Observations of Tidally Induced Currents over the Continental Slope of the Laptev Sea, Arctic Ocean

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ABSTRACT

Two year-long (2004–05 and 2005–06) records of currents from two moorings deployed at the continental slope of the Laptev Sea (78°26′N, 125°40′E) are used in order to define the properties of tidal currents in the upper ~200-m ocean layer. Harmonic and spectral analyses of currents showed that the semidiurnal tidal constituent \( S_2 \) dominates over the semidiurnal \( M_2 \) and diurnal constituents. This dominance of the \( S_2 \) constituent in the tidal currents is due to resonant interaction of the superinertial wave with sloping bottom topography. In contrast to the tidal currents, sea level changes are dominated by the \( M_2 \) constituent, as seen from a tidal model by L. Padman and S. Erofeeva, using assimilation of observational data. Strong anticorrelation (\(-0.73 \pm 0.05\)) was found between the upper 50-m \( S_2 \) current amplitudes and local sea ice concentration, with fourfold (from ~2.0 to 8.5 cm s\(^{-1}\)) amplification of tidal currents under ice-free conditions. This is probably due to a change of local resonance conditions for the \( S_2 \) tidal current. These findings may be important for understanding the increasing role of tides in a seasonally ice-free Arctic Ocean.

1. Introduction

This paper discusses the properties of tidally induced currents derived from two year-long acoustic Doppler current profiler (ADCP) records from the mooring located at the continental slope of the Laptev Sea in the Arctic Ocean in 2004–05 and 2005–06. Recent Arctic Ocean Model Intercomparison Project (AOMIP) studies suggest that Arctic Ocean modeling may be improved if tides are explicitly included in coupled ice–ocean models (Holloway and Proshutinsky 2007). At the same time, most of the modern large-scale ocean models do not have explicit tidal forcing and thus do not incorporate tidal energy dissipation and the corresponding mixing. Consequently, simulated oceans (including the Arctic Ocean) may lack realistic heat and momentum fluxes. Observations of high-latitude tidal currents (especially for ice-covered regions) become critically important for improving model physics and validating numerical models because we are nearing a time when model experiments will routinely use explicit tidal forcing.

Dynamical processes over the continental slopes can be strongly influenced by tidal motions where the bottom topography is favorable for resonant amplification and enhanced mixing (Munk et al. 1970). Confirming the important role of slope topography for the Arctic Ocean tides, Hunke (1986) and later D’Asaro and Morison (1992), Padman et al. (1992), and Plueddemann (1992) reported significant amplification of the diurnal tides resulting in enhanced tidal mixing over the Yermak Plateau in the Eurasian Basin. An amplified barotropic tide interacts with stratification to generate baroclinic tidal currents, which absorb a significant part of tidal energy (Cummins et al. 2001). Dushaw and Worcester (1998) and Robertson (2001) reported that areas of steep shelf break located near the critical latitude (the latitude at which the frequency of a tidal constituent is equal to the inertial frequency) are favorable for generating internal tidal waves. Because the critical latitudes are 74.47°N for the \( M_2 \) constituent and 85.76°N for the \( S_2 \) constituent, the Arctic Ocean slopes, including the Laptev Sea slope, may be an area of strong internal wave production caused by the interaction of tidal waves with bottom topography and stratification. The internal tides affect the current structure and induce velocity shear in the water column, which can produce mixing through shear instability in low gradient Richardson number environment (Robertson 2001).

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Tidal dynamics in high-latitude regions is complicated by the presence of ice. Previous studies demonstrated the importance of sea ice conditions for tides in the Arctic seas (e.g., Sverdrup 1926; Zubov 1945; Murty 1985; Prinsenberg 1988; Bourke and Parsons 1993; Kowalik and Proshutinsky 1994; Pease et al. 1994, 1995). The impact of tides on ice is not well understood. Satellite imagery from Radarsat showed persistent oscillatory ice motion at mixed inertial–tidal frequencies (Kwok et al. 2003). The model experiments with a barotropic tidal model coupled with a simple sea ice dynamics model showed that the impact of ice on tidal properties is not strong and is generally local (Proshutinsky 1988). Still, there are numerous gaps in our knowledge of ice–tide interaction. Therefore, the role of ice in the formation of Arctic Ocean tides requires further clarification, including additional direct observations of tidal currents complemented by sea ice monitoring.

Observations of tidal currents over the Laptev Sea continental slope are rare and fragmentary. Analyzing an approximately 6-month-long ADCP record from 2004–05, Dmitrenko et al. (2008) found that the semidiurnal lunar tide ($M_2$; period $\approx 12.42$ h) was dominant with a maximal 4–5 cm s$^{-1}$ amplitude of tidal current. Recent studies by Lenn et al. (2011) confirmed that semidiurnal tides dominate tidal currents at the mooring located over the Laptev Sea slope (76°44′N, 125°55′E). In this study, we extend existing estimates of tidal properties for this region using an extended series of observations combined with the results of model simulations by Padman and Erofeeva (2004).

The article is organized as follows: Sections 2 and 3 provide a detailed description of the observational data used in this study and the method of analysis. In section 4, we analyze the general properties of tidal currents, their vertical structure within the upper ~200-m layer, and their temporal variability throughout the year. Sections 5 and 6 describe the impact of slope topography and sea ice conditions on tidal currents. Section 7 contains a discussion of the major results and presents general conclusions.

2. Data

In this study, we used two year-long records of ADCP observations at mooring M1 (78°26′N, 125°40′E; ~2690 m) located at the Laptev Sea continental slope (eastern Eurasian Basin; Fig. 1). The mooring position was in proximity to the Atlantic Water (AW) core as determined by the temperature maximum. The 300-kHz upward-looking ADCP deployed at a depth of ~56 m in September 2004 (mooring M1c) provided information about the vertical current structure in the upper ~50-m layer (Fig. 1, right). Horizontal current velocities were sampled over 12 equally spaced bins with a vertical resolution of 4 m; the first ADCP cell was located at ~48 m and the last sampling level was located at ~4 m. The velocity profiles were sampled at 30-min intervals, which is sufficient to resolve the major tidal harmonics. The ensemble time interval used for averaging the raw current velocity measurements was ~9 min, with 54 pings per ensemble. The total length of the ADCP record was approximately one...
year. To examine the vertical structure of tidal currents below 100-m depth, we used downward-looking ADCP observations collected at the same location in 2005–06 (mooring M1d). The instrument was deployed in September 2005 at a depth of ~108 m, so the first cell provided observations at ~115 m (Fig. 1, right). The vertical bin size was 4 m, with actual sampling depth ranges of 115–183 m for 18 measuring cells. The total length of the ADCP record was approximately one year with a one-hour time resolution of current profiling. The instrumental ensemble time interval used for raw current velocity averaging was 10 min, with 60 pings per ensemble.

In ADCP records, any single data gaps in velocity components that were marked as bad measurements by the ADCP software were filled in using linear time interpolation between the nearest two good data points. The total number of restored data was less than ~0.003% of the total number of measurements for 2004–05 and ~0.005% for the 2005–06 data. All data gaps in 2004–05 were located in the uppermost layer close to the surface. Data gaps in 2005–06 were concentrated in the last sampling cell at ~183-m depth. All current vectors were rotated by adding the magnetic deviation (9°31’W) determined from the International Geomagnetic Reference Field for the mooring position (see http://www.ngdc.noaa.gov/IAGA/vmod/).

According to the manufacturer’s description, the accuracies of current measurements with the 300-kHz ADCP are 0.5% of reading for magnitude and 2° for current direction. To check the actual instrument depth during both years of observations we used hydrostatic pressure records from the internal ADCP sensor. The accuracy of the ADCP internal pressure sensor is 0.25% of the full sensor range of 200 dbar.

The 2004–05 ADCP pressure record at mooring M1c has weak (~2 dbar) annual drift (not shown), which allowed us to use the entire record of current observations without correction for sampling levels. In 2005–06, there were five events of pronounced (up to ~60 m) deepening of the instrument, which occurred on 5–11 October, 15–25 October, 6–12 November, 13–27 June, and 7–17 July. Therefore, data collected before 15 November 2005 and after 1 June 2006 were eliminated from our analysis, and the total length of continuous current record used in this study was ~6.5 months.

Tidal currents and sea level amplitudes at the M1 mooring location simulated by Padman and Erofeeva (2004) using a barotropic inverse tidal model for the Arctic Ocean interior and marginal Arctic seas complemented our analysis. The output from this model is available online (at http://www.esr.org/ptm_index.html). The model uses a data assimilation routine of ~300 coastal tide gauges and satellite radar altimetry and has a 5-km horizontal resolution. The simulated sea level amplitudes for the four major constituents ($M_2$, $S_2$, $K_1$, and $O_1$) as well as the maximal tidal current of these four harmonics were linearly interpolated to the mooring location.

The Special Sensor Microwave Imager (SSM/I) satellite observations of sea ice concentration (taken from Maslanik and Stroeve 2010) were used to estimate the ice conditions at the M1 mooring location during the period of oceanographic measurements. The gridded dataset of sea ice concentration has 25-km equally spaced horizontal resolution and daily temporal resolution. The maps of monthly-mean sea ice concentration and time series of daily ice concentration were used to illustrate seasonal changes of the ice conditions over the Laptev Sea slope in 2004 and 2005.

In addition, the global one-degree-resolution monthly-mean temperature and salinity distributions from Polar Science Center Hydrographic Climatology (PHC; available from http://psc.apl.washington.edu/Climatology.html) (Steele et al. 2001) were used to examine the seasonal changes of vertical density stratification at the mooring location. The monthly 0–250-m-depth layer data were interpolated to the M1 mooring location using the inverse distance method (Shepard 1968).

3. Method of tidal analysis

Harmonic analysis was applied for the vectors of horizontal current according to the method described in Foreman (1978) using the MATLAB tide package (for details, see Pawlowicz et al. 2002). The standard properties of tidal currents (major and minor axes of tidal ellipses, Greenwich tidal phase, ellipse inclination, and corresponding 95% confidence intervals) were derived. In this study, the harmonic analysis for each level of observation was conducted using a 30-day running window shifted by the sampling intervals (0.5 and 1 h for 2004 and 2005, respectively). Derived characteristics of tidal currents for each window were assigned for the center of the window.

The M1 mooring was deployed in an area where the inertial frequency (0.0819 cph) is between the semidiurnal tidal frequencies of the $M_2$ (0.0805 cph) and $S_2$ (0.0833 cph) constituents, so that the $M_2$ and $S_2$ tidal waves are subinertial and superinertial, respectively, and their separation requires care because influence of oceanic background vorticity in the vicinity of critical latitudes on planetary vorticity may be important (D’Asaro and Morison 1992). Emery and Thompson (2001) estimated the frequency resolution $\Delta f_L = 1/L$ of the harmonic analysis applied to a record of length $L$, so that two constituents with frequencies closer than $\Delta f_L$ cannot...
Table 1. Tidal characteristics for the four major tidal constituents ($M_2$, $S_2$, $K_1$, and $O_1$) derived from the ADCP current record averaged over the upper ~50 m (2004–05) and 115–183 m (2005–06). Data from the M1 mooring (78°26′N, 125°40′E) are complemented by Padman and Erofeeva’s (2004) model estimates for the same location. All tidal properties are accompanied by corresponding 95% confidence intervals. SNR is determined according to Pawłowicz et al. (2002).

<table>
<thead>
<tr>
<th></th>
<th>$M_2$</th>
<th>$S_2$</th>
<th>$K_1$</th>
<th>$O_1$</th>
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<tr>
<td>Major axis (cm s$^{-1}$)</td>
<td></td>
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<tr>
<td>2004–05 (4–48 m)</td>
<td>1.2 ±0.3</td>
<td>1.8 ±0.3</td>
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<tr>
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<td>0.8 ±0.2</td>
<td>0.3 ±0.2</td>
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<tr>
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<td>0.18</td>
<td>0.07</td>
<td>0.04</td>
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<tr>
<td>Minor axis (cm s$^{-1}$)</td>
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<td>2004–05 (4–48 m)</td>
<td>0.7 ±0.3</td>
<td>0.7 ±0.3</td>
<td>0.0 ±0.1</td>
<td>0.0 ±0.1</td>
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<tr>
<td>2005–06 (115–183 m)</td>
<td>0.1 ±0.2</td>
<td>0.3 ±0.2</td>
<td>0.0 ±0.1</td>
<td>0.0 ±0.1</td>
</tr>
<tr>
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<td>0.00</td>
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<tr>
<td>Phase (°)</td>
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<td></td>
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<tr>
<td>2004–05 (4–48 m)</td>
<td>217 ±25</td>
<td>184 ±15</td>
<td>344 ±72</td>
<td>222 ±87</td>
</tr>
<tr>
<td>2005–06 (115–183 m)</td>
<td>32 ±31</td>
<td>156 ±23</td>
<td>346 ±38</td>
<td>315 ±49</td>
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<td>Inclination (°)</td>
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<tr>
<td>2004–05 (4–48 m)</td>
<td>93 ±26</td>
<td>86 ±13</td>
<td>88 ±77</td>
<td>69 ±70</td>
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<tr>
<td>2005–06 (115–183 m)</td>
<td>87 ±32</td>
<td>45 ±26</td>
<td>97 ±82</td>
<td>25 ±42</td>
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<tr>
<td>Model</td>
<td>95</td>
<td>95</td>
<td>93</td>
<td>98</td>
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<tr>
<td>SNR</td>
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<td>0.9</td>
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<tr>
<td>2005–06 (115–183 m)</td>
<td>4.5</td>
<td>11</td>
<td>2.6</td>
<td>0.3</td>
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</table>

be separated. For the 30-day window used for our analysis, $df = 0.00138$ cph; therefore, the applied window was sufficient to separate the tidal constituents $M_2$ and $S_2$ and the inertial oscillations, because the closest frequencies (inertial and $M_2$) are separated by $df = 0.0014$ cph, which is greater than $\Delta f_L$. Munk and Hasselmann (1964) proposed a more sophisticated estimate of the tidal record length, required for high-quality frequency resolution. They suggested that tidal frequency resolution may be estimated as $\Delta f_L = 1/(L \sqrt{\text{SNR}})$, where the square root of the signal-to-noise ratio (SNR) is taken. According to Pawłowicz et al. (2002), SNR can be computed based on the squared ratio of tidal amplitude to amplitude error. The year-long 2004–05 record and the 6.5-month-long 2005–06 record provide the correct separation of frequencies if $\text{SNR} \gtrsim 0.007$ and $\text{SNR} \gtrsim 0.02$, respectively. SNRs derived from standard MATLAB tidal package output for 2004–05 and 2005–06 are several orders of magnitude larger, thus confirming the robustness of the frequency resolution in our analysis (Table 1).

To ensure that the 30-day running window provides an accurate separation of the major tidal constituents, we repeated the tidal analysis for the current record with a 60-day running window. The experiment showed the robustness of our tidal analysis. For example, the correlation coefficients between the two series of tidal current amplitudes at 48-m depth based on a 30- and 60-day running window were $R = 0.89$ and $R = 0.78$ for the $S_2$ and $M_2$ constituents, respectively. The root-mean-square difference between these series was ~0.3 cm s$^{-1}$ for the $S_2$ constituent and ~0.1 cm s$^{-1}$ for the $M_2$ constituent, or ~13% of the current amplitudes for both constituents.

4. Tidal wave properties

Velocity components measured by the ADCP in the upper 50-m layer in 2004–05 and within the 115–183-m layer in 2005–06 show pronounced variability at different time scales resolved by the records (Fig. 2). The currents within the upper 50-m layer demonstrate enhanced variability during the last 1.5 months of observations in 2005, with the strongest variability (ranging from ~30 to 40 cm s$^{-1}$) at 4-m depth and somewhat increased variability at 48-m depth (Fig. 2). The current record at 48 m shows three events (in November 2004 and January and February 2005) of current amplification (up to 30 cm s$^{-1}$), which can probably be explained by mesoscale eddies (for details, see Dmitrenko et al. 2008).

a. Spectral analysis of current records

To attribute power to tidal oscillations we used rotary spectral analysis of the current records, providing separate estimates of the power for clockwise (CW) and counterclockwise (CCW) rotary components of the currents. For more accurate frequency resolution, we analyzed the entire length of the 2004–05 ADCP record, which was the longest series available to us. The spectra in Fig. 3 demonstrate the predominance of the CW component of the rotary spectra across a wide frequency range from ~0.07...
to \(-0.2\) cph; this is about an order of magnitude more energetic than the CCW component. Explaining the strong dominance of the CW rotation in the Northern Hemisphere, Gonella (1972) suggested a simple relationship between CCW and CW spectra, such that, for frequencies \(\omega\) close to the inertial frequency \(f\), this relationship tends to zero and the CW component dominates the spectra,

\[
S^+ (\omega)/S^- (\omega) = (\omega - f)^2/(\omega + f)^2,
\]

where \(S^+ (\omega)\) is the CCW spectral power and \(S^- (\omega)\) is the CW spectral power.

The spectral analysis showed that a substantial part of the energy is concentrated in periods from several days to months with a strong signal at the semidiurnal band of frequencies. It clearly identified \(S_2\) and \(M_2\) tidal harmonics (Fig. 3, top) and a lack of energy at the inertial frequency (the latter does not exclude possibility of significant bursts of near inertial energy with fairly random phase). The maximal spectral power (current) is concentrated at the \(S_2\) frequency, which dominates the \(M_2\) and diurnal frequencies for both the CW and CCW spectra (Fig. 3). Because the inertial oscillations are not evident in the CCW spectra in the Northern Hemisphere (Gonella 1972), we conclude that the predominance of the \(S_2\) tidal current over the \(M_2\) current is not caused by inertial or tidal current interference. CW spectra have wider peaks at semidiurnal frequencies than CCW spectra do (Fig. 3), which indicates a strong impact of baroclinicity (Munk 1997). Kulikov et al. (2004) also pointed out that wide semidiurnal peaks of the CW rotary spectra of tidal currents are an indication of baroclinic tides. The power of the diurnal frequency band is significantly weaker (by approximately an order of magnitude) than the power of the semidiurnal band (Fig. 3). Spectral peaks associated with high-order tides (with periods less than 0.5 days) caused by the nonlinear interaction of tidal constituents are also weak.

The spectral properties within the semidiurnal band of frequencies demonstrate significant changes throughout the year (Fig. 3, bottom). For example, during the last 1.5 months of the record, the energy associated with the \(S_2\) frequency is about 8 times higher for both the CW and CCW components of the rotary spectra than it is during the first 10.5 months of observations (Fig. 3, bottom). The increase of spectral power at the \(M_2\) frequency is weaker than the increase at the \(S_2\) frequency and is evident only for the CW spectral component.

In conclusion, we note that both the CW and CCW spectra demonstrate that \(S_2\) tidal currents dominate over \(M_2\) and diurnal currents within the upper \(-50\) m layer. However, because spectral analysis provides estimates of spectral power at discrete frequencies, which do not match exactly the frequencies of tidal constituents, the ratio between the powers at the \(S_2\) and \(M_2\) frequencies derived from the spectral analysis (Fig. 3) differs from what was obtained from the harmonic analysis of the current records (Table 1). The diurnal and high-order tidal currents as well as the inertial oscillations have weak spectral power in comparison with the semidiurnal constituents. CW power spectra peaks are wider than CCW peaks for the semidiurnal band of frequencies, indicating baroclinic tidal current behavior within the upper 50-m layer. The

![Fig. 2. Time series of (right) meridional and (left) zonal components of currents at (top) 4 and (bottom) 48 m measured by the upward-looking ADCP at mooring M1c (78°26'N, 125°40'W) in 2004–05.](image)

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strong increase of CW spectral power at the $S_2$ frequency during the last ~1.5 months of observations suggests that tidal properties are not static throughout the year.

b. Harmonic analysis of tidal currents

In this section, we provide results of our harmonic analysis of the ADCP records. The characteristics of the four major tidal constituents derived from harmonic analysis are presented in Table 1. Confirming results of the spectral analysis shown in Fig. 3, the harmonic tidal analysis shows dominance of the $S_2$ tidal current constituent over both the $M_2$ constituent and the diurnal current for both years of observations (Table 1). The $S_2$ tidal current derived from year-long ADCP records is ~1.5 times stronger than the $M_2$ current in the upper 50-m layer and ~1.3 times stronger in the 115–183-m layer. The amplitudes of diurnal tidal currents were found to be weak, just slightly exceeding the level of significance.

Because observations in 2004–05 and 2005–06 were made in different layers, differences of tidal current phase between these two layers probably reflect the impacts of stratification and boundary layer on the structure of tidal flow in the upper layer (Table 1). At the ~2690-m-deep mooring site, tidal properties within the 115–183-m layer (the deepest layer available for analysis) may be associated with the barotropic tides of the basin interior. The barotropic properties of tides in this layer are confirmed by better (in comparison with tidal properties in the upper ~50-m layer) agreement with the simulated tidal currents from the barotropic model, which used depth-averaged shallow-water equations (Table 1). We note that the barotropic model reproduces properties of $M_2$ tidal currents closer to observations in comparison with properties of $S_2$ tidal currents. Amplitudes of semidiurnal currents in the upper 50-m layer are about 2–2.3 times larger than those in the 115–183-m layer for both the $M_2$ and $S_2$ constituents, demonstrating the possibly significant baroclinicity of tidal currents. Within the upper 50-m layer, tidal properties are under the influence of strong seasonal stratification and thus may be strongly influenced by baroclinic tides.

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**FIG. 3.** (top) Rotary spectra (red line is for CW component of rotation and blue line is for CCW component) of depth-averaged ADCP currents within the upper 50-m layer at the mooring M1c (78°26′N, 125°40′E) in 2004–05. (bottom) Rotary spectra for the semidiurnal frequency band (left) for the period of compact ice conditions (September 2004–August 2005) and (right) for the ice-free period (August–September 2005). Horizontal black lines show the 95% confidence interval. Vertical dashed lines show the inertial frequency.
The contribution of baroclinic tide to the tidal currents in the upper 50-m layer in 2004–05 was estimated using amplitudes and phase of barotropic tidal currents derived from the 2005–06 record in the deep (115–183 m) layer, where the baroclinic currents are assumed to be relatively small. This assumption that in the deeper layer the tidal currents are barotropic probably works because at the mooring site the bulk of the ocean interior is much thicker than the upper layer. Tidal current $V(t)$ in the upper 50-m layer may be described as a superposition of barotropic and baroclinic tides,

$$V(t) = A_{S_0} \cos(\omega_1 t + \phi_{S_0}) = A_{bt} \cos(\omega_1 t + \phi_{bt}) + A_{bc} \cos(\omega_1 t + \phi_{bc}),$$

where $A_{S_0} \cos(\phi_{S_0})$ are the tidal current amplitude and phase; $A_{bt}$ and $\phi_{bt}$ are the amplitude and phase of the barotropic tidal current; $A_{bc}$ and $\phi_{bc}$ are the amplitude and phase of the baroclinic tidal current; $\omega_1$ is frequency of tidal constituent; and $t$ is time. From this, $A_{bc} \cos(\phi_{bc})$ may be estimated as

$$A_{bc}^2 = A_{S_0}^2 + A_{bt}^2 - 2A_{S_0}A_{bt} \cos(\phi_{S_0} - \phi_{bt})$$

$$\phi_{bc} = \tan^{-1}\left(\frac{[A_{S_0} \sin(\phi_{S_0}) - A_{bt} \sin(\phi_{bt})][A_{S_0} \cos(\phi_{S_0}) - A_{bt} \cos(\phi_{bt})]}{A_{S_0}^2 + A_{bt}^2 - 2A_{S_0}A_{bt} \cos(\phi_{S_0} - \phi_{bt})}\right),$$

where $\tan^{-1}$ is the inverse tangent function. Using this relationship, we found that the amplitude of the $M_2$ baroclinic tidal current in the upper ~50-m layer is ~1.8 cm s$^{-1}$, which is approximately 3 times higher than the amplitude of the $M_2$ barotropic current of ~0.6 cm s$^{-1}$ (Table 2). The confidence intervals of the amplitude and phase of the barotropic tidal current for both semi-diurnal $M_2$ and $S_2$ constituents were derived from the upper and lower borders of the 95% confidence intervals of amplitudes and phases of the barotropic and total tidal currents (Table 1). The extreme estimates of the derived baroclinic tidal amplitudes and phases in the uppermost 50-m layer are shown in Table 2 as the corresponding errors. Because these estimates are derived using nonlinear function, they exceed the corresponding errors of the total and barotropic tidal currents. Stronger $M_2$ baroclinic tidal currents in comparison with the total $M_2$ tidal currents derived from the upper 50-m layer record result from out-of-phase barotropic and baroclinic currents (Table 2). The $S_2$ baroclinic current amplitude is ~1.2 cm s$^{-1}$, which is 1.5 times higher than the barotropic tidal current amplitude. In contrast to the $M_2$ current, the $S_2$ baroclinic and barotropic tidal currents are in phase, so the total current due to this constituent is stronger than either the baroclinic or the barotropic tidal component alone (Table 2). Using spectral estimates of tidal currents, Kulikov et al. (2004) found approximately the same ratio (1.5–2.5:1) for the amplitudes of baroclinic to barotropic tides in the surface layer on the shelf break in the southern Beaufort Sea. Based on the squared ratio of the derived amplitudes of barotropic to baroclinic semi-diurnal tides, we conclude that the energy of baroclinic tidal currents is approximately 2 and 9 times larger than the energy of the barotropic tidal current for the $S_2$ and $M_2$ constituents, respectively.

The amplitudes of the tidal currents determined using a 30-day window demonstrate strong variability throughout the year-long record (Figs. 4, 5). The tidal currents within the uppermost 50-m layer vary across a wide range, from ~0.2 to ~10 cm s$^{-1}$. The tidal currents in the ocean interior (115–183 m) are more stable and vary only from ~0.1 to ~1.5 cm s$^{-1}$ (Figs. 6, 7). A strong increase of tidal currents occurred during the last ~1.5 months of observations in August–September 2005 (Figs. 4, 5). These amplified tidal currents (up to 10 cm s$^{-1}$) were evident at all observed levels for the $S_2$ constituent and for the layer above 12 m for the $M_2$ constituent, where the maximal shear of $M_2$ tidal currents was found. There were several additional events of strong (up to ~4–5 cm s$^{-1}$) $S_2$ tidal current amplification; particularly strong events occurred...
in mid-March 2005 and in late May 2005, but these events were not evident in the uppermost (<12 m) layer (Fig. 4). In the upper 16-m layer, the $M_2$ constituent generally (except for the last month) dominates over the $S_2$ constituent. Below 16 m, $S_2$ tidal currents are ~1.7 times stronger than $M_2$ currents. Stronger $S_2$ tidal currents at this location are confirmed by the 2005–06 mooring dataset, in which maximal tidal currents due to the $M_2$ constituent for the 115–183-m layer are ~1.3 times weaker than are the $S_2$ currents (Table 1 and Figs. 6, 7). Thus, we conclude that $S_2$ tidal currents dominate, at least in the ~20–200-m depth range over the Laptev Sea slope.

Ellipses of $S_2$ tidal currents provide additional evidence for the strong temporal variability of tidal currents (Fig. 8). We note that the depth of maximal tidal currents for this constituent (Fig. 8, black line) roughly follows the depth of the maximal vertical gradient of the potential density derived from the monthly PHC, which is associated with the bottom of the upper mixed layer (Fig. 8, green line; Steele et al. 2001). Within the layers above and below the depth of maximal $S_2$ tidal currents were found to be significantly weaker. It is possible that the differences between the depth of the upper mixed layer and the maximum $M_2$ current are due to the use of climatological temperature and salinity data, which may differ substantially from the actual temperature and salinity state of the upper ocean for any particular year. In the absence of upper-ocean temperature and salinity observations over the 2004–05 seasonal cycle, the use of PHC is our best choice. The $M_2$ tidal ellipses also show seasonal variations in the depth of maximal currents, which is possibly related to changes in properties of the upper mixed layer (Fig. 9). In contrast to the $S_2$ constituent, for which the maximal currents were found at the base of the mixed layer, strong $M_2$ tidal currents during the cold period (from October through May) occupied the entire mixed layer. Because of this difference, we used a different criterion to define the lower boundary of the layer with maximal $M_2$ tidal currents. This layer was defined as the depth at which the kinetic energy of the tidal current was 3 times lower than the kinetic energy of the surface tidal current (Fig. 9, black line). The lower boundary of the layer with maximal $M_2$ currents and the depth of the mixed layer show reasonable agreement, demonstrating significant seasonal deepening during the cold season.

The direction of tidal current rotation is generally anticyclonic for both the semidiurnal $M_2$ and the $S_2$ constituents. However, there are three months in which rotation occurs in the opposite (i.e., cyclonic) direction within the upper ~40-m layer. These are March and April 2005 for the $S_2$ constituent and July 2005 for the $M_2$ constituent (Figs. 8, 9). According to Carbajal (2004), a change in the direction of tidal current rotation was linked to changes in the velocity of the lower boundary of the mixed layer and the interior baroclinic flow. The changes in tidal current direction observed in this study may be related to similar changes in the interior baroclinic flow, which may affect the stratification and the depth of the mixed layer. Further studies are needed to understand the mechanism behind these changes and their impact on the upper ocean dynamics in the Laptev Sea.
of tidal current rotation may occur because of a shift in the position of a large-scale amphidromic tidal system. Because the amphidromic tidal system is individual for each tidal harmonic, the change of rotational direction for one constituent is not necessarily accompanied by a change for another constituent. Carbajal (2004) also demonstrated that additional friction is an important mechanism, which can facilitate change in the direction of tidal current rotation. Cyclonically rotated tidal currents are generally associated with tidal ellipses with small minor axes (Figs. 8, 9), so that, for example, small changes in ice conditions may lead to reduced or enhanced friction.

Fig. 5. As in Fig. 4, but for $M_2$.

Fig. 6. (top) Maximal tidal currents (cm s$^{-1}$) for the $M_2$ constituent and (bottom) the width of the 95% confidence interval. Data are from a downward-looking ADCP deployed at mooring M1d ($78^\circ26\text{'}N$, $125^\circ40\text{'}E$) in 2005–06.
in the upper-ocean boundary layer and a shift in the direction of tidal current rotation.

c. Tidal sea level oscillations versus tidal currents

Here, using the results of model simulations, we show that tidal sea level variations and tidal currents may be dominated by different constituents. For estimates of tidal sea level oscillations, we used modeling data derived from a regional tidal model of the Arctic Ocean (Padman and Erofeeva 2004). In contrast to the observed tidal currents dominated by the semidiurnal $S_2$ constituent in the upper 50-m and deeper (115–183 m) layers (Table 1), the model shows that the $M_2$ constituent dominates sea level changes at the ∼2690-m-deep M1 mooring location (Table 3). The model amplitude of the $S_2$ tidal component, ∼5 cm, is less than half the sea level amplitude of the $M_2$ constituent, ∼12 cm. The diurnal constituents $K_1$ and $O_1$ have amplitudes of ∼5 and ∼2 cm, respectively (Table 3). The model-based tidal currents at the mooring location are dominated by the $M_2$ constituent with current amplitude more than 2 times stronger than that of the $S_2$ constituent. However, we note that the model systematically underestimates the current amplitudes for the semidiurnal constituents. A particularly large (up to ∼1.6 cm s$^{-1}$) difference between simulated and observed currents in the upper ∼50-m layer is due to the impact of baroclinicity on the observed tides (Table 1). To ensure that the differences between the observed and simulated tidal currents are not due to poor representation of bottom topography in the model, we analyzed the cross-slope distribution of the sea level and current amplitudes for the four major tidal constituents along 125°E (Fig. 10). The model demonstrates the strong influence of the $M_2$ constituent on sea level changes and tidal currents as compared to the influence of the $S_2$ and diurnal constituents along the entire ∼1000-km length of the section. The maximal tidal current and sea level amplitude in the model are found close to the shelf break, which is ∼150 km away from the M1 mooring location. In conclusion, the model demonstrates that the $M_2$ tidal constituent is the primary driver of sea level change over the Laptev Sea continental slope. At the same time, observations show that depth-averaged tidal currents are primarily influenced by the semidiurnal constituent $S_2$.

5. Resonance amplification of tidal currents on steep topography

In this section, we explain the observed dominance of $S_2$ tidal currents over $M_2$ currents and the lack of such dominance in sea level changes using the linear theory of trapped long waves over variable topography. The processes of resonant amplification of tidally induced currents by local topography in the Arctic Ocean are well known (e.g., Hunkins 1986; D’Asaro and Morison 1992; Padman et al. 1992). In most cases, tidal amplification due to the impact of the continental shelf and slope on tidal waves is evident for the diurnal tidal constituents; the frequencies of which are close to half of the inertial frequency (Hunkins 1986; Chapman 1989). However, Gammelsrød and Rudels (1983) and later D’Asaro and Morison (1992)
found slope-related enhanced tidal variability over the Yermak Plateau for the semidiurnal tides as well. The spectrum of barotropic oceanic long waves includes the fundamental (zero) mode that exists at both subinertial and superinertial frequencies; the properties of which resemble Kelvin and Stokes (edge) wave properties at frequencies lower and higher than the inertial frequency (Huthnance 1975; Mysak 1980; Ke and Yankovsky 2010). Munk et al. (1970) suggested that semidiurnal tides may propagate along the continental slope as hybrid Kelvin–edge waves; the characteristics of such waves depend on the earth’s rotation and on slope properties. Ke and Yankovsky (2010) showed that propagation of a tidal wave as a hybrid Kelvin–edge wave may lead to strong spatial variability of the tidally induced currents due to convergence of the alongshore energy fluxes, especially in the proximity of the continental shelf and slope.

To examine the cross-slope modal structure of the sea level and current amplitudes of the barotropic tidal wave interacting with the Laptev Sea continental slope, we used a simple linear model proposed by Caldwell and Longuet-Higgins (1972) and further developed by Huthnance (1975). The full set of model equations and corresponding boundary conditions are presented in the appendix. The model momentum equations were obtained from the system of nonstationary linear vertically integrated shallow-water equations, in which the pressure gradients are balanced by the Coriolis force. The momentum equations were complemented by the continuity equation. The solution is sought in the form of a wave propagating along the slope. Eliminating velocity components from the continuity equation yields the second-order ordinary differential equation for the sea level amplitude [Eq. (A1)]. The program code for this model of trapped waves over the continental shelf and slope was developed by Brink and Chapman (1987) and was kindly provided by K. Brink (2010, personal communication). The model simulates the cross-slope structure of currents and sea level amplitudes for the specified depth distribution. The length of the model domain in the cross-slope direction was 1200 km with a horizontal resolution of 12 km. The model depths reproduce the major properties of the Laptev Sea slope.

**Table 3.** Estimates of sea level amplitudes and phases for the four major tidal constituents $M_2$, $S_2$, $K_1$, and $O_1$ by Padman and Erofeeva’s (2004) model for the M1 mooring location (78°26′N, 125°40′E).

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Frequency (cph)</th>
<th>Amplitude (cm)</th>
<th>Phase (degrees)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_2$</td>
<td>0.080 511 4</td>
<td>12</td>
<td>124</td>
</tr>
<tr>
<td>$S_2$</td>
<td>0.083 333 3</td>
<td>5</td>
<td>184</td>
</tr>
<tr>
<td>$O_1$</td>
<td>0.038 730 7</td>
<td>2</td>
<td>343</td>
</tr>
<tr>
<td>$K_1$</td>
<td>0.041 780 7</td>
<td>5</td>
<td>2</td>
</tr>
</tbody>
</table>
along 125°E, including its wide (~250 km) shelf and rapidly deepening continental slope (Fig. 11, bottom). For this bathymetry, the M1 mooring is located ~380 km from the shallow border placed at 75°N.

The cross-slope distribution of the simulated sea level amplitudes for both the $S_2$ and the $M_2$ constituents is shown in Fig. 11 (top). Following Dale et al. (2001) and Ke and Yankovsky (2010), we show the cross-slope amplitude distribution in dimensionless form normalized to a maximum absolute value for each constituent so that the range for all variables shown does not exceed an interval from $-1$ to $1$. The maximal simulated sea level amplitude for all constituents is found at the model shelf border and decreases rapidly toward the location of the M1 mooring, where sea level amplitudes for both $M_2$ and $S_2$ waves are nearly equal. However, the gradual decrease of $S_2$ sea level amplitudes is disrupted by additional low-amplitude (~10% of maximal amplitude) wave-like features with several peaks separated by ~400 km (Fig. 11, top). On average, the distance between two neighboring peaks is comparable to the width of the Laptev Sea continental slope; this phenomenon may be viewed as a resonance enhancement of the oscillation on sloping topography.

The structure of along-slope current amplitudes for the waves at subinertial and superinertial frequencies differs significantly. For the subinertial $M_2$ frequency, the maximal (negative) along-slope current coincides with a maximum sea level elevation on the shelf, with an additional local maximum close to the shelf break (at ~200 km) interrupting the monotonic decrease of current amplitudes toward open ocean (Fig. 11). Simulated sea level amplitudes at the M1 mooring were found to be nearly equal for both waves, with a small predominance of the $M_2$ wave. At the same time, the model shows the dominance of the $S_2$ current over the $M_2$ current at the M1 mooring location, thus confirming results of the harmonic and spectral analysis of the observational data. Along-slope current amplitudes for the superinertial $S_2$ wave have several extrema over the slope. One of them with a current amplitude of ~80% of the maximal current is found at $y = 480$ km (Fig. 11). Another extrema with

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**Fig. 10.** Cross-slope (along 125°E) distribution of (top) sea level amplitudes and (middle) tidal current amplitudes simulated by Padman and Erofeeva’s (2004) Arctic Ocean Tidal Inverse Model (AOTIM-5). (bottom) Model bottom relief. The colored symbols show observed amplitudes derived from upper 50-m ADCP data at mooring M1c in 2004–05. The length of vertical lines drawn from the center of each circle denotes the range of the 95% confidence interval. The observed tidal currents for the diurnal constituents $K_1$ and $O_1$ are indistinguishable from zero at this scale.
a current amplitude of ~30% of the maximal current is found at $y = 324$ km. This structure of current amplitudes is related to the wave-like feature in $S_2$ sea level amplitudes, with maximal tidal currents caused by maximal cross-slope changes of the sea level amplitudes. Even though the mooring location does not coincide exactly with these extrema, the $S_2$ current amplitude is significantly amplified at this location compared with the $M_2$ current amplitude (Fig. 11).

This difference in the solution for the $M_2$ and $S_2$ tidal waves may be explained by the change in the type of the governing Eq. (A1) from elliptic for the subinertial frequencies ($M_2$) to hyperbolic for the superinertial frequencies ($S_2$) (Dale et al. 2001). As a result, the solution changes its structure from an exponential decrease of sea level amplitude for the $M_2$ frequency to an oscillatory type for the superinertial frequency $S_2$. Therefore, the model simulated the oscillatory behavior of $S_2$ tidal characteristics with extrema; the locations of which depend on the wavelength and the slope properties. Thus, we conclude that an edge wave with relatively small amplitudes in sea level and wavelength comparable to the width of the continental slope may lead to a substantial increase of tidal currents for the superinertial harmonics, as we see from observations, which exhibit the dominance of superinertial $S_2$ currents over subinertial currents at the $M_2$ frequency. Simulations with the baroclinic version of the Brink and Chapman model with horizontally uniform density stratification from the PHC dataset (Steele et al. 2001) show no significant change of the cross-slope structure of tidal current and sea level amplitudes for both $S_2$ and $M_2$ waves compared with the barotropic experiment.

### 6. Impact of sea ice on tidal currents

In this section, we show that tidal currents may be amplified by local sea ice conditions. As we already noted in section 4b, the series of current amplitudes shows strong amplification of tidal currents during the last ~1.5 months of observations in August–September 2005 (Figs. 4, 5 and Table 4). The $S_2$ tidal currents increase more than 4 times (from ~1.5 to ~8.5 cm s$^{-1}$) in comparison with currents observed during the first 10.5 months of the record (Table 4). For the $M_2$ constituent the seasonal increase was not as strong (from ~1.2 to ~2.0 cm s$^{-1}$, or ~1.6 times); this is evidence for the selective nature of this amplification. We speculate that a possible mechanism for this amplification of the $S_2$ tidal current is the change of resonance properties of the upper-ocean boundary layer because of the very different properties of the upper-ocean boundary layer in the presence/absence of ice.

The monthly-mean ice concentration for the eastern Eurasian Basin from September 2004 to October 2005 is shown in Fig. 12. Typically, August–September is characterized by the minimum arctic sea ice concentration. The 2004–05 mooring record covers one winter season and two periods of seasonally reduced ice cover (September

### Table 4. Tidal characteristics and corresponding 95% confidence intervals for the four major tidal constituents $M_2$, $S_2$, $K_1$, and $O_1$ derived from the depth-averaged ADCP current record at the M1c mooring location (78°26′N, 125°40′E) in 2004–05.

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Major axis ($\text{cm s}^{-1}$)</th>
<th>Minor axis ($\text{cm s}^{-1}$)</th>
<th>Phase (degrees)</th>
<th>Inclination (degrees)</th>
<th>SNR</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_2$</td>
<td>1.2 ±0.3</td>
<td>0.7 ±0.3</td>
<td>217 ±25</td>
<td>93 ±26</td>
<td>13</td>
</tr>
<tr>
<td>$S_2$</td>
<td>1.8 ±0.3</td>
<td>0.7 ±0.3</td>
<td>184 ±15</td>
<td>86 ±13</td>
<td>24</td>
</tr>
<tr>
<td>$K_1$</td>
<td>0.1 ±0.1</td>
<td>0.0 ±0.1</td>
<td>344 ±72</td>
<td>88 ±77</td>
<td>1.4</td>
</tr>
<tr>
<td>$O_1$</td>
<td>0.1 ±0.1</td>
<td>0.0 ±0.1</td>
<td>222 ±87</td>
<td>69 ±70</td>
<td>0.9</td>
</tr>
</tbody>
</table>

**Entire record** (September 2004–September 2005)

| $M_2$       | 1.2 ±0.3                      | 0.7 ±0.3                      | 221 ±24         | 94 ±23               | 11  |
| $S_2$       | 1.5 ±0.3                      | 0.6 ±0.3                      | 185 ±17         | 87 ±14               | 21  |
| $K_1$       | 0.1 ±0.1                      | 0.1 ±0.1                      | 345 ±77         | 91 ±65               | 1.5 |
| $O_1$       | 0.1 ±0.1                      | 0.0 ±0.1                      | 220 ±84         | 68 ±90               | 0.7 |

**First 10.5 months** (September 2004–July 2005)

| $M_2$       | 2.0 ±1.3                      | 1.4 ±0.9                      | 137 ±80         | 107 ±63              | 2.2 |
| $S_2$       | 8.5 ±1.3                      | 5.1 ±0.8                      | 172 ±16         | 97 ±14               | 40  |
| $K_1$       | 0.3 ±0.2                      | 0.2 ±0.1                      | 178 ±62         | 160 ±91              | 1.4 |
| $O_1$       | 0.1 ±0.2                      | 0.0 ±0.2                      | 193 ±118        | 56 ±71               | 0.5 |

**Last 1.5 months** (August–September 2005)
2004 and August–September 2005). However, the maps in Fig. 12 demonstrate that ice-free conditions were observed over the M1 mooring location in August–September 2005, coinciding with the time when the $S_2$ tidal currents were amplified. The time series of ice concentration and upper 50-m tidal currents demonstrate a strong anticorrelation of $R = -0.73 \pm 0.05$ for the $S_2$ constituent (Fig. 13). The correlation of ice concentration with the tidal currents of other constituents (including $M_2$) was negligible at the 95% level of significance (Table 5). This selective correlation suggests that $S_2$ wave experiences selective amplification under ice-free conditions, possibly associated with the dynamics of superinertial waves. To confirm robustness of the identified relation between the ice concentration and $S_2$ tidal current amplitudes, we derived additional nonparametric Wilcoxon’s statistic (Wilcoxon 1945) reflecting degree of similarity between the two time series. The main advantage of this
statistic is the absence of an a priori hypothesis about internal structure of the series (i.e., their autocorrelation and nonstationarity). The test confirms the similarity between time series of ice concentration and tidal current at 95% confidence level, thus confirming the existing relationship between the ice concentration and tides.

A relationship between tidal current speed and ice conditions was previously reported by Kulikov et al. (2004) for the Canadian Arctic. Examining baroclinic tidal currents in the southeastern Beaufort Sea, the authors pointed out that both inertial currents and internal tides exhibit seasonal variability. We note, however, that the nature of this variability is different than that of the seasonal amplification found in our data. Kulikov et al. (2004) hypothesized that the properties of internal tides vary in response to seasonal change of vertical stratification; variability of inertial currents is caused by more effective atmospheric impact during the ice-free period.

The suppression of barotropic tides under compact ice condition caused by enhanced friction at the interface between the sea ice and the ocean was reported earlier (e.g., Zubov 1945; Kowalik and Proshutinsky 1994; St-Laurent et al. 2008). St-Laurent et al. (2008) provided an observational evidence of ~1.5 amplification of $M_2$ tidal current under ice-free conditions in a sound of the Hudson Bay (see their Fig. 6) and seasonal modulation of $M_2$ sea level amplitudes and phases at several locations within Foxe Basin, Hudson Strait, and Hudson Bay. The authors used a numerical model in order to investigate the factors responsible for this seasonal modulation of sea level oscillations and found that it is probably due to change of friction in the under-ice boundary layer. Our analysis does demonstrate approximately the same (~1.6) rate of amplification of $M_2$ tidal current under ice-free conditions and the pattern of this amplification with a maximum of amplification just at the surface (Fig. 5), the same pattern reported by St-Laurent et al. (2008). In contrast, the observed $S_2$ tidal currents demonstrate much stronger, four-fold, amplification and close to vertically uniform pattern of amplification in the upper ~50-m layer (Fig. 4). Rainville and Woodgate (2009) also found a strong impact of seasonally varied sea ice on the generation of baroclinic internal waves at the inertial frequency at the northern Chukchi Sea slope. In our case, the spectral power of tidal currents for the last 1.5 months of the record (Fig. 3, bottom) does not demonstrate a significant signal at the inertial frequency, so our finding of strong current amplification is not related to aliasing of tidal currents with seasonally amplified inertial motions. Thus, selective strong amplification of $S_2$ tidal currents suggests that change of local resonance properties may be a plausible explanation for the observed phenomena of the upper-ocean sensitivity of the baroclinic tides to local sea ice conditions.

7. Discussion and conclusions

Using two year-long ADCP current records in 2004–05 and 2005–06 at the M1 mooring located in 2690 m of water on the continental slope of the Laptev Sea, the properties of tidal currents within the upper ~200-m depth layer have been examined. Harmonic and spectral analyses of the current records showed that the maximal spectral power (current amplitude) was concentrated at the $S_2$ frequency, which dominated the $M_2$ frequency and the diurnal band of frequencies. The vertical structure of tidal currents demonstrates strong dominance of the baroclinic tide within the upper 50-m ocean layer, where the energy of baroclinic tidal currents is at least 2 times larger than the energy of barotropic tidal currents.

Modeling experiments demonstrated that the existence of wave-like features (edge waves) in the cross-slope distribution of level amplitudes leads to a series of local extrema in tidal currents. In these experiments, properties

| Table 5. Correlation coefficients for time series of sea ice concentration and depth-averaged tidal current. |
|-------|-------|-------|-------|
| $M_2$ | $S_2$ | $K_1$ | $O_1$ |
| Entire record (September 2004–September 2005) | $-0.02 \pm 0.1$ | $-0.73 \pm 0.05$ | $-0.15 \pm 0.1$ | $-0.18 \pm 0.1$ |
of continental slope and shelf width define the basic characteristics of the edge waves. Munk and Phillips (1968) proposed a mechanism for resonant amplification of internal tidal currents near the critical latitudes, where the tidal frequencies are close to the inertial frequency. This amplification mechanism enhances the CW current component and is effective only in a relatively narrow latitude band near the critical latitude (Foldvik et al. 1990). Here we found that similar amplification is possible for the semidiurnal band of tidal waves over the Laptev Sea continental slope.

Kulikov et al. (2004) found that the internal tides have small (in comparison with the amplitudes of the original barotropic tidal waves) variations related to the seasonal changes in stratification. We, however, found very strong temporal variability of tidal currents within the upper ~50-m layer, with fourfold amplification of the S2 tidal current associated with observed changes in local sea ice conditions. Amplification of the S2 tidal currents in August–September 2005 coincides with the seasonal clearance of the Laptev Sea from ice. Comparison of year-long series of sea ice concentration and S2 tidal current amplitudes showed strong anticorrelation, with $R = -0.73$. The $M_2$ tidal currents do not demonstrate such a strong anticorrelation. We speculate that this is probably due to the change of resonant properties of the upper-ocean boundary layer and conclude that the selective nature of amplification makes a resonant hypothesis of this amplification plausible. We further hypothesize that sea ice may be an important regulator of tidal current strength in this region. This is potentially an important finding, because a fourfold amplification of tidal currents under ice-free conditions leads to a 16-fold increase of kinetic energy, resulting in enhanced tidal energy dissipation and mixing rates in the upper-ocean layer. Arctic summer sea ice has recently exhibited an important finding, because a fourfold amplification of tidal currents under ice-free conditions leads to a 16-fold increase of kinetic energy, resulting in enhanced tidal energy dissipation and mixing rates in the upper-ocean layer. Arctic summer sea ice has recently exhibited a dramatic retreat (Comiso et al. 2008), and, in consequence, the Arctic Ocean slopes have become more and more exposed to seasonally ice-free conditions. Global climate models predict that the ice will retreat even more in the future. Therefore, the strong tidal amplification we found over the slopes under ice-free conditions may play an increasingly important role in the energy budget of the upper-ocean layer for these areas in the future (e.g., Wang and Overland 2009). However, achieving a detailed understanding of the mechanisms of this amplification, the rate of mixing, and the geographical areas where this amplification is possible requires more targeted and well-orchestrated observational and model experiments.

Acknowledgments. This study was supported by NASA Grant NNX08A050G and a JAMSTEC grant.

APPENDIX

A Barotropic Long-Wave Model

A barotropic wave interacting with the continental slope may be described by a system of ordinary differential equations for sea level amplitudes [(A1)] and cross- [(A2)] and along-slope [(A3)] components of current (Caldwell and Longuet-Higgins 1972),

\[ \eta_t(y) + \frac{h_t(y)}{h(y)} \eta(y) + \left[ \frac{\omega^2 - f^2}{gh(y)} \frac{h'(y) f k}{\omega} - k^2 \right] \eta(y) = 0, \] (A1)

\[ u(y) = \frac{-g}{\omega^2 - f^2} \left[ f \eta'(y) - 8k \eta(y) \right], \quad \text{and} \quad (A2) \]

\[ v(y) = \frac{ig}{\omega^2 - f^2} \left[ (\omega^2 - f^2) \eta'(y) - k \eta(y) \right], \] (A3)

where $\eta(y)$ is sea level amplitude; $\omega$ is wave frequency; $k$ is the wavenumber; $f$ is the inertial (Coriolis) frequency; $u$ is the along-slope velocity; $v$ is the cross-slope velocity; $g$ is the acceleration due to gravity; $h(y)$ is the depth; $y$ is the cross-slope coordinate; and $i$ is the imaginary unit.

The solution for $\eta$ is sought in the form of a wave propagating along the slope (along the $x$ coordinate) as $\eta(x, y, t) = \eta(y)e^{i(\omega t - kx)}$. The system (A1)–(A3) is complemented by the following boundary conditions: At the shelf border ($y = 0$), we used a solid wall condition for the cross-slope component of current [(A4)] and diminishing wave amplitudes toward the deep ocean [(A5)],

\[ y = 0; \quad u = 0 \quad \text{and} \quad (A4) \]

\[ y \rightarrow \infty; \quad \eta \rightarrow 0. \quad (A5) \]

Because an analytical solution of (A1)–(A3) with corresponding boundary conditions (A4)–(A5) exists only for limited classes of simple bottom profiles, we used the numerical solution of this problem with the code developed by Brink and Chapman (1987).

REFERENCES


