Baroclinic and Barotropic Tides in the Ross Sea

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The barotropic and baroclinic tides in the Ross Sea were simulated using primitive equation, sigma-coordinate model, the Regional Ocean Model System (ROMS), for four tidal constituents, $M_2$, $S_2$, $K_1$, and $O_1$. The model results predicted the elevation amplitudes to be small, with a combined standard deviation < 40 cm over most of the basin. Larger amplitudes, with standard deviations ranging from 40-60 cm, occurred in the southeast region of the ice shelf cavity and over the continental slope and Iselin Bank. Most of the elevation variation was associated with the diurnal constituents ($K_1$ and $O_1$), with elevation amplitudes reaching 50 cm under the ice shelf and at points along the continental slope where continental shelf waves were generated by the diurnal tides. For the semidiurnal constituents ($M_2$ and $S_2$), the elevation amplitudes were generally small, less than 10 cm, except under the Ross Ice Shelf, where they reached 30 cm. The simulated tides were barotropic for the diurnal constituents, but baroclinic tides were generated at locations of steep topography for the semidiurnal constituents. Diurnal continental shelf waves were found to amplify the semidiurnal elevations and baroclinic tidal velocities over the continental slope. Comparisons to observations in both elevation and velocities, using the elevation amplitude and phase and the major axis of the tidal ellipses respectively, showed very good agreement for the semidiurnal constituents, good agreement for the $K_1$ constituent, but poor agreement for $O_1$. The $O_1$ constituent overestimates the generation of continental shelf waves along the shelf break resulting in overestimations along the continental shelf break and underestimates in the ice shelf cavity. The baroclinic tides induce both small scale spatial shear and vertical shear in the velocity fields in the Ross Sea.

**Key Words:** Tides, Ross Sea, Internal tides
1. Introduction

Tides have been recognized to play a significant role in ocean mixing (Munk & Wunsch 1998; Muench et al. 2002; Garrett 2003). Tidal mixing over rough topography in the deep ocean is estimated to account for \( \sim 1 \) TW (terawatt; \( 1 \text{TW}=10^{12} \text{W} \)) of energy due to the \( M_2 \) constituent (Egbert & Ray 2001) and another 0.5 TW due to the solar constituents (Garrett 2003). Tidal energy maps indicate more tidal energy is lost to mixing in the shallow seas than in the deep ocean.

How do the tides influence mixing? In the polar regions, tides have the capability to influence mixing and heat transfer through several mechanisms. Barotropic and baroclinic tides affect the ocean-atmosphere heat transfer (Padman & Kottmeier 2000; Robertson 2001b). Increased benthic mixing is induced by barotropic tides through higher benthic shear. In addition, mixing has the potential to affect the heat transport from the Warm Deep Water to the surface ice and influence the ice cover (McPhee et al., 1996; Muench et al., 2002; Beckmann et al., 2001). Tides induce lead formation in the ice due to convergence and divergence of the velocities (Nansen 1898). The lead formation then affects the ocean-atmosphere heat transfer.

In the Antarctic seas, mixing along the continental slope can affect the formation of deep and bottom waters and their ventilation. Internal tidal mixing at the shelf fronts around Antarctica contributes to deep water production. The cross-slope exchange in the Ross Sea both from tidal mixing and other processes is being investigated by the Antarctic Slope Initiative (AnSlope), an observational program, in order to determine the contributions of the different mechanisms to the cross-slope exchange.

In order to understand and quantify the tidal contributions to mixing, lead formation, and other processes in the Weddell Sea, the structure of the tides in the Ross Sea must be known. Since only sparse tidal observations are available in the Ross Sea (dots and triangles in Figure 1), models are being used to fill in data gaps and provide estimates of the tidal structure. Barotropic tides in the Ross Sea have been modeled for multiple tidal constituents by MacAyeal [1984] and Padman and associates (Padman & Kottmeier 2000; Padman et al. 2003). The vertical structure of baroclinic tides, however, has only been simulated for one semidiurnal constituent (\( M_2 \)) (Robertson et al. 2003). The goal of this project was to extend this modeling effort to four major constituents (semidiurnals: \( M_2 \) and \( S_2 \) and diurnals \( K_1 \) and \( O_1 \)) and develop a description of the three-dimensional structure of the tides in the Ross Sea. For this study, the Regional Ocean Modeling System (ROMS), described in Section 2, was used. In Section 3, the model results are discussed and compared to existing observations. A summary is given in section 4.

2. Model Description
The ROMS model in this application is fully described elsewhere (Robertson, 2001a; Robertson 2003). Consequently, only the pertinent details and recent modifications will be discussed here. Recently, improvements have been made in the vertical mixing parameterization for primitive equation models. Different vertical mixing parameterizations available in ROMS were evaluated using a sensitivity study at a site with extensive hydrographic, velocity, and turbulence observations (Robertson 2003b). For vertical mixing, the Generic Length Scale (GLS) parameterization of Umlauf & Burchard (2003) was found to replicate the observed velocities and vertical diffusivities the best and was used for these simulations. Tidal forcing was implemented by setting elevations along all the open boundaries, with the coefficients taken from a two-dimensional model of the region ((CATS-99) Padman & Kottmeier 2000). Four major tidal constituents were used, two semi-diurnals, $M_2$ and $S_2$, and two diurnals, $K_1$ and $O_1$.

ROMS had been modified previously to include the presence of a floating ice shelf (Robertson et al. 2003), including both mechanical and thermodynamic effects following Robertson (1999; 2001a) and Hellmer & Olbers (1989), respectively.

The model domain (Figure 1) covered the Ross Sea with a spacing of 0.045° in latitude, 0.186° in longitude, and with 24 vertical levels. The ice thickness and water column thickness, defined as the distance between the bottom and the ice shelf base (or ocean surface), were taken from BEDMAP (Lythe et al. 2000). In the northern region beyond the extent of BEDMAP, the water column thickness was taken from Smith & Sandwell (1997). No smoothing was performed on either the water column or the ice shelf thicknesses. The minimum water column thickness was set to 75 m, which entailed artificially deepening a small portion of the region under the ice shelf, particularly near the eastern grounding line. The barotropic and baroclinic mode time steps were 6 s and 180 s, respectively, and the simulations were run for 30 days, with hourly data from the last 15 days saved for analysis. The kinetic and potential energies stabilized around 15 days, thus the first 15 days of simulated data was discarded.

Initial potential temperature, $\theta$, and salinity, $S$, fields were assembled from the World Ocean Circulation Experiment (WOCE) Hydrographic Programme Special Analysis Center (Gouretski & Janke 1999), except under the ice shelf. Here, the hydrography was essentially a four layer system based on a $\theta$ and $S$ profile through the ice shelf at a single location (J9 (Jacobs 1989)). The uppermost layer was 40 m thick, where possible, with $\theta =$ the freezing point of water at the depth of the ice shelf base and $S=34.39$ psu. The second layer was 100 m thick with $\theta = 0.06^\circ$C warmer than the freezing point of water for the depth. $\theta$ increased linearly between this value and that of the lowermost layer within the 50 m thick third layer. The lowermost layer encompassed the remaining water column (deeper than 190 m below the ice shelf base) with $\theta =$
1.87°C and \( S =34.83 \) psu. \( S \) increased linearly between values in the uppermost and the lowermost layers over the 150 m of the middle two layers. The model elevation and velocities were initialized with the geostrophic velocities associated with the initial hydrography as determined from a simulation without tidal forcing.

Elevation, \( \zeta \), and depth-independent velocities, \( U_2 \) and \( V_2 \), are produced by the two-dimensional (barotropic) mode of the model and \( \zeta, S \), and depth-dependent velocities, \( U_3, V_3 \), and \( W_3 \), by the three-dimensional (baroclinic) mode. Foreman’s tidal analysis routines were used to analyze these fields (Foreman 1977; 1978).

3. Results
3.1 Elevations and Depth-Independent Velocities
3.1.1 Semidiurnal constituents (M\(_2\) and S\(_2\))

The elevation amplitude response for both semidiurnal constituents, M\(_2\) and S\(_2\), were similar. The prominent feature in both was an amphidromic point located in the western Ross Sea under the ice shelf at \( \sim 180^\circ W, 81^n S \) (Figures 2a and 2b). These amphidromic points result from destructive interference of the tidal Kelvin wave, which is \( \sim 180 \) out of phase on opposite sides of the basin under the ice shelf. The locations are dependent on the relative size of the basin, its water column thickness, and the tidal wavelength (MacAyeal 1984). The semidiurnal tidal elevation amplitudes in the Ross Sea were generally small, \( \sim 10 \) cm, although values > 30 cm were reached for both semidiurnal constituents in the channel under the southeastern portion of the ice shelf (Figure 2a).

To evaluate the model performance, the modeled elevation amplitudes and phases were compared to those of the existing observations, which are concentrated under the ice shelf (locations are shown as dots in Figure 1). With one exception for M\(_2\) and five exceptions for S\(_2\), the model elevation amplitudes matched the observations within the observational uncertainty, 2 cm (Williams & Robinson 1980) (Table 1). The phases agreed within 45° for ten and seven of the twelve locations for M\(_2\) and S\(_2\), respectively (Table 1). The phase changes rapidly in the vicinity of the amphidromic point and small errors in the location of this point can result in significant errors for the phases.

It is believed that most of the differences between the observations and the model estimates are a result of topographic errors. The bathymetry and ice shelf thickness under the Ross Ice Shelf are not well known and the model is quite sensitive to the topography. For an example of the accuracy of the topography, it was necessary to shift the position of a couple of the elevation points to the nearest submerged grid cell, since the observation location was on land in the model topography. The observation location exceeding the observational uncertainty for M\(_2\) was one of
these points. This effect has been previously seen in model results (MacAyeal 1984; Robertson et al. 1998; Padman et al. 1999). A small shift in the position of the amphidromic point location will result in significant amplitude and phases changes.

The model estimates were compared to the results of two different two-dimensional tidal models (MacAyeal 1984; Padman & Kottmeier 2000) (Table 2). ROMS had better agreement with observations for both semidiurnal constituents than either Padman’s model (CATS-99 grid spacing: ~6 km in longitude and ~9 km in latitude) or MacAyeal’s model (grid spacing: ~10 km in longitude and ~10 km in latitude) with lower rms differences for ROMS (Table 2). ROMS also had better phase agreement than the other two models (Table 2). The differences between the model results are attributed to two factors 1) ROMS is a three-dimensional model and the other two models are two-dimensional, and 2) the models use different topographies. Although there is good agreement, particularly for M\textsubscript{2}, and ROMS outperformed the other models, the observation locations are too sparse and are not well enough distributed over the domain to provide a definitive comparison and verify the model results.

The response of the semidiurnal constituents is affected by the presence of the diurnal constituents in the shelf break region. As will be discussed more fully later, the diurnal constituents excite continental shelf waves (CSWs) along the continental slope. Since the semidiurnal constituent frequencies are near the first harmonic of the diurnal constituents, interactions between the CSW and variations in topography transfer energy from the diurnal tides to their first harmonic, which is incorporated into the semidiurnal tides in this region. This can be seen by the local regions of higher amplitude over Iselin Bank and along the continental slope in Figures 2a and 2b, which are present when four tidal constituents were used for forcing. These regions are not present when forced by only the semidiurnal constituent M\textsubscript{2} (Figure 2c).

The major axes of the tidal ellipses of the semidiurnal constituents for the depth-independent velocities were quite small, generally < 5 cm s\textsuperscript{-1}. Major axes > 5 cm s\textsuperscript{-1} occurred over Iselin Bank, along the continental shelf break, and at the entrance to the channel in the southeast corner under the ice shelf (Figures 3a and 3b). The large major axes over Iselin Bank and along the shelf break correspond to the harmonics of the CSW and those under the ice shelf to the restricted topography in that corner of the domain. Again the topography was the primary controlling factor for the depth-independent velocities, as has been previously noted for a two-dimensional tidal model (Robertson et al. 1998).

3.1.2 Diurnal Constituents (K\textsubscript{1} and O\textsubscript{1})

The tidal elevation response of the diurnal constituents, K\textsubscript{1} and O\textsubscript{1}, was quite different than the semidiurnal response. The diurnal constituents progressed from east to west as a Kelvin wave.
without the generation of an amphidromic point (Figures 2d and 2e, respectively). Large elevation amplitudes occurred in the ice shelf cavity, particularly in the narrow channel in the southeastern corner of the ice shelf cavity and over Iselin Bank and along the continental slope, with the latter a result of the generation of continental shelf waves.

Continental shelf waves (CSW) are excited by interactions of velocities with topographic irregularities (Huthnance 1978; Brink 1991). Excellent summaries of their properties are given by Huthnance (1978) and Brink (1991). They are free coastally trapped waves with sub-inertial frequencies (Huthnance 1978; Brink 1991). For the Ross Sea, they are excited by the diurnal tides, but not the semi-diurnal tides. For long waves, the group velocity propagates in the same direction as the phase velocity. In the southern hemisphere, CSW propagate from west to east (Huthnance 1978; Brink 1991) and in weak stratification, they are essentially barotropic (Huthnance 1978). Increasing stratification will increase the wave speed (Huthnance 1989) and induce the frequency to increase. The CSW decay primarily through bottom friction (Brink 1991). Realistic topography was found to concentrate the motion over the upper slope and shelf (Huthnance 1978). In the presence of mean alongshore flow, the properties of CSW change due to a combination of Doppler shift, changes in the background vorticity, and the growth of instabilities (Brink 1991). The behavior of the mode one CSW present in the model results is consistent with this behavior, propagating westward, and a concentration of the motion over the upper slope and shelf.

Comparisons of the model estimates of the elevation amplitudes to the observations for the diurnal constituents, $K_1$ and $O_1$, showed good agreement for $K_1$, but poor agreement for $O_1$ with rms values of 2.5 and 6.8 cm for the $K_1$ and $O_1$ constituents, respectively. Most of the $K_1$ mismatches occurred in shallow waters under the eastern portion of the ice shelf. They are attributed to topographic errors. The $O_1$ elevations are significantly underestimated by about 10 cm deep under the ice shelf and by about 5 cm at other locations nearer the ice shelf edge. The $O_1$ discrepancy will be discussed more in Section 3.2.2.

ROMS outperformed the MacAyeal (1984) and Padman & Kottmeier (2000) models in replicating tidal elevation amplitudes for $K_1$, but not for $O_1$, where ROMS performs more poorly (Table 2). The phase rms for the three models are roughly equivalent.

With the exception of over the continental slope, Iselin Bank and two regions under the ice shelf, the diurnal major axes are small < 10 cm s$^{-1}$ (Figures 3c and 3d). However, over the continental slope and Iselin Bank, CSW induced high velocities in the diurnal constituents (Figures 3c and 3d). The $O_1$ frequency CSW exceeded those at the $K_1$ frequency even though
observations show less of an O₁ CSW response. In a prior two-dimensional model of the Weddell sea, O₁ CSW were also overestimated (Robertson et al., 1998; Robertson, 1999).

3.2 Depth-Dependent Response

3.2.1 Semidiurnal Constituents (M₂ and S₂)

To further evaluate the model performance, the major axes of the tidal ellipses for depth-dependent velocities were compared to those of fifty-four current meters from twenty separate moorings, whose locations are shown as triangles in Figure 1. (Although there are fifty-four current meters, there are only fifty-one separate sites, since three sites had repeat observations.) The model estimates for the M₂ major axes at forty-eight of the fifty-one sites agreed with the observed values (Figure 4 and Table 3) within the observational uncertainty of 1.7 cm s⁻¹ (Pillsbury & Jacobs 1984) and the estimates for the S₂ major axes agreed for forty-nine of the fifty-one sites (Table 3). The rms differences for the major axes were 5.3 and 3.5 cm s⁻¹ for M₂ and S₂, respectively (Table 4). For one of the locations of disagreement, a location along the continental slope (76° 30.093' S in Figure 8), the model overestimates the velocity for the upper current meter. This overestimation does not occur in the simulation with only M₂ tidal forcing, thus it represents a harmonic response semidiurnal constituent to the diurnal CSW. Two other disagreement locations are over Iselin Bank and the observations reported extremely high tidal benthic velocities (bottom values at 72° 28.813' S, 76° 29.808' S in Figure 4). These high benthic values were not seen in the model results. Most of the rms differences were associated with the site having the larger of these benthic values, since the rms differences were reduced to 1.0 and 0.9 cm s⁻¹ for M₂ and S₂, respectively, when this site was excluded (Table 4). A possible explanation for these high observational values is that they are a first harmonic response from the CSW induced by the diurnal constituents in response to Iselin Bank. The observations may also be locally high values, which result from small-scale topographic features irresolvable with the model topography. There is also the possibility that the high observational values are an inertial response, however, their benthic location and strong depth dependence reduce this possibility.

The rms difference reflects only the errors in the major axes. A more comprehensive measure is the absolute error, which combines errors in the amplitude with those in phase. The absolute errors (Eₐ) for each constituent were calculated for both positive (anticlockwise) and negative (clockwise) components according to

\[ E_\text{A} = \sqrt{ \frac{1}{L} \sum_{i=1}^{L} \left( A_0 - A_m \right)^2 + \sum_{i=1}^{L} \left( \phi_0 - \phi_m \right)^2 } \]

where L is the number of observations and A₀ and Aₘ and \( \phi_0 \) and \( \phi_m \) are the amplitudes and phases, respectively, with the subscripts of \( o \) and \( m \) representing the observations and model results, respectively (Cummins and Oey 1997). The
absolute error is determined for each component for each constituent. The absolute errors in the positive components for $M_2$ and $S_2$ were 0.73 and 0.76 cm s\(^{-1}\) respectively (Table 4). Again, a significant portion of these errors was associated with the one location with high benthic values and the errors were reduced to ~0.4 cm s\(^{-1}\) for both $M_2$ and $S_2$, when this location was excluded (Table 4). In the Southern Hemisphere, most of the baroclinic response for the boundary layers should be associated with the positive component. The negative component is barotropic in the Southern Hemisphere. The absolute errors in the negative components for $M_2$ and $S_2$ were smaller, 0.58 and 0.59 cm s\(^{-1}\), respectively (Table 4). Again, a significant portion of these errors was associated with the one location with high benthic values and the errors were reduced to ~0.3 cm s\(^{-1}\) for both $M_2$ and $S_2$, when this location was excluded (Table 4). The absolute errors indicate that more error is associated with the baroclinic response than the barotropic response.

For the $M_2$ semidiurnal constituent, the critical latitude, $74^\circ 28.8'$ S, crosses through the domain (Figure 1 and dashed line in Figure 5). At the critical latitude the inertial frequency equals the tidal frequency, which affects both the generation and propagation of internal tides. Linear internal wave theory predicts that internal waves will not be generated nor propagate poleward of the critical latitude. Robertson (2001a) gives a more extensive explanation of the relevant linear internal wave theory.

Generation and propagation of internal tides occurred for the semidiurnal constituents over the continental slope and steep topography (for example, with $M_2$ areas indicated with IT in Figures 5a-5e). The semidiurnal internal tides had the expected theoretical wavelengths for mode 1 waves and propagated along the expected wave characteristics according to linear internal wave theory. A previous study showed that critical latitude effects increased generation, especially in the western Ross Sea (Figures 5a-5c), where the continental slope is equatorward of critical latitude (Robertson et al, 2001). However, with multiple constituents, the critical latitude effects were overwhelmed by amplification by CSWs. The locations of intense internal tide generation (velocities > 6 cm s\(^{-1}\) over the continental slope and steep topographic features in Figure 5 have significantly smaller velocities, < 4 cm s\(^{-1}\), in the simulation with only the $M_2$ constituent. Furthermore, these areas coincide with the locations of high diurnal velocities (Figures 3c-d and 6). Thus internal tide generation for the semidiurnal constituents is amplified by diurnal CSW.

Internal tides were also generated under and at the front of the ice shelf (areas marked 2L in Figures 5a-5f). Although the existence of internal tides is not expected poleward of the critical latitude in a continuously stratified fluid according to linear internal wave theory, it is quite possible for a baroclinic mode tide to exist in the multiple layer system (Kundu 1990, p. 233), such as that of the initial hydrographic conditions in the ice cavity. A two-layer response is
Robertson: Baroclinic and Barotropic Tides in the Ross Sea

apparent under the ice shelf with the upper layers (~140 m) essentially responding together as the
top layer and the lowermost layer as the bottom layer (for example, with M\textsubscript{2} area marked 2L in
Figures 5b-5d).

Comparison of the M\textsubscript{2} major axes of the tidal ellipses for depth-independent (Figure 3a) and
the surface depth-dependent (Figure 7) velocities show the depth-dependent velocities to have
more spatially variability in the depth-dependent velocities with many small scale local features
and velocities amplified over the depth-independent values. This spatial variability is induced by
spatial differences in the baroclinic tides. The spatial variations, in turn, induce divergence and
convergence in the surface velocity fields, which affects lead formation and pack ice dynamics.

3.2.2 Diurnal Constituents (K\textsubscript{1} and O\textsubscript{1})

Agreement between the model estimates and the observations was not as good for the diurnal
constituents as the semidiurnals, particularly along the continental slope, with estimates at
seventeen and forty of the fifty-one sites falling within the observation uncertainty for K\textsubscript{1} and O\textsubscript{1},
respectively (Table 3). The rms differences for the major axes were 14.7 and 16.7 cm s\textsuperscript{-1} for K\textsubscript{1}
and O\textsubscript{1}, respectively (Table 4). Again a significant portion of the K\textsubscript{1} rms difference is due to the
one large observation value, without which the rms difference reduces to 9.7 cm s\textsuperscript{-1} (Table 4).
The O\textsubscript{1} rms difference is not affected by exclusion of this point. The absolute errors for the
positive and negative rotating components was 3.2 and 1.8 cm s\textsuperscript{-1} for K\textsubscript{1} and O\textsubscript{1}, respectively,
when the high observational value was excluded.

The model overestimated the CSW along the continental shelf break resulting in higher
values than those observed for those stations with the exception of the large benthic value at 76°
30.095' S (for K\textsubscript{1}, top row in Figure 8). Disagreement with the observations also occurred at the
two easternmost sites along the ice shelf (for K\textsubscript{1}, right two plots on second row in Figure 8) and at
other locations along the ice shelf edge (for K\textsubscript{1}, bottom row in Figure 8) where the observations
varied vertically, but the model estimates were barotropic. The model overestimated the CSW for
the O\textsubscript{1} constituent to even a greater degree than K\textsubscript{1} and the O\textsubscript{1} estimates disagreed at more sites
along the ice shelf edge, again with the observations indicating more baroclinicity than the model
estimates.

The overestimation of the O\textsubscript{1} CSW along the continental slope can account for much of the
disagreement for the O\textsubscript{1} elevations and velocities. Most of the velocity sites are located along the
continental slope or the ice shelf edge. In both of these regions, O\textsubscript{1} velocities are severely
overestimated due to the overexcitation of O\textsubscript{1} CSWs. The overestimations are larger near the
continental slope, a factor of ~5, whereas along the ice shelf edge, it is a factor of ~2. So the
overestimations are worse near the continental slope where the continental slope waves are
generated and are reduced further away from their source. Most of the elevation sites are poleward of the continental slope in the ice shelf cavity. The model underestimates the elevations with the underestimations increasing with increasing distance under the ice shelf cavity or away from the continental slope. The continental slope wave response redistributes the O$_1$ tidal energy, generating larger elevations and velocities along the continental shelf. Consequently, tidal energy is dissipated along the continental slope and upper continental shelf and less O$_1$ tidal energy is available to flow into the ice shelf cavity. This results in an underestimation of the elevations and velocities in the ice shelf cavity and the elevations discrepancies increase with increasing distance under the ice shelf.

Unfortunately, the cause for the overestimation of O$_1$ CSW by the model is unknown, although there are several possible sources. It is possibly linked to the lack of wind-driven mean flows in the model. Mean alongshore flows can modify the CSW properties. Also stratification can cause the frequency of the CSW to shift. It is possible that in the natural conditions, a shift is caused by the stratification and/or the presence of a mean alongshore flow. This shift transfers energy from the diurnal components into other frequencies where they are not picked up in the tidal analysis. The model conditions may not reproduce this shift due to the lack of a mean flow and/or different hydrographic conditions. Resonance effects associated with errors inherent in the model are another potential source for the discrepancy.

The diurnal response of the depth-dependent velocities was essentially barotropic. Large major axes occurred at sites along the continental slope, over Iselin Bank and areas of steep topography and are associated with CSWs (for example, with K$_1$ areas marked CSW in Figure 6). The baroclinic response was slight, essentially a surface and benthic boundary layer effects (for example, with K$_1$ areas marked VS in Figure 6). A slight baroclinic response also occurred near the ice shelf edge (Figures 6). The Ross Sea is 40° or more poleward of the diurnal critical latitudes, so the response is expected to be barotropic without baroclinic tidal generation. Despite the theory, the observations do show some variation in velocities in the vertical direction (Figure I and Table 3).

3.3 Combined Tidal Response

To this point the discussion has considered the tidal response separately for the different constituents. This is appropriate when evaluating the model performance; however, the dynamics respond to the combined elevation and velocity fluctuations. The combined response will be evaluated using the standard deviations of the elevations and velocities over the fifteen days of model results.
The largest elevation standard deviations occurred in the channel in the southeastern corner under the ice shelf and over Iselin Bank (Figure 9a). The high variance over Iselin Bank is attributable to the generation of CSW. The restrictive topography of the channel in the southeastern corner and the increase of Kelvin wave height in shallow water induces large standard deviations in that region. The depth-independent velocity standard deviations are also highest along the continental slope and over Iselin Bank (Figure 9b) again due to CSW. It should be noted that this combined estimate of the variability is slightly larger than it should be along the continental shelf break and slightly smaller under the ice shelf due to the overestimation and underestimation of the $O_1$ constituent, respectively.

Standard deviations of the depth-dependent velocities also are highest over regions of steep topography, such as the continental slope and Iselin bank, (Figure 10 for the east-west and Figure 11 for the north-south depth-dependent velocities, respectively). Differences in the magnitude of the standard deviations for the two directions occur due to the alignment of the continental slope with respect to the velocities (Figures 10 and 11), with the largest standard deviations in the direction perpendicular to the direction of continental slope. In the eastern region, where the continental slope runs nearly east-west, the north-south velocities show more variability (continental shelf break in Figures 11d and 11e compared to those in Figures 10d and 10e). Near 178° W, the continental slope turns northward around Iselin Bank and the east-west velocities show more variability (plateau features in Figures 11c and 11d compared to those in Figures 10c and 10d). West of Iselin Bank, the continental slope is again east-west and the larger standard deviations are associated with the north-south velocities (Figures 10a and 10b). Vertical differences are apparent in the standard deviations for both directions indicating a baroclinic response (Figures 11 and 10) and vertical shear in the horizontal velocities. Since the diurnal response was nearly barotropic, this baroclinic response is induced by the semidiurnal constituents.

Vertical shear in the horizontal velocities can induce mixing in the water column, through increased bottom (and surface under the ice shelf) friction and shear instabilities. Inspection of the total vertical shear in the horizontal velocities show that the tides induce shear in the water column (Figure 12). Tides can induce vertical shear in two ways: 1) increasing the velocities increases the benthic (and surface under the ice shelf) boundary layer(s) quadratically as the velocity increases and 2) through baroclinic tides. Both effects are present. The vertical shear in the benthic boundary layer and the surface boundary layer are apparent in Figure 12, particularly Ha with the enlarged vertical scale. Theoretically, the benthic boundary layer should increases near the critical latitude (Furevik & Foldvik, 1996). The model replicates this effect, which is
most apparent in Figure 12e, where the benthic boundary layer is thicker equatorward of the $M_2$ critical latitude and then thins poleward. Vertical shear attributable to baroclinic tides is also apparent, particularly north of the $M_2$ critical latitude over steep topographic features (Figure 12).

4. Summary

Simulations of the barotropic and baroclinic tides in the Ross Sea basin and the cavity under the Ross Ice Shelf were performed both with the major semidiurnal constituent, $M_2$, and for four tidal constituents combined, $M_2$, $S_2$, $K_1$, and $O_1$, using ROMS. Comparison of the model results to tidal estimates from elevation and velocity observations showed excellent agreement for both semidiurnal constituents. The agreement for the elevations for the diurnal $K_1$ constituent was also good, however, the agreement with the major axes from the observations was not as good, particularly along the continental slope, where the model overestimated the tidal response in the form of continental shelf waves. Likewise, the model generated more continental shelf waves than were observed for the diurnal $O_1$ constituent, which resulted in poor agreement with the observations and less energy flowing into the ice shelf cavity at that frequency.

The diurnal frequency continental shelf waves were found to transfer energy to the semidiurnal constituents through their first harmonic. Thus, simulations with all the constituents combined showed more energy along the continental slope and less energy in the ice shelf cavity at the $M_2$ frequency than the single constituent simulation.

The diurnal tides were essentially barotropic with their baroclinic response associated with the benthic and surface boundary layers. Most of the baroclinic response occurred at the semidiurnal frequencies. Semidiurnal baroclinic tides were generated along the continental shelf break and over regions of steep topography, particularly over steep topography north of the critical latitude. The semidiurnal baroclinic tides were significantly amplified by the diurnal continental shelf waves in these regions. The baroclinic tidal behavior was consistent with that for internal waves in a continuously stratified fluid and propagated along wave rays with wavelengths corresponding to the theoretical estimates for mode one internal waves.

Under the ice shelf, the hydrography was specified by a series of four layers. Model velocities show that the top three layers responded together to effectively give a two-layer baroclinic response. In the ice shelf cavity, this baroclinic response was often present inducing shear into the water column in this region.

There is room for improvement in the model predictions. The model estimates will improve as input topography and bathymetry improves. Presently, a topographic survey of the southeastern portion of the Ross Ice Shelf cavity is being planned by investigators at Lamont-Doherty Earth Observatory. The Anslope initiative is obtaining hydrographic and bathymetric
Robertson: Baroclinic and Barotropic Tides in the Ross Sea

data for the shelf slope area of the Ross Sea. Data from both of these sources will be incorporated into the initial conditions for the model in future efforts. Also, the problem of the overexcitation of $O_1$ continental shelf waves needs to be addressed. The $O_1$ overexcitation may be linked to the lack of alongshore mean currents in the model. One of two improvements to the model that are presently in-progress is addition of the wind-induced mean circulation. The other improvement is coupling to dynamic sea ice model of Tremblay and Mysak [1997]. There improvements along with improved hydrography may alleviate the excessive $O_1$ continental shelf wave excitation. Inclusion of more of the forcing factors and dynamics will improve the model estimations. Additionally, through the Anslope program more tidal data is becoming available, which will be incorporated into future comparisons. Furthermore, it is planned to delve into what the effects of tides are on the surface pack ice, heat transfer to the atmosphere, and mixing in future efforts.

Acknowledgements:

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Robertson: Baroclinic and Barotropic Tides in the Ross Sea


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Robertson: Baroclinic and Barotropic Tides in the Ross Sea

List of Figures

**Figure 1.** The model domain with the water column thickness contoured at 100, 500, 1000, 2000, and 3000 m. The Ross Ice Shelf edge is indicated by a dashed line. The locations of tidal elevation observations are indicated by dots and those of velocity observations by triangles.

**Figure 2.** The amplitude of the tidal elevation for the a) $M_2$, b) $S_2$, c) $M_2$ only, d) $K_1$, and e) $O_1$ constituents contoured at 1, 2, 5, 10, 20, 30, 40, and 50 cm. The overlaid heavy lines indicate the phase for the elevation, with the zero phase line indicated.

**Figure 3.** The major axes of the tidal ellipses for the depth-independent velocities for the $M_2$, b) $S_2$, c) $K_1$, and d) $O_1$ constituents, contoured at 1, 5, 10, 20, 30, 40, 50, 60, 80, and 100 cm s$^{-1}$.

**Figure 4.** Major axes of the tidal ellipses from ROMS compared against observational data for the $M_2$ constituent. The error bars indicate the observational uncertainty (1.7 cm s$^{-1}$).

**Figure 5.** The major axes of the $M_2$ tidal ellipses from the depth-dependent velocities along N-S transects at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The pairs of letters indicate points discussed in the text (IT—internal tides; 2L—two-layer response). The location of the critical latitude is indicated by a dashed line.

**Figure 6.** The major axes of the $K_1$ tidal ellipses from the depth-dependent velocities along N-S transects at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The pairs of letters indicate points discussed in the text (CSW—continental shelf waves; VS—vertical shear).

**Figure 7.** The major axes of the $M_2$ tidal ellipses for the depth-dependent velocities at the surface with only $M_2$ tidal forcing, contoured at 1, 5, 10, 20, 30, 40, 50, 60, 80, and 100 cm s$^{-1}$.

**Figure 8.** Major axes of the tidal ellipses from ROMS compared against observational data for the $K_1$, constituent. The error bars indicate the observational uncertainty (1.7 cm s$^{-1}$).
Figure 9. The standard deviation of the a) east-west and b) north-south depth-dependent velocities at the surface.

Figure 10. The standard deviation in the east-west depth-dependent velocities over 15 days at a) 169°30' E, b) 176° E, c) 179°30' W, d) 175° W, and e) 160° W. The M2 critical latitude is indicated by a dashed white line.

Figure 11. The standard deviation in the north-south depth-dependent velocities over 15 days at a) 169°30' E, b) 176° E, c) 179°30' W, d) 175° W, and e) 160° W. The M2 critical latitude is indicated by a dashed white line.

Figure 12. Vertical shear in the depth-dependent velocities along N-S transects at a) 169°30' E, b) 176° E, c) 179°30' W, d) 175° W, and e) 160° W. The M2 critical latitude is indicated by a dashed white line.
List of Tables

Table 1. Elevation amplitudes and phases from ROMS compared against observational data and other two-dimensional models. Underlined values for the amplitude exceed the observational uncertainty for the amplitude and underlined values for the phase exceed a 45° difference. * indicates sites where multiple current observations were made.

Table 2. Rms differences for the elevation amplitude and phases at the twelve observation locations for each constituent for the ROMS simulations and for two-dimensional simulations of MacAyeal (1984) and Padman & Kottmeier (2000).

Table 3. Major axes of the $M_2$ tidal ellipses from the ROMS depth-dependent velocities from ROMS compared against observational data. Underlined values for the amplitude exceed the observational uncertainty (1.7 cm s$^{-1}$).

Table 4. Rms differences for the major axes of the tidal ellipses and absolute and relative errors for the positive and negative rotation components from the depth-dependent velocities at the observation locations for each constituent for ROMS.
Table 1. Elevation amplitudes and phases from ROMS for the M₂, S₂, K₁, and O₁ constituents compared against observational data and two two-dimensional models. Note: Although Padman has more recent model runs (Padman et al. 2003), the data is unavailable for comparison. Underlined values for the amplitude exceed the observational uncertainty.
<table>
<thead>
<tr>
<th>Constituent</th>
<th>M$_2$</th>
<th>S$_2$</th>
<th>K$_1$</th>
<th>O$_1$</th>
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<td></td>
<td>Amplitude (cm)</td>
<td>Phase (°)</td>
<td>Amplitude (cm)</td>
<td>Phase (°)</td>
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<tr>
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<td>4.4</td>
<td>65</td>
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<tr>
<td>Padman (2000)</td>
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<td>125</td>
<td>4.2</td>
<td>98</td>
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</table>

Table 2. Rms differences for the elevation amplitude and phases at the 12 observation locations for each constituent for the ROMS simulations and for two-dimensional simulations of MacAyeal (1984) and Padman & Kottmeier (2000).
<table>
<thead>
<tr>
<th>Latitude</th>
<th>Longitude</th>
<th>Water Depth (m)</th>
<th>Instr. Depth (m)</th>
<th>$M_2$</th>
<th>$S_2$</th>
<th>$K_1$</th>
<th>$O_1$</th>
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<td>1.0</td>
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<td>0.6</td>
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<td>174° 39.00' W</td>
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<td>237</td>
<td>0.5</td>
<td>0.7</td>
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<td>0.4</td>
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<td>170° 36.72' W</td>
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<td>228</td>
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<td>0.5</td>
<td>1.3</td>
<td>0.6</td>
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<tr>
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<td>174° 30.78' W</td>
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<td>210</td>
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<td>0.5</td>
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<td>0.6</td>
<td>0.3</td>
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<td>0.6</td>
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<td>1.0</td>
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<td>175</td>
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<td>0.8</td>
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<td>0.5</td>
</tr>
<tr>
<td>76° 30.093' S</td>
<td>167° 29.971' W</td>
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<td>241*</td>
<td>0.3</td>
<td>0.1</td>
<td>2.7</td>
<td>0.3</td>
</tr>
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<td>76° 29.808' S</td>
<td>174° 59.595' W</td>
<td>569</td>
<td>241*</td>
<td>0.8</td>
<td>0.1</td>
<td>1.0</td>
<td>0.9</td>
</tr>
</tbody>
</table>

* Measurements from ROMS simulation.
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Figure 2. The amplitude of the tidal elevation for the a) M$_2$, b) S$_2$, c) M$_2$ only, d) K$_1$, and e) O$_1$ constituents contoured at 1, 2, 5, 10, 20, 30, 40, and 50 cm. The overlaid heavy lines indicate the phase for the elevation, with the zero phase line indicated.
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Figure 4. Major axes of the tidal ellipses from ROMS compared against observational data for the $M_2$ constituent. The error bars indicate the observational uncertainty (1.7 cm s$^{-1}$). * indicates sites where multiple current observations were made.
Figure 5. The major axes of the $M_2$ tidal ellipses from the depth-dependent velocities along N-S transects at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The pairs of letters indicate points discussed in the text (IT-internal tides; 2L-two-layer response). The location of the critical latitude is indicated by a dashed line.
Figure 6. The major axes of the $K_1$ tidal ellipses from the depth-dependent velocities along N-S transects at a) 169° 30' E, b) 176° E, c) 179° 30' W, d) 175° W, and e) 160° W. The pairs of letters indicate points discussed in the text (CSW-continental shelf waves; VS-vertical shear).
Figure 7. The major axes of the M$_2$ tidal ellipses for the depth-dependent velocities at the surface with only M$_2$ tidal forcing, contoured at 1, 5, 10, 20, 30, 40, 50, 60, 80, and 100 cm s$^{-1}$. 
Figure 8. Major axes of the tidal ellipses from ROMS compared against observational data for the $K_1$ constituent. The error bars indicate the observational uncertainty (1.7 cm s$^{-1}$). * indicates sites where multiple current observations were made.
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Figure 10. The standard deviation in the east-west depth-dependent velocities over 15 days at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The M₂
critical latitude is indicated by a dashed white line.
Figure 11. The standard deviation in the north-south depth-dependent velocities over 15 days at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The M_2 critical latitude is indicated by a dashed white line.
Figure 12. Vertical shear in the depth-dependent velocities along N-S transects at a) 169° 30’ E, b) 176° E, c) 179° 30’ W, d) 175° W, and e) 160° W. The M$_2$ critical latitude is indicated by a dashed white line.