) in pack ice. Following Padman and Kottmeier (2000), their Eq. (8), we calculate \( C_L = \pi \sigma(\nabla_H \cdot \mathbf{u}) / 2 \alpha \), where \( \alpha \) is the frequency of the tidal constituent (~1.4 × 10^{-4} s^{-1} for \( M_2 \)). Maps of \( C_L \) for \( \mathbf{u}_{bt} \) and \( \mathbf{u} \) (Fig. 7) demonstrate that high values of \( C_L \) (>10%, say) occur over a broad region surrounding the SSR, while significant values (>2%) are found throughout most of the Scotia Sea and northern Weddell Sea. When multiple tidal constituents are considered together, spring tidal divergence values may significantly exceed the values presented in Fig. 7.

4. Discussion and conclusions

The average magnitude of the total (northward plus southward) baroclinic energy flux \( \langle |J| \rangle \) along
the most tidally active sections of the SSR is \( \sim 150 \text{ W m}^{-1} \) for the \( M_2 \) harmonic in Run#7 and Run#8, the runs with the most realistic set of model parameters (high horizontal and vertical resolution, bathymetry based on ETOPO-2, and stratification obtained from Powell Basin CTD profiles). This value is \( \sim 1/3 \) of the \( \sim 500 \text{ W m}^{-1} \) obtained by Muench et al. (2002) using the Morozov (1995) analytical model applied to idealized ridge geometry for the \( M_2 \) tide, but is roughly consistent with the energy flux required to support the mixing rates inferred from CTD and

Fig. 7. Time-averaged open water (“lead”) percentage (100 \( \% \)) for the barotropic \( M_2 \) tide (upper panel) and the total surface \( M_2 \) current (barotropic + baroclinic; lower panel), based on the root-mean-square horizontal divergence \( \sigma(v_H \cdot u) \) from Run#6 (see Table 1 for model details). Black contours indicate the 1000, 2000, and 3000 m isobaths.
ADCP profiles collected in this region during the DOVETAIL program.

The \( M_2 \) baroclinic fluxes based on our own regional models with SHA parameters (coarser horizontal and vertical grid spacing, and SHA “global” stratification) are about six times higher, between 800 and 1070 W m\(^{-1} \) depending on choice of bathymetry grid. By comparing a range of runs with different grids, stratification and bathymetry, we conclude that \(|J|\) is most sensitive to stratification. By changing stratification from the global-average used by SHA to our own profile derived from Powell Basin CTD profiles, we reduced \(|J|\) by a factor of \( \sim 300 \). Improving grid resolution relative to SHA (from \( 1/8^\circ \times 1/8^\circ \times 15 \) levels, to \( 1/16^\circ \times 1/25^\circ \times 41 \) levels) increased \(|J|\) by a factor of \( \sim 20 \), while rougher and more recently compiled bathymetry (ETOPO-2 vs. ETOPO-5) further increased \(|J|\) by a factor of \( \sim 2–3 \). The divergence of \( \mathbf{J} \), averaged over a small region of energetic baroclinic tide generation along the central SSR (see Fig. 4, lower panel, for location), is \( \sim 1 \) mW m\(^{-2} \) for ETOPO-2 bathymetry (Run#7; see Table 1). This value can be interpreted, approximately, as the area-averaged barotropic-to-baroclinic generation rate (\( \langle C_S \rangle \)). For comparison, the global average value for \( M_2 \) is \( \sim 2 \) mW m\(^{-2} \) (\( \sim 700 \) GW over a global ocean area of 361 million km\(^2 \)).

Our values of \(|J|\) and \( \langle \nabla \cdot \mathbf{J} \rangle \) are 2 orders of magnitude below the values reported by SHA. However, given the DOVETAIL physical oceanographic evidence (Muench et al., 2002), we trust our lower values for \(|J|\) (of order 0.1–1 kW m\(^{-1} \)) rather than the SHA value of \( \sim 45 \) kW m\(^{-1} \). We conclude that, contrary to the suggestion by SHA, the SSR is not a significant “hot spot” in the world map of baroclinic tides. The SHA study and a companion paper by Arbic et al. (2004) demonstrated, however, that global baroclinic tide models can now be run, and such models will soon contribute significantly to mixing and sea-ice studies. It is, therefore, essential that we identify the source of discrepancies between our regional models and the global solutions. While we can easily explain the differences between our own model runs (Table 1) as responses to changing model parameters, we do not understand the factor of 50 difference in \(|J|\) between our own “SHA” regional runs (Run#1–Run#3) and the SHA solution. One component of this discrepancy is the difference between the strength of the barotropic tides predicted by the SHA forward model and our own barotropic models (CATS02.01 and AP) used to provide open boundary conditions. The SHA model is more energetic than observed, whereas CATS02.01 is known to be reasonably accurate around Antarctica (Padman et al., 2002; King and Padman, 2005; King et al., 2005). SHA note that increasing the benthic quadratic drag coefficient \( C_D \) from its typical value of \( \sim 0.003 \) to 0.3 leads to significant improvement in SHA model accuracy for sea-surface elevation (primarily a barotropic response), while the barotropic-to-baroclinic conversion rate is halved. For one regional comparison, the SHA model predicts an area-integrated conversion rate for the Hawaiian Ridge which is a factor of \( \sim 2 \) higher than values from Egbert and Ray (2001) and Merrifield et al. (2001); see Table 3 in SHA. Although \( C_D = 0.3 \) is physically unreasonable, this result suggests that errors in the barotropic currents may account for some of the strong baroclinic generation in the SHA model. We tested the effect of overly energetic barotropic tides by multiplying the normal-flow currents along the open boundaries in our own model by a factor of two. The conversion rate doubled, still leaving a factor of 25 discrepancy between SHA and our own regional runs based on the global-averaged stratification used by SHA.

It is possible that the two types of models, our sigma-coordinate version and the SHA isopycnal model, have very different capabilities for the generation of baroclinic tides over ridges. Our regional solutions also may be excluding significant baroclinic energy that has radiated into our domain from outside. We note also that SHA’s WKB-like normalizations (their Eqs. (14) and (15)) do not adequately suppress internal tide generation for the weak Southern Ocean stratification. Their normalization for the SSR reduces \( \langle C_S \rangle \) by a factor of \( \sim 2 \); our own test of changing stratification from the SHA global to a Powell Basin regional profile reduces \( \langle \nabla \cdot \mathbf{J} \rangle \) by a factor of order 100. We plan further studies to develop
our understanding of why the SHA global model generates much more energetic baroclinic tides in this region than our own models and oceanographic data suggest.

While our own models indicate that production of baroclinic tidal energy along the SSR is weak relative to major world hot spots, the internal tide in this region can elicit a significant sea-ice response to the ocean surface velocity field. The rms lateral divergence $\sigma(\nabla_H \cdot \mathbf{u})$ of the pack ice is an order of magnitude higher in the baroclinic tidal model than in our barotropic (depth-integrated) model. This increase arises primarily through a reduction in the spatial scales of current variability rather than through significant changes in the magnitude of the surface currents when the baroclinic component is added to the barotropic tide. The effect of periodic tidal divergence is to increase the mean open-water fraction in the winter ice pack, consequently increasing the rate of ice formation in fall and winter and accelerating the rate at which solar radiation melts the sea ice in spring and summer. It has already been shown (Koentopp et al., 2005) that the addition of barotropic tides to a general circulation, coupled ocean/ice model has a significant effect on ice concentration distribution and temporal variability. According to Koentopp et al. (2005), the role of tides is particularly marked in the marginal ice zones such as the SSR region: the new analyses indicate that the impact of tides on ice concentration can be much greater when baroclinicity is included in the ocean tide model (see Fig. 7).

Our study points to the need, as SHA have already noted, to use realistic stratification and adequate resolution in global baroclinic tide models, especially for the purpose of predicting the tide’s contribution to ocean mixing and sea-ice response at high latitudes. This goal is now within reach of supercomputer resources. For high-latitude researchers, however, the primary needs now are to improve the resolution and accuracy of bathymetric grids, and to develop the coupling between ocean baroclinic models and the overlying sea ice. Refining and testing recently developed sea-ice models is needed, with an emphasis on realistically representing the dynamic response of ice to stress, strain and divergence that is applied at high frequencies (> 1 cycle day$^{-1}$) and short spatial scales (1–10 km).

Acknowledgments

This study was funded by Grants OPP-0003423 and OPP-0338101 from the National Science Foundation, Office of Polar Programs and Grant N00014-03-1-0067 from the Office of Naval Research. We thank H. Simmons for providing the values of $C$ and $J$ from his global modeling study and for energetic discussions on possible sources of model discrepancies, and G. Egbert and S. Erofeeva for their important contributions to our Antarctic barotropic tide modeling program. Two reviewers provided valuable advice on the first version of this manuscript. This is ESR contribution number 90.

References


and observations in the western Weddell Sea during 1992.

SAR-derived surface fluxes in the western Weddell Sea
during Ice Station Weddell 1992. Journal of Geophysical
Research 110, C04002.

Holloway, P.E., Merrifield, M.A., 1999. Internal tide genera-
tion by seamounts, ridges, and islands. Journal of Geophy-
sical Research 104 (C11), 25,937–25,951.

variability and age of the internal tide at the Hawaiian

tide models around Antarctica. Geophysical Research

King, M.A., Penna, N.T., Clarke, P.J., King, E.C., 2005.
Validation of ocean tide models around Antarctica using
onshore GPS and gravity data. Journal of Geophysical
Research 110, B08401.

Koentopp, M., Eisen, O., Kottmeier, C., Padman, L., Lemke,
P., 2005. Influence of tides on sea ice in the Weddell Sea:
investigations with a high-resolution dynamic-thermo-
dynamic sea ice model. Journal of Geophysical Research
110, C02014.

In: Johannessen, O.M., Muench, R.D., Overland, J.E.
(Eds.), The Polar Oceans and Their Role in Shaping the
Global Environment, Geophysical Monograph 85. AGU,
Washington, DC, pp. 137–158.

Sub-daily sea ice motion and deformation from RADAR-
SAT observations. Geophysical Research Letters 30 (23),
2218.

testing of a lateral boundary scheme as an open boundary
condition in a barotropic ocean model. Coastal Engineering
11, 603–627.

closure models for planetary boundary layers. Journal of
Atmospheric Sciences 13, 1791–1806.

closure model for geophysical fluid problems. Reviews of

Merrifield, M.A., Holloway, P.E., Johnson, T.M.S., 2001. The
generation of internal tides at the Hawaiian Ridge.


Upper ocean diapycnal mixing in the northwestern Weddell

Munk, W., Wunsch, C., 1998. Abyssal recipes II: Energetics of

Padman, L., Kottmeier, C., 2000. High-frequency ice motion
and divergence in the Weddell Sea. Journal of Geophysical
Research 105, 3379–3400.

Padman, L., Fricker, H.A., Coleman, R., Howard, S.L.,
Erofeeva, S., 2002. A new tidal model for the Antarctic

Ray, R.D., Mitchum, G.T., 1996. Surface manifestation of
internal tides generated near Hawaii. Geophysical Research
Letters 16, 2101–2104.

Simmons, H.L., Hallberg, R.W., Arbic, B.K., 2004. Internal
wave generation in a global baroclinic tide model. Deep-Sea
Research II 51, 3043–3068.

Stashchuk, N., Vlasenko, V., 2005. Topographic generation of
internal waves by nonlinear superposition of tidal harmo-

Wadhams, P., 2000. Ice in the Ocean. Gordon and Breach,
London (364pp.).

Wunsch, C., 1975. Internal tides in the ocean. Reviews of
Geophysics and Space Physics 13, 167–182.

743–744.