Internal Waves and Mixing in the Marginal Ice Zone near the Yermak Plateau*

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ABSTRACT

Observations were made of oceanic currents, hydrography, and microstructure in the southern Yermak Plateau in summer 2007. The location is in the marginal ice zone at the Arctic Front northwest of Svalbard, where the West Spitsbergen Current (WSC) carries the warm Atlantic Water into the Arctic Ocean. Time series of approximately 1-day duration from five stations (upper 520 m) and of an 8-day duration from a mooring are analyzed to describe the characteristics of internal waves and turbulent mixing. The spectral composition of the internal-wave field over the southern Yermak Plateau is 0.1–0.3 times the midlatitude levels and compares with the most energetic levels in the central Arctic. Dissipation rate and eddy diffusivity below the pycnocline increase from the noise level on the cold side of the front by one order of magnitude on the warm side, where 100-m-thick layers with average diffusivities of $5 \times 10^{-5}$ m$^2$ s$^{-1}$ lead to heat loss from the Atlantic Water of 2–4 W m$^{-2}$. Dissipation in the upper 150 m is well above the noise level at all stations, but strong stratification at the cold side of the front prohibits mixing across the pycnocline. Close to the shelf, at the core of the Svalbard branch of the WSC, diffusivity increases by another factor of 3–6. Here, near-bottom mixing removes 15 W m$^{-2}$ of heat from the Atlantic layer. Internal-wave activity and mixing show variability related to topography and hydrography; thus, the path of the WSC will affect the cooling and freshening of the Atlantic inflow. When generalized to the Arctic Ocean, diapycnal mixing away from abyssal plains can be significant for the heat budget. Around the Yermak Plateau, it is of sufficient magnitude to influence heat anomaly pulses entering the Arctic Ocean; however, diapycnal mixing alone is unlikely to be significant for regional cooling of the WSC.

1. Introduction

The oceanic heat transport to the Arctic constitutes a potentially important term in the delicate energy balance of the Arctic Ocean (Morison 1991). Warm and saline Atlantic Water (AW) carried northward with the West Spitsbergen Current (WSC) is the principal contribution of salt and oceanic heat to the Arctic Ocean (Aagaard et al. 1987; Schauer et al. 2008). On its northward path, the WSC cools and freshens through several processes. At the surface, heat is lost directly to the atmosphere and to melt ice (Aagaard et al. 1987; Boyd and D’Asaro 1994; Saloranta and Haugan 2004; Cokelet et al. 2008; Sirevaag and Fer 2009). Subsurface heat loss is comparable and is attributed to eddies diffusing heat along steeply sloping isopycnal surfaces (Boyd and D’Asaro 1994), to cross-shelf exchange with colder water (Saloranta and Haugan 2004; Nilsen et al. 2006; Cottier et al. 2007), and to turbulent mixing induced by currents and internal waves over topography (Padman and Dillon 1991; Sirevaag and Fer 2009).

The final passage of the WSC to the Arctic is over the Yermak Plateau (YP), a bathymetric feature northwest of Svalbard (Fig. 1). The YP is identified as a region of...
enhanced tidal variability for both diurnal (Hunkins 1986; Padman and Dillon 1991; Padman et al. 1992) and semi-diurnal tides (Gammelsrød and Rudels 1983; D’Asaro and Morison 1992). Earlier measurements typically obtained from buoys or ice stations drifting over the outer northern or central parts of the plateau characterized the YP as a region of enhanced internal-wave activity and mixing (e.g., Padman and Dillon 1991; Wijesekera et al. 1993).

The diverging isobaths south of the YP split the WSC and distribute the AW in three main branches (Quadfasel et al. 1987; Manley et al. 1992): the Svalbard branch (Aagaard et al. 1987; Manley et al. 1992; Cokelet et al. 2008) along the upper continental slope inshore of the plateau, the Yermak branch (Manley et al. 1992; Manley 1995) along the northwestern slope of the YP, and the recirculating branch in Fram Strait between 78° and 80°N flowing south with the Return Atlantic Current. According to the estimates of Manley (1995), only about 55% of AW enters the Arctic Ocean (about 35% in the Svalbard branch and 20% in the Yermak branch). The fate of the Yermak branch (i.e., the path and partitioning between the recirculation in Fram Strait and the Arctic inflow around or over the YP) is relatively uncertain (Perkin and Lewis 1984; Bourke et al. 1988; Muench et al. 1992; Manley 1995; Gascard et al. 1995). In this study, we present data from the southern YP contrasting previous observations of enhanced mixing from the northern flanks. Profoundly different mixing rates between the northern and southern flanks of the plateau suggest that the cooling and freshening of the WSC will depend on the path it follows over or around the plateau.

The aim of this study is to investigate the role of tides, internal waves and topography in mixing near the Yermak Plateau. To this extent, we use day-long time series of finescale and microstructure measurements at stations near the ice edge over the southern YP. The present work has implications beyond the study site, for the Arctic.
Ocean in general. Although peripheral regions such as the YP are home to anomalously large tidal velocities (30–40 cm s\(^{-1}\); Padman et al. 1992; Padman and Erofeeva 2004), the maximum tidal velocities over most of the central Arctic Ocean are sufficient (5–10 cm s\(^{-1}\); Kowalik and Proshutinsky 1993) to generate internal waves over suitable topography such as midbasin ridges and the continental shelf break. Including a simple representation of tidal mixing in a coupled ocean–ice model for the Arctic Ocean, Holloway and Proshutinsky (2007) show that tides enhance loss of heat from AW and have profound implications for Arctic hydrography and circulation. Furthermore, studies in the marginal ice zone (MIZ) can aid understanding how the decreasing trend in Arctic ice cover (Giles et al. 2008) will affect the vertical mixing. Currently, the deep Arctic Ocean away from coasts and submarine topographic features is a remarkably quiescent environment (Rainville and Winsor 2008; Fer 2009). Maintenance of this weak turbulent mixing in the interior Arctic Ocean is crucial for the cold halocline layer (Fer 2009), hence the ice cover, and the AW layer circulation (Zhang and Steele 2007). However, recent finding shows that large inertial waves, enhanced shear, and mixing are tightly related to the absence of sea ice (Rainville and Woodgate 2009). In a seasonally ice-free Arctic, vertical mixing can be comparable to the levels in the MIZ with potential impacts on the heat content in the upper-layer circulation, nutrients, and the ecosystem.

In this paper, we present the oceanographic context, hydrography, and mixing in the MIZ over the southern flanks of the YP and describe the frequency and wave-number content of the internal-wave field at the site. Using the observations, we estimate the contribution of diapycnal mixing over rough topography and discuss the implications for the regional and the Arctic Ocean heat budget.

2. Location and sampling

We use a subset of data collected aboard the R/V Håkon Mosby near the YP and eastern Fram Strait in summer 2007. The R/V Håkon Mosby is an open water vessel, so our sampling was limited by the ice edge, which was located over the southern YP during the cruise (Fig. 1). In this study, we analyze dedicated microstructure and fine-structure profiles from five stations, with each approximately one-day occupation, and data from an 8-day-long mooring close to the ice edge (Fig. 1), in the Arctic Front.

A total of 185 conductivity–temperature–depth (CTD) full-depth profiles were collected using a SeaBird Electronics (SBE) SBE911plus system. To correct the CTD-derived salinity, one salinity sample per cast was drawn and analyzed with a Guildline Portasal 8410 salinometer. The CTD sensor accuracies provided by the manufacturer are 1 dbar, 1 \times 10^{-3} {^\circ}C, and 3 \times 10^{-4} S (siemens) m\(^{-1}\) for pressure, temperature, and conductivity, respectively. Downcast CTD data, sampled at 24 Hz, were processed in 1-dbar (~1 m) vertical averages following standard postprocessing routines.

Microstructure profiles were collected in the upper 520 m or down to about 10 m above bottom for shallower depth using a loosely tethered free-fall profiler [model MSS90L (MSS)] manufactured by ISW Was sermesstechnik, Germany. The profiler was ballasted to fall at a rate of about 0.6 m s\(^{-1}\). The MSS was equipped with two airfoil shear probes, fast response conductivity and fast response temperature; an acceleration sensor; and conventional CTD sensors, all sampling at 1024 Hz. In total, 222 casts were made while the vessel was drifting with the engines and thrusters switched off. Postprocessing follows Fer (2006). MSS-derived CTD profiles were averaged to 10-cm vertical resolution and corrected against the SBE CTD from 86 joint casts separated by less than 1.5 h. Profiles of dissipation rate of turbulent kinetic energy (TKE) per unit mass \(\epsilon\) are obtained as 1-dbar vertical averages, assuming isotropy. Diapycnal eddy diffusivity, \(K_\rho = \Gamma \epsilon N^{-2}\), is calculated assuming a balance between the production of TKE, its dissipation \(\epsilon\), and the diapycnal buoyancy flux \(\Gamma \epsilon\) (Osborn 1980). We use the common value of \(\Gamma = 0.2\), corresponding to 17% mixing efficiency, and obtain the buoyancy frequency \(N = -\left[(g/\rho)(dp/dz)\right]^{-1/2}\) from sorted \(\sigma_\rho\) profiles smoothed over 4 m vertically. The noise level in \(\epsilon\) measurements is 3–5 \times 10^{-10} W kg\(^{-1}\).

A vessel-mounted RDI Narrowband 150-kHz acoustic Doppler current profiler (VM-ADCP) continuously collected 5-min-averaged current profiles in the upper 360 m in 4-m bins. Additionally, current measurements were made by expendable current profilers (XCP; Lockheed Martin Sippican Inc.; Sanford et al. 1993). The XCP samples relative horizontal velocity, compass, and temperature to a depth of 1500 m. Depth is inferred from time using a known fall rate at a vertical resolution of about 0.4 m.

During the cruise, five day-long stations were occupied (Fig. 1 and Table 1), during which microstructure profiles were collected approximately every 30–60 min with concurrent continuous VM-ADCP sampling. The ship was repositioned typically 5–7 times during each station, depending on the drift. VM-ADCP profiles were excluded during maneuvering the ship and when maintaining the position by dynamical positioning during a CTD cast, leaving gaps in the time series. Six XCP probes were successfully deployed at station 1 at approximately 4-h intervals. Only two out of the seven XCPs dropped at station 3 returned usable data.
Table 1. Summary of station position and sampling duration: $\Delta x$ is the mean $\pm 1\sigma$ standard deviation horizontal displacement from the mean station position; echo depth is the mean $\pm 1\sigma$ over the station duration; and 5-M is the mooring located near station 5.

<table>
<thead>
<tr>
<th>Station</th>
<th>Lat (N)</th>
<th>Lon (E)</th>
<th>$\Delta x$ (km)</th>
<th>Duration (h) MSS/ADCP</th>
<th>Echo depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>80°7.95'</td>
<td>4°18.10'</td>
<td>2.0 $\pm$ 1.3</td>
<td>24.1/24.3</td>
<td>1253 $\pm$ 57</td>
</tr>
<tr>
<td>2</td>
<td>80°25.14'</td>
<td>6°55.24'</td>
<td>0.9 $\pm$ 0.6</td>
<td>23.1/24.7</td>
<td>637 $\pm$ 9</td>
</tr>
<tr>
<td>3</td>
<td>80°34.0'</td>
<td>9°46.73'</td>
<td>1.5 $\pm$ 1.4</td>
<td>22.7/8.8</td>
<td>1000 $\pm$ 86</td>
</tr>
<tr>
<td>4</td>
<td>80°0.57'</td>
<td>9°59.25'</td>
<td>1.7 $\pm$ 1.2</td>
<td>24.4/18.8</td>
<td>484 $\pm$ 4</td>
</tr>
<tr>
<td>5</td>
<td>80°0.1'</td>
<td>5°58.10'</td>
<td>2.2 $\pm$ 1.2</td>
<td>24.2/4.6</td>
<td>879 $\pm$ 29</td>
</tr>
<tr>
<td>5-M</td>
<td>79°59.78'</td>
<td>5°55.95'</td>
<td>—</td>
<td>195.5/15.1*</td>
<td>889</td>
</tr>
</tbody>
</table>

* Duration of Microcat Longranger sampling.

Shipboard measurements were supplemented by a short-term mooring at 79°59.78’N, 5°55.95’E (collocated with station 5) at a depth of 889 m. The mooring was deployed at 1039 UTC 23 July 2007 and retrieved at 1424 UTC 31 July 2007. The mooring consisted of 19 SBE37 Microcats distributed between 99–873 m, a RDI short-term mooring at 864-m depths (576-m vertical range), using both three- and 5-M is the mooring located near station 5.

Wind speeds during the cruise were moderate to low, especially calm during the later stations, and exceeded 10 m s$^{-1}$ only for short stretches of time. Air temperatures were near or below zero near the ice edge at stations 1, 2, and 5, and between 3° and 6°C elsewhere over the YP. Marked differences between air and sea surface temperatures during stations 1–3 correspond to non-radiative heat fluxes [sum of sensible and latent heat flux using Fairall et al. (1996)] of 25 W m$^{-2}$ (station 1) and $-50$ to $-20$ W m$^{-2}$ (stations 2 and 3) where positive heat flux is directed into the ocean.

4. Garrett–Munk spectral model

In the stratified ocean interior, the internal-wave continuum is reasonably well represented by the empirical model of Garrett–Munk (GM; Garrett and Munk 1972; Cairns and Williams 1976). GM spectral description serves as a tool for both comparing different datasets and for inferring an order of magnitude estimate of diapycnal mixing using finescale parameterizations (section 9). The GM model has been revised through the years, and earlier observations were compared with various versions. In this paper, to be consistent in comparing with earlier work, we use three variants of the GM spectra: GM79 in the form given in Levine et al. (1985), GM75 (Garrett and Munk 1975), and GM76 (Cairns and Williams 1976). We use the frequency spectrum, the so-called moored spectrum, of GM79. We use the vertical wavenumber spectrum, the so-called dropped spectrum, of GM75 and GM76 for the horizontal current, shear (vertical derivative of horizontal current), and strain (vertical derivative of isopycnal displacement). GM spectra are characterized by a prescribed spectral slope and four model parameters; $E_{GM} = 6.3 \times 10^{-5}$ is the GM non-dimensional energy level, $b = 1300$ m is the vertical depth scale of exponential stratification profile $N(z) = N_0 \exp(-z/b)$ with amplitude $N_0 = 3$ cph, and $j_0$ is the non-dimensional vertical mode number. GM75 and GM76 spectra differ in the slope of the spectrum (2.5 and 2, respectively) and the vertical mode number $j_0$ (6 and 3, respectively). In GM79, following Levine et al. (1985), the four GM model parameters can be written in terms of an energy level $r$ and a wavenumber bandwidth parameter $t$:

$$r = E_{GM} b^2 N_0^2$$

and

$$t = \frac{j_0}{2N_0 b}$$

3. Environmental conditions

Meteorological data recorded by the ship’s mast and the tidal elevation for the duration of the cruise show the variation in conditions during the occupation of the different stations (Fig. 2). Tidal surface elevation is inferred from the 5-km resolution Arctic Ocean Tidal Inverse Model (AOTIM-5; Padman and Erofeeva 2004) for the position of station 5 (collocated with the mooring). This location is chosen to allow for comparison with the near-bottom pressure recorded by the deepest Microcat (gray trace in Fig. 2a). Keeping in mind that the Microcat is about 15 m above the sea bed (i.e., total pressure at the bottom is not resolved), the pressure anomaly compares well with the AOTIM-5 tidal elevation both in amplitude and phase and lends confidence for the model results. The first two stations were worked during neap tides, followed by stations 3–5 during transition to spring tides. Among all stations, station 5 had the largest tidal range, comparable to the spring tides. The mooring was deployed prior to station 1 and recovered after station 5.

Wind speeds during the cruise were moderate to low, especially calm during the later stations, and exceeded 10 m s$^{-1}$ only for short stretches of time. Air temperatures were near or below zero near the ice edge at stations 1, 2, and 5, and between 3° and 6°C elsewhere over the YP. Marked differences between air and sea surface temperatures during stations 1–3 correspond to non-radiative heat fluxes [sum of sensible and latent heat flux using Fairall et al. (1996)] of 25 W m$^{-2}$ (station 1) and $-50$ to $-20$ W m$^{-2}$ (stations 2 and 3) where positive heat flux is directed into the ocean.
The canonical GM value of the energy level is \( r_{GM} = 320 \text{ m}^2 \text{ cph} \) (standard midlatitude value). The values of \( r \) in the Arctic Ocean vary between 3 and 100 \( \text{ m}^2 \text{ cph} \), according to the data compiled by Levine et al. (1985).

The dimensional GM energy density in units of joules per kilogram is

\[
r = \frac{rN}{N_0} - 0.003 \left( \frac{N}{N_0} \right),
\]

where \( N \) is the local stratification. In comparing our data with GM (section 6), we use the canonical \( r_{GM} = 320 \text{ m}^2 \text{ cph} \): that is, not scaled for the local stratification. The moored frequency spectra of vertical displacement, total horizontal velocity, and the horizontal kinetic energy (HKE) density, expressed in cyclic units, can be written as

\[
\Phi_v(\omega) = 2\pi f \frac{(\omega^2 - f^2)^{1/2}}{\pi N} \quad \text{and}
\]

\[
\Phi_{HKE}(\omega) = 8\pi f N \frac{(\omega^2 + f^2)}{\omega^3(\omega^2 - f^2)^{1/2}},
\]

as a function of local values of the inertial frequency \( f \) and the buoyancy frequency \( N \).

The coherence between isopycnal displacements at two levels separated vertically by \( \Delta z \), the so-called moored vertical coherence (MVC), is

\[
MVC(\omega) = \exp\left[-2\pi f \Delta z \left( N^2 - \omega^2 \right)^{1/2} \right].
\]

This form is after Desaubies (1976), where coherence increases toward the local buoyancy frequency; that is, the GM approximation \( N^2 - \omega^2 \approx N^2 \) is not made in Eq. (6). In calculating GM76 shear and strain spectra, we follow the appendix in Gregg and Kunze (1991).

5. Hydrography and mixing

The main hydrographic feature interacting with the topography of the YP is the northern tongue of the WSC. The depth of the AW temperature maximum descends (i.e., submerges) as heat is removed from the layers closer to the surface through loss to the atmosphere and ice melt north of Svalbard (Aagaard et al. 1987; Cokelet et al. 2008). The temperature signal of the WSC can be seen in Fig. 3. The WSC, identified by water temperature above 3°C, follows the shelf break west of Svalbard and then
partly crosses the plateau northwest of Svalbard. Surface temperatures rapidly decline toward the ice edge, whereas subsurface temperatures suggest submerging of the warm current under the colder meltwater close to the ice edge.

In this section, salient hydrographic features and turbulence measurements are presented using station-averaged profiles. As a result of internal-wave-induced oscillations, the maxima in $T$ and $S$ (signature of the AW core) are smeared out when profiles are time averaged in depth coordinates. Averaging is therefore done over isopycnal surfaces with mean vertical separation of 2 m and then converted to depth coordinate using the mean density profile. Each profile is sorted before calculations. Average profiles of temperature $T$ and salinity $S$, together with the corresponding $T$–$S$ diagram, are contrasted for all five stations in Fig. 4. The site of measurements is characterized by the frontal structure of the marginal ice zone and the Arctic Front. Stations 3 and 4 are in the pathway of the Svalbard branch of the WSC, with a layer of AW ($T > 3^\circ$C and $S > 35$) capped by a slightly freshened, less than 50-m-thick surface layer (Fig. 4). Station 4 with a thick layer of AW is located closest to the core of the Svalbard branch, which can be identified by the warm temperatures in Fig. 3. At station 2, AW is submerged below a clearly defined fresher surface layer. AW is absent at stations 1 and 5 during the time of our measurements, suggesting they are between the Svalbard and Yermak branches, in the cold side of the front. Station 1, close to the ice edge, is the only station where surface temperatures approach the freezing point. There is a distinct change of slope in the $T$–$S$ curves near $\sigma_\theta = 27.95$, which contains the $T$–$S$ maxima of stations 1 and 5. At lower densities, there is a significant temperature and salinity gradient along isopycals between the cold and warm core of the front, and substantial cooling and freshening of the AW layer can be achieved by isopycnal mixing.

Isopycnally averaged profiles inferred from the microstructure profiler are shown for the five stations in Fig. 5. In addition to hydrography, Fig. 5 shows the gradient Richardson number (ratio of squared buoyancy frequency and squared velocity shear evaluated at 8-m vertical scale from the VM-ADCP data), dissipation $\epsilon$, eddy diffusivity $K_\rho$, turbulent heat flux $F_H = -\rho_0 C_\rho K_\rho \langle dT/dz \rangle$, and the turbulent activity index $I_A = \nu^{-1} N^2$. Here, $C_\rho$ is the heat capacity of seawater, $\langle dT/dz \rangle$ is the average temperature gradient, and $\nu$ is the viscosity. In a stratified flow, turbulence cannot overcome the stratification to produce significant diapycnal buoyancy flux (i.e., mixing) for $I_A \leq 20$, although the threshold is uncertain (Thorpe 2005).

The cold side of the front (stations 1 and 5), in relatively deeper water and close to the ice edge, is quiescent below about 150-m depth, where both $\epsilon$ and $K_\rho$ are close

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**Fig. 3.** Temperature fields at (a) 3- and (b) 100-m depth derived using all CTD and MSS stations (dots). Day-long station positions (open circles) and the ice edge on 25 Jul 2007 (dashed white–magenta line) are shown for reference. Contours are drawn using objective analysis with a decorrelation scale of 0.7° in the meridional direction and 1.5° in the zonal direction.
to the instrumental noise level (Fig. 5). Despite the elevated $\epsilon$ in the upper layers, strong stratification prohibits mixing across the pycnocline at stations 1 and 5 ($I_A < 20$ between 25 and 80 m and between 17 and 34 m at stations 1 and 5, respectively) and marginally at station 3 at about 50 m. In the MIZ, Fer and Sundfjord (2007) found significant correlations between the dissipation rate integrated over the mixed layer and the work done by stress under the ice and the wind work at 10 m. Assuming a similar correlation, enhanced dissipation in the upper layer at station 1 can be due to 2–3 times stronger wind (Fig. 2b). Note that stations 1 and 5 are closely located, with a separation of about 35 km, but are occupied in contrasting tidal forcing: station 1 during neap tides and station 5 approaching spring tides. The similarity of dissipation rate profiles and the lack of energetic turbulence suggest that tidal forcing is not significant in stirring the deep layers in the southwestern YP, keeping in mind that depth below 520 m is not sampled.

All casts at station 1 show a clear warming toward the surface, corresponding to a downward heat flux of about 5 W m$^{-2}$. This accounts for 20% of the estimated bulk surface heat flux of 25 W m$^{-2}$ directed into the ocean (section 3). The convergence of heat flux downward from the surface and upward in the thermocline (note the upward $F_H$ between 50 and 120 m in Fig. 5a) leads to warming of the surface layer and consequent melting of ice, which in turn enhances the stratification in the upper layer ($N$ reaches 15 cph). Deeper in the water column, the weak turbulent mixing is reflected in the weak heat flux profiles within ±2 W m$^{-2}$ for both stations 1 and 5.

Stations 2–4 all have $\epsilon$ and $K_\rho$ above the noise level and $I_A$ sufficiently large to allow for turbulent mixing at all depths. We hypothesize that these are linked to the shallower water depth and stronger tidal currents. The largest measured dissipation rates are at station 4, especially near the bottom, leading to strong mixing of cold near-bottom water with the core of the Svalbard branch. The $K_\rho$ averaged in the bottom 100 m is $10^{-4}$ m$^2$ s$^{-1}$, yielding an average heat flux of 7.5 W m$^{-2}$ directed downward. Near-bottom values of $K_\rho$ reach $3 \times 10^{-4}$ m$^2$ s$^{-1}$, and the maximum heat removed from the AW layer is 15 W m$^{-2}$. Both stations 2 and 3 show discernible, large $\epsilon$ between 400- and 500-m depths (i.e., about 150–250 m and 500–600 m above the seabed, respectively), with average $K_\rho$ of $5 \times 10^{-5}$ m$^2$ s$^{-1}$ and $F_H$ of −2.5 W m$^{-2}$ and with maximum cooling of AW of about 4.3 W m$^{-2}$. This turbulent mixing occurs below the AW core and contributes to cooling of the WSC.

Our microstructure observations can be compared to data collected from the Coordinated Eastern Arctic Experiment (CEAREX) Oceanography Camp (O Camp) on the northwestern flanks, where energetic mixing events with $\epsilon O(10^{-7})$ W kg$^{-1}$ and $K_\rho \sim 10^{-7}$ m$^2$ s$^{-1}$ led to heat
loss from the AW layer core of about 25 W m$^{-2}$ (Padman and Dillon 1991). Station 4, away from the YP but near the Svalbard shelf, has $K_\rho = 10^{-4}$ m$^2$ s$^{-1}$ in the bottom 100 m, but the resulting heat flux is about half that observed at the CEAREX O Camp. In stations 1 and 5, on the southern flanks of the YP, the eddy diffusivity is at the noise level below the $T$–$S$ maxima and the heat flux is one order of magnitude less than that on the northern flanks.
6. Moored spectra

Time series spanning 8 days from the moored instruments are used to describe the frequency content of the horizontal currents and the isopycnal displacements. Vertical isopycnal displacement \( \xi \) profiles are calculated relative to 24-h moving-averaged \( \sigma_0 \) profiles. Frequency spectra of \( \xi \) between 100- and 800-m depths (Microcat data) and the rotary component velocity spectra at 23-m depth (in the pycnocline, sampled by the Seaguard) are calculated (Figs. 6, 7).

Total horizontal velocity spectrum [sum of the spectra of clockwise (CW) and counterclockwise (CCW) components, equivalent to 2 times the spectrum of HKE density] is dominated by a peak at the diurnal frequency, followed by a slightly less energetic peak at the semidiurnal frequency (Fig. 6). Total velocity variance integrated in the \( K_1 \) and \( M_2 \) frequency bands (0.0365–0.0469 and 0.0781–0.0885 cph) are \( 7 \times 10^{-3} \) and \( 2.2 \times 10^{-3} \) m\(^2\) s\(^{-2}\), respectively. Corresponding HKE, obtained as density (1027 kg m\(^{-3}\)) times half the velocity variance, are 3.6 and 1.1 J m\(^{-3}\). For reference, 2 J m\(^{-3}\) corresponds to a root-mean-square horizontal velocity of about 6 cm s\(^{-1}\).

In the following, we compare our observations to the GM model, which is appropriate for deep ocean, away from boundaries, in which wave–wave interactions create an energy cascade in frequency–wavenumber space to create a continuum (section 4). Close to internal-wave generation sites, such as the YP, the GM model is not expected to be representative of the internal-wave field; however, it serves as a useful tool to compare our data with other internal-wave observations. At station 5, the slope of the velocity spectrum in the internal-wave continuum (frequencies between \( f \) and \( N \)) is consistent with GM79 (Fig. 6). A quantitative measure of the observed spectral level is obtained following Levine et al. (1985).

The GM79 spectral shape for horizontal velocity [Eq. (5) using local \( N \) and \( f \) but retaining \( r \) as a free parameter] is fitted to the observed spectrum by adjusting \( r \). The energy level obtained by least squares fitting spanning frequencies 0.1–1 cph is \( r = 103 \) m\(^2\) cph, 0.32 times that of the canonical GM79 value (\( r_{GM} \)). The southern YP is thus less energetic than typical midlatitude open ocean. The range of \( r \) in the central Arctic spans 3–100 (Levine et al. 1985); hence, the southern YP is comparable to the most energetic Arctic internal-wave field. The rotary frequency spectra show that the CW-rotating component dominates particularly in the diurnal but also in the semidiurnal band. The ratios of CW to CCW variance at the corresponding bands are 37 (\( K_1 \)) and 5 (\( M_2 \)). This is comparable to the rotary spectra of the YP section from the Arctic environmental drifting buoy (AEDB) drift (Plueddemann 1992, his Fig. 13).

Time series of isopycnal depths show relatively smooth curves between 100 and 200 m and higher-frequency displacements between 250 and 450 m (Fig. 7a). Frequency spectra of isopycnal excursion, like the velocity spectra, are less energetic than the GM79 level and approximately follow the GM79 slope (Fig. 7b). In our calculations, we use displacements inferred from the density profiles, not the temperature records alone, following the observations of Foster and Eckert (1987) that, because of intrusive features, temperature spectra cannot be used to determine internal-wave activity of the upper ocean in MIZ. Spectra from three chosen depth ranges, marked in Fig. 7a, are scaled by the local buoyancy frequency and averaged (Fig. 7b). In agreement with the velocity spectrum, the diurnal band dominates, and there is significant variance contained in the semidiurnal band. When fitted to the GM79 shape between 0.1 and 1 cph, the spectra are 0.11, 0.17, and 0.12 times GM79. The curve fit to the average spectrum gives \( r = 42 \) m\(^2\) cph (Fig. 7b).

The frequency spectra for the horizontal current and vertical displacements can be compared to those obtained from previous drifts. The frequency spectrum of velocity observed in the pycnocline resembles those reported by Wijesekera et al. (1993) for the CEAREX O Camp drift.
Levine et al. (1985) recorded isotherm displacements for 5-days duration as the Fram-III camp drifted on the northern flank of the YP over about 800-m isobath (marked by the black circle in Fig. 1). Their measurements between 49 and 149 m covered the thermocline. Although the slope of the frequency spectra of isotherm displacements were comparable to GM79 slope, the energy level was 0.15–0.24 times $r_{GM}$ ($r = 47$ and 75 m$^2$ cph at 67 and 137 m, respectively). These observations are in agreement with the spectra reported here. Eckert and Foster (1990) reported that the vertical displacement spectrum was about 0.1 times GM79 at low frequency but increased to 3 times GM79 at higher frequencies near the average buoyancy frequency, based on data from the MIZEX84 program, which sampled close to our mooring location.

7. Vertical coherence

For a random, linear internal-wave field, coherence reduces with increasing vertical or horizontal separation. This reduction is proportional to the wavenumber bandwidth. For a broad wavenumber bandwidth, an increasing
number of waves with various wavelengths will interact with increasing spatial separation, acting to obscure the coherent wave motion. We use the isopycnal displacement time series inferred from the moored Microcats to analyze coherency with increasing vertical separation. We contrast the structure in the upper and middle layers (1 and 2, respectively, marked in Fig. 7) with relatively smooth displacements versus displacement with relatively high-frequency oscillations. Isopycnals with the mean depths of 160 and 300 m are chosen, and the coherence spectra are calculated between deeper isotherms with 20-m mean separation increments. Frequency distributions for the first three pairs are shown in Fig. 8. There is a broad low-frequency band from the near-inertial frequency to about 0.4 cph, showing high coherence. Coherency at 20- and 40-m vertical separation in the low-frequency band is in remarkable agreement with GM79. For larger separations, coherence falls rapidly (suggesting a broad wave-number bandwidth). There is an increase in coherence near the local $N$. Deeper in the water column, starting with the isopycnal at 300-m depth, there is a band of high frequency (about 0.5 cph) with high coherence (Fig. 8b). It should be noted that the tides are a significant fraction of the velocity variance (section 6), and the high coherence
in the low-frequency band as well as the vertical coherence scales may be biased by the highly correlated barotropic and low-mode baroclinic tides.

Mean coherence and phase profiles in the two indicated frequency bands are calculated up to 160-m vertical separation. The low-frequency coherence persists, different than zero at 95% confidence, down to about 120 m, but it drops steeper than the GM curve for both depth ranges. The high-frequency band loses coherence at about 60-m separation and drops steeper than the low-frequency band average profile. In all cases, the oscillations are in phase at all depths within \pm 20°. The structure is reminiscent of Levine et al.’s (1985) observations suggesting packet-like waves excited by the internal tide over the YP with short vertical range of in-phase coherence but in their case at much higher frequency (2.6 cph).

Our results can be compared to AIWEX (Beaufort Sea in the Arctic Ocean) and CEAREX (northern YP). During AIWEX, coherences were found to be very low compared to typical midlatitude open ocean values (Levine et al. 1985), implying a factor of 10 higher wavenumber bandwidth than at lower latitudes. The highest coherency was found near the local buoyancy frequency. The high bandwidth was consistent with independent estimates from vertical wavenumber spectra of HKE (D’Asaro and Morehead 1991). The high coherence in the low-frequency band apparent in our data is in contrast to Levine et al.’s (1985) observations suggesting packet-like waves excited by the internal tide over the YP.

8. Dropped spectra

Vertical profiles of horizontal velocity and density are used to derive vertical wavenumber spectra of horizontal velocity, velocity shear, and strain. Shear spectra are calculated using the 2-m vertical-resolution XCP profiles, limiting the analysis to stations 1 and 3. Rotary velocity spectra over 512-m (256 data points) segments are converted to shear spectra by multiplication with \((2\pi k_z)^2\), where \(k_z\) is the cyclic vertical wavenumber. The shear spectra are normalized by the average buoyancy frequency \(N^2\) in the corresponding depth range. Horizontal velocity spectra for stations 1 and 3 are shown in Fig. 9 and compared to the levels observed in STREX (D’Asaro 1984) and AIWEX (D’Asaro and Morehead 1991). STREX data collected during a period of storms in the northeastern Pacific Ocean are representative of a high range of GM levels. AIWEX data collected in the Beaufort Sea are typical of the quiescent Arctic Ocean. These two curves were used in D’Asaro and Morison (1992) and serve as a good reference for comparison.

We infer the energy level and the vertical wavenumber bandwidth from dropped spectra following D’Asaro and Morehead (1991). We retain both \(E\) and \(j_s\) as free parameters and obtain the best-fit values to the GM75 curve in the least squares sense in the range 0.004–0.1 cpm. The resulting values are \(j_s = 16\) and \(E = 3.4 \times 10^{-6}\), about 6 times wider and 19 times less energetic than GM75 at station 1, and \(j_s = 7\) and \(E = 1 \times 10^{-5}\), roughly 1.2 times wider and 6 times less energetic than GM75 at station 3. For comparison, AIWEX spectra adhere to \(j_s = 60\) and \(E = 1.57 \times 10^{-6}\). These low levels are consistent with the isopycnal displacement frequency spectra, which is about 7 times less energetic on average (Fig. 7b). Clearly, the vertical wavenumber spectral shape and the bandwidth at the YP are significantly different than the central Arctic Ocean.

D’Asaro and Morison (1992) reported vertical wavenumber spectra inferred from XCP data deployed during the MIZEX83 drift. The MIZEX83 drift passes across a 500-m-deep seamount where measured currents increased markedly (D’Asaro and Morison 1992). There, D’Asaro and Morison (1992) found that the level of the mean vertical wavenumber spectrum was about one order of magnitude above the GM level. The cross-YP drift of the AEDB between 22 November and 22 December 1987 [portion marked by 11–22 to 12–22 in Fig. 7 of Plueddemann (1992)] highlights the variability of relative velocities over the plateau (Fig. 15 of Plueddemann 1992). Between 12 and 15 December 1987, as the AEDB drifts near the seamount, very close to the MIZEX83 drift (Fig. 1), energetic oscillations in relative velocity were observed, which is consistent with the MIZEX83 observations. D’Asaro and Morison (1992) suggested that enhanced shear levels were caused by a baroclinic internal tide that was generated on the seamount by the barotropic tide, trapped to the seamount by the barotropic vorticity field. Our data are less energetic and suggest large variability over the plateau and that the enhanced energy level in MIZEX83 is likely a localized feature related to the seamount. Using the present dataset, we cannot conclude with confidence whether the variability is temporal or spatial. The velocity time series acquired by the AEDB shows that the relative velocity is significantly less energetic after the seamount, as the buoy proceeds toward the southern flanks. This is consistent with our observations and suggests significant spatial variability over the YP.

Corresponding vertical wavenumber shear spectra are compared to GM76 in Fig. 9b for the CW and CCW rotating shear. At 512-m scale \((k_z = 2 \times 10^{-3} \text{ cpm})\), the ratio of CW to CCW variance, \(r_{\text{CW-CCW}}\), is 1.1 and 0.6 in
the upper water column at stations 1 and 3, respectively. Near the bottom 512 m, CCW variance is 3.6 and 2.1 times more energetic for stations 1 and 3, respectively, suggesting upward energy propagation. Unfortunately, microstructure profiles do not penetrate below 520 m and do not resolve the lower 512-m portion of the water column where XCP spectra indicate upward energy propagation. We therefore cannot assess the importance of this feature for mixing in the region. However, the spectra are not elevated above the GM76 level, and finescale parameterizations (section 9) would lead to insignificant dissipation levels accordingly.

9. Finescale parameterizations

Measuring dissipation by direct means, such as microstructure profiles, is not a widely used method. In a slowly varying wave field, the rate of energy dissipation resulting from wave breaking can be approximated from the net energy transfer toward smaller scales associated with nonlinear interactions (Gregg 1989). The advantage of relating to internal-wave energy is that $\epsilon$ and $K_p$ can be estimated from relatively simple routine measurements. During the last two decades, this principle has been used for developing finescale parameterizations for $\epsilon$ based on the theoretical GM model using a quadratic scaling with finescale shear variance (Gregg 1989), or strain variance (Wijesekera et al. 1993) or, more generally applicable, using both shear and strain (Polzin et al. 1995; Gregg et al. 2003). Shear variance estimates, however, are dominated by instrument noise in weak stratification, and care is needed in estimating turbulence quantities from internal-wave shear, particularly using acoustic data that are prone to be noisy (Kunze et al. 2006). In the abyssal Southern Ocean, for example, diffusivity estimates using both shear and strain variance are one order of magnitude higher than estimates based on strain variance alone (Kunze et al. 2006, their Fig. 16). This suggests that the applications of finescale parameterization involving shear will be dominated by noise in weakly stratified environments, resulting in spuriously large
diffusivities such as those inferred in the abyssal Scotia Sea (Naveira Garabato et al. 2004b). The scaling by finescale shear variance is also shown to fail close to generation sites (Klymak et al. 2008). With these caveats in mind, a convenient formulation of the finescale parameterization for eddy diffusivity (using $K_{p} = 0.2 c N^{-2}$) is (Kunze et al. 2006)

$$K_{p} = K_{0} \frac{\langle V_{z}^{2} \rangle}{\langle V_{z}^{2} \rangle_{GM}} h(R_{w}) L(f, N),$$  

(7)

where $K_{0} = 5 \times 10^{-6} \ m^{2} \ s^{-1}$, $\langle V_{z}^{2} \rangle$ is the observed finescale shear variance, $\langle V_{z}^{2} \rangle_{GM}$ is the corresponding GM value,

$$h(R_{w}) = \frac{3}{2 \sqrt{2} R_{w}} + \frac{1}{R_{w} - 1},$$

and

$$L(f, N) = \frac{f \ arccosh \left( \frac{N}{7} \right)}{f_{30} \ arccosh \left( \frac{N_{0}}{f_{30}} \right)}.$$  

Here, $L(f, N)$ accounts for the latitudinal variation, $N$ is the local buoyancy frequency, $f$ is the local inertial frequency, $N_{0} = 5.2 \times 10^{-3} \ s^{-1}$ (=3 cph) is the reference stratification, and $f_{30}$ is the inertial frequency at 30° latitude. In Eq. (7), $h(R_{w})$ corrects for the variation in shear–strain ratio

$$R_{w} = \frac{\langle V_{z}^{2} \rangle}{N^{2} \langle V_{z}^{2} \rangle},$$  

(8)

The quadratic shear variance ratio in Eq. (7) is equivalent to $(0.1/k_{c})^{2}$, where $k_{c}$ is the high wavenumber cutoff up to which the integrated shear variance is about $0.7 N^{2}$ (0.1 cpm is the GM value for $k_{c}$).

VM-ADCP current profiles are not of sufficient quality to integrate the vertical wavenumber spectra to order 10-m vertical scales. XCP probes, on the other hand, measured velocity at station 1 with high resolution and also resolved the tidal cycle with six casts over approximately 20 h. We therefore use station 1 XCP-derived shear and MSS-derived isopycnal displacement spectra to evaluate shear and strain. Spectra are calculated using 128-m length segments to obtain profiles of shear–strain ratio $R_{w}$, cutoff wavenumber $k_{c}$, and dissipation rate (Fig. 10). We obtain $k_{c}$ as the wavenumber at which the integrated shear spectrum reaches $0.7 N^{2}$. When evaluated at $k_{c}$ (i.e., using shear and strain variances integrated to $k_{c}$ instead of 0.1 cpm), $R_{w} = 4.5$. When evaluated at 10-m scale, $R_{w} = 11 \pm 4$ but is 2.9 between 125- and 170-m depths. For comparison, the GM value is $R_{w} = 3$; for the CEAREX data, using Table 4 of Wijesekera et al. (1993), we infer average $R_{w} = 1.5$ both in the upper thermocline (between 100 and 170 m) and between 100 and 270 m (i.e., significantly lower than our data). Data from dedicated surveys conducted at mid-latitudes (Polzin et al. 2003) and high latitudes (Naveira Garabato et al. 2004a,b; Daae et al. 2009) show that $R_{w}$ varies between 3 and 14.

The cutoff wavenumber is a fairly constant value of about 0.04 cpm below 170 m and rises to much larger value above 150 m. Finescale parameterization gives dissipation rates comparable to the station-averaged profiles; however, the variability in $\epsilon$ at this station is not sufficient to conclude on the skill of the model. The ratio of CW to CCW variance at 128-m length scale suggests upward energy propagation in the deepest part of the profile, which is consistent with the bottom 512-m spectra (Fig. 9b).

D’Asaro and Morison (1992) used XCP-derived 10-m shear over the YP using the Gregg (1989) version of Eq. (7) (i.e., $h = 1$ and $L = 1$). They inferred $K_{p} \sim 10^{-4} \ m^{2} \ s^{-1}$ near the seamount close to the passage of MIZEX83 drift marked in Fig. 1. From our data, the average shear–strain ratio for 10-m vertical scale is $R_{w} = 11$, giving $h = 0.37$. This is partially compensated by the latitude dependence giving $L = 1.7$, leading to a total correction $hL$ of a factor of 0.6. Using the present values of $R_{w}$, $h$, and $L$ yields a revised estimate of the upper range of diffusivity of about $6 \times 10^{-5} \ m^{2} \ s^{-1}$, comparable to the near-bottom average values observed at stations 2 and 3.

10. Implications for regional scale and the
Arctic Ocean

We assume that the elevated mixing between 400 and 500 m observed at stations 2 and 3 is typical of the mixing over rough topography (away from abyssal plains) in the Arctic Ocean, resulting in a downward heat flux of 2–4 W m$^{-2}$. This is exactly the same range suggested by D’Asaro and Morison (1992), supported here by direct measurements. Integrated over an area of $3 \times 10^{12} \ m^{2}$ of the Arctic Basin, excluding abyssal plains but including ridges and continental shelf break regions, a total of 6–12 TW is thus lost through internal-wave-induced mixing. Because this downward heat loss is below the depth of temperature maximum in the Atlantic layer, it will not contribute to the surface heat flux, but it will be lost to the Arctic Intermediate Water below, which leaves the basin in the East Greenland Current. Because stations 2 and 3 were occupied during neap tides and in calm weather, inferred 6–12 TW of heat loss is likely a conservative estimate. Stronger forcing during spring tides and enhanced near-inertial shear following strong winds in the absence of sea ice (Rainville and Woodgate 2009) are expected to elevate the internal-wave-induced mixing.

On the regional scale, our measurements combined with the previous work from the northern YP show that
the total heat loss at the YP resulting from diapycnal mixing ranges within 2–25 W m\(^{-2}\). The range mirrors the lateral and temporal variability over and around the plateau and corresponds to the trend in ocean-to-ice heat flux observations by McPhee et al. (2003) showing heat fluxes ranging from 0 to 110 W m\(^{-2}\). Their buoys drifting from the northeastern to western YP recorded the largest surface heat flux at the northeastern and western flanks and showed less activity at the central part of the plateau. The upper limit of 25 W m\(^{-2}\) (Padman and Dillon 1991), typical for the northern flanks, is indeed comparable to the average surface heat flux of 22 W m\(^{-2}\) inferred by McPhee et al. (2003). If we account for the lateral variability by assuming that 30% of the plateau area is characterized by enhanced diapycnal fluxes of 20 W m\(^{-2}\) and the remaining 70% is characterized by more typical 3 W m\(^{-2}\), the weighted average is 8 W m\(^{-2}\). This corresponds to 0.3 TW using an approximate area of \(4 \times 10^{10} \text{ m}^2\) (200 km \(\times\) 200 km) for the YP. The heat loss resulting from diapycnal mixing over the YP thus compares with the heat loss required to cool 1 Sv (1 Sv = \(10^6 \text{ m}^3 \text{ s}^{-1}\)) of AW by about 0.1 K. Although small, this cooling can be important for the variability of heat anomalies entering the Arctic Ocean through Fram Strait and constitutes about 15% of Rudels et al.’s (2008) estimated 2-TW storage rate in the Arctic (i.e., change in heat content of the AW layer). Deep diapycnal mixing at the Yermak Plateau alone (0.3 TW), however, cannot explain the strong cooling of the WSC north of Svalbard. When compared to the heat lost to ice and atmosphere in this region (14 TW in total; Rudels et al. 2008), internal-wave-induced mixing is likely not significant on the regional scale.

11. Summary

Observations over the southern part of the Yermak Plateau (YP) highlight the characteristics of hydrography, internal waves, dissipation, and turbulent mixing during summer 2007. The plateau northwest of Svalbard is of interest because it is the main topographic obstacle for the Atlantic inflow to the Arctic via the West Spitsbergen Current (WSC). Measurements were made from five day-long stations and an 8-day-duration mooring in neap to early spring tides. The sampling suffers from not resolving the strong forcing during spring tides and the dissipation rates below 520 m. The latter will fail to capture plausible bottom-enhanced mixing on deeper slopes. With these caveats in mind, a summary of our results is as follows.

Previous reports on energetic tidally induced internal waves observed over the northern YP suggested that diapycnal mixing over the YP might play a significant role in cooling of the WSC in the area north of Svalbard. Overall, diapycnal mixing at our study site is less energetic than that in the northern YP. Furthermore, on the southern YP, we
found significant differences between stations on either side of the Arctic Front, where dissipation rate and eddy diffusivity below the pycnocline increase from the noise level (on the cold side) by one order of magnitude (on the warm side). Station 4, close to the Svalbard shelf inshore of the plateau, reveals elevated diffusivity by another factor of 3–6, but this cannot be linked clearly to the internal waves using the present dataset. Stations 1 and 5 (cold side of the front), occupied on neap and spring tides, respectively, show no evidence of increased stirring during spring tide, with the caveat that only the upper half of the ~1200-m water column was sampled. Dissipation at the pycnocline at these stations is well above the noise level, but strong stratification prohibits mixing across the pycnocline. Stations 2 and 3 (warm side of the front) show 100-m-thick deep layers with average diffusivity of $5 \times 10^{-5}$ m$^2$ s$^{-1}$ and heat loss from the Atlantic Water (AW) layer of 2–4 W m$^{-2}$. We suggest that these values are representative of the mixing in the Arctic over rough topography. Inshore of the plateau, at the core of the Svalbard branch of WSC near-bottom mixing removes 15 W m$^{-2}$ of heat from the Atlantic layer.

Both the horizontal velocity and vertical displacement frequency spectra show significant peaks at the semidiurnal and diurnal periods. Over the southern YP, the internal-wave field is found to be less energetic and less coherent than that typical of midlatitudes (GM; Garrett and Munk 1972). The frequency content of the internal-wave field is 0.1–0.3 times the corresponding GM level. This is consistent with previous observations close to our study site (Levine et al. 1985; Eckert and Foster 1990). Although these levels are much lower than those typical of midlatitude and of the northern flanks of the YP (Plueddemann 1992), they compare to the highest levels in the central Arctic.

Vertical displacements of isopycnals vertically separated by 20 m show significant coherence at low frequencies consistent with GM up to about 40-m vertical separation. Coherence decreases with increasing separation, more rapidly than that predicted by GM, suggesting a wider wavenumber bandwidth. Analysis of vertical wavenumber spectra suggests that the wavenumber bandwidth is indeed 1.2–6 times wider than GM.

Application of the Gregg et al. (2003) finescale parameterization to infer dissipation rate from the measured shear and strain fields predicts dissipation rates within a factor of 2 of the observed average dissipation profile at station 1, the only station with sufficient XCP and microstructure profiles. We cannot reach a firm conclusion on the skill of the parameterization at high latitudes, because the dissipation rates are close to the instrumental noise level.

Using diapycnal mixing rates representative of the mixing in the Arctic over rough topography, we estimate that internal-wave mixing away from abyssal plains in the Arctic Ocean can account for 6–12 TW of heat loss, downward from the AW layer to the Arctic Intermediate Water below, which will eventually leave the basin in the East Greenland Current. Around the YP, on the other hand, diapycnal mixing alone is unlikely to be significant for regional cooling of the WSC.

Our observations made during neap tides and in calm weather are likely biased and do not resolve the possible temporal variability. A comparison of approximately collocated two stations occupied in neap and spring tides shows no evidence of enhanced mixing in the resolved upper half of the water column. We compare our observations to those made in different seasons as well as at different locations around the YP. We expect significantly different internal-wave environment, because of different stratification and forcing, between summer and winter observations. Winter observations will be influenced by strong convection and storms. The character of ice deformation is also likely to be different in winter and summer conditions. We therefore expect large temporal as well as spatial variability. With this inherent variability in mind, we report strongly variable internal-wave activity and mixing both within the present dataset and when compared to observations from the central (Eckert and Foster 1990; D’Asaro and Morison 1992) and northern (Levine et al. 1985; Hunkins 1986; Padman and Dillon 1991) part of the plateau. We hypothesize (to be addressed in a future paper) that spatially varying tides are largely responsible for this mixing. Holloway and Proshutinsky (2007) have previously shown that tidally induced mixing along the continental boundaries significantly enhances the ventilation of heat from Atlantic Water in the Arctic Ocean and influences the long-term state of the modeled Arctic. Considering that the pathway of the Atlantic inflow over and around the YP is highly variable (Bourke et al. 1988; Padman and Dillon 1991), cooling and freshening of the Atlantic inflow will be affected by the path it follows.

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REFERENCES


