Early Spring Oceanic Heat Fluxes and Mixing Observed from Drift Stations North of Svalbard*

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ABSTRACT

From several drifting ice stations north of Svalbard, Norway, observations were made in early spring of the ocean turbulent characteristics in the upper 150 m using a microstructure profiler and close to the under-ice surface using eddy correlation instrumentation. The dataset is used to obtain average heat fluxes at the ice–water interface, in the mixed layer, across the main pycnocline, as well over different water masses in the region. The results are contrasted with proximity to the branches of the warm and saline Atlantic water current, the West Spitsbergen Current (WSC), which is the main oceanic heat and salinity source both to the region and to the Arctic Ocean. Hydrographic properties show that the surface water mass modification is typically due to atmospheric cooling with relatively less influence of ice melting. Surface heat fluxes of $O(100)$ W m$^{-2}$ are found within the branches of the WSC and over shelf areas with elevated levels of mixing due to strong tides. Away from the shelves and WSC, however, ocean-to-ice turbulent heat fluxes are typical of the central Arctic. Deeper in the water column, entrainment from below together with equally important horizontal advection and diffusion increase the heat content of the mixed layer and contribute to the heat flux maximum in the upper layers. The results in this study emphasize the importance of mixing along the boundaries, over shelves, and topography for the cooling of the Atlantic water layer in the Arctic in general, and for the regional heat budget, hence the ice cover and cooling of the WSC north of Svalbard, in particular.

1. Introduction

At high latitudes, air–sea interaction is affected by the presence of sea ice, which partly insulates the upper ocean from solar radiation and wind stress. A significant source of oceanic heat flux is thus the upward turbulent mixing and entrainment of heat from warmer layers underlying the cold well-mixed layer below the ice (Aagaard et al. 1987). The West Spitsbergen Current (WSC) is the northernmost extension of the Norwegian Atlantic Current and carries warm and saline Atlantic water (AW). North of Svalbard, Norway, where several drift stations reported here are occupied, is the site where the WSC enters the Arctic. In this area, the exchange of heat between the ocean and the ice/atmosphere plays an important role in modifying the core of the AW and thickness of the ice cover (Untersteiner 1988).

Fram Strait, the passage between Greenland and Spitsbergen, Norway, is a site of dramatic exchange of mass, heat, and salt. On the eastern side, AW is carried northward by the WSC and on the western side, the East Greenland Current brings colder and less saline water southward. The WSC transports a major fraction of the heat and salt supplied to the Arctic Ocean (Rudels et al. 2000; Saloranta and Haugan 2001). Primarily confined to the upper continental slope along Svalbard, the WSC bifurcates into two branches around 79°N where it meets the Yermak Plateau (YP; Fig. 1) (Bourke et al. 1988; Manley 1995). The Svalbard branch continues eastward, inshore of the plateau and along the upper part of the slope between the 400- and 500-m isobaths. The Yermak...
branch flows along the offshore western flank of the YP following the 1000-m isobath. While a fraction of the Yermak branch rejoins the Svalbard branch and enters the Arctic Ocean, another fraction detaches north of 80°N and turns westward, contributing to the recirculating Atlantic water (Bourke et al. 1988). Roughly, 20% and 35% of the Atlantic water in the region is carried by the Yermak and Svalbard branches, respectively, whereas about 45% recirculates (Manley 1995). Furthermore, the Yermak branch loses significant heat because of intense tidal mixing at the YP (Padman and Dillon 1991; Padman et al. 1992).

The path of the WSC branches includes areas of intense air-sea ice interaction where the warm core supplies heat to the mixed layer, leading to surface heat fluxes of $O(100)\ W\ m^{-2}$ (Aagaard et al. 1987; Dewey et al. 1999). As the WSC progresses into the Arctic, it loses its heat and surface fluxes are subsequently lower: 0–37 W m$^{-2}$ reported at 32°E (Wettlaufer et al. 1990).

The objective of this study is to quantify the oceanic heat flux toward the ice west and north of Spitsbergen at varying proximity to the warm AW core and over varying topography. During several drifting experiments performed in early spring (within 1 April–1 May) in the area 78.8°–82°N, 0°–15.4°E (Fig. 1), the turbulent heat fluxes were either measured directly in the mixed layer by an eddy correlation method or inferred from dissipation measurements from a microstructure profiler in the upper 150 m. The individual drifts were located within, close to, or away from the branches of the WSC and due to their short durations provide only a snapshot of the early spring conditions in this area. The measurements are from separate years, however at the same time of year, and span the period from 2003 to 2007 (Table 1).

An overview of the measurement locations along with the instruments and the data reduction details is presented in section 2. In section 3, a general description of each drift is given together with salient observations. Oceanic turbulent fluxes are presented and discussed in section 4, followed by a summary and concluding remarks in section 5.
2. Measurements and methods

a. Drift stations and sampling

Geographically, the drift stations are located in the Fram Strait, over the eastern flank of the Yermak Plateau and on the shelf north of Svalbard, all occupied within the period 1 April–1 May of a given year. Upper ocean turbulence measurements were conducted using two separate systems: (i) direct eddy correlation measurements with a turbulence instrument cluster (TIC) and (ii) dissipation measurements in the upper 150 m using a microstructure profiler (MSS). The TIC is equipped with an acoustic 5-MHz Sontek/YSI Acoustic Doppler Velocimeter, an SBE04 standard conductivity sensor, an SBE03 fast-response temperature sensor, and an SBE07 microconductivity sensor, all manufactured by SeaBird Electronics. All TIC sensors sample at the same vertical level, providing three-dimensional velocity, temperature, and salinity at 2 Hz, resolving turbulence well into the inertial subrange (McPhee 2002; Fer et al. 2004; McPhee 2008). MSS, on the other hand, is a loosely tethered free-fall profiler equipped with precision CTD sensors and a suite of turbulence sensors including two shear probes, a fast-response thermistor Fasttip probe (FP07), and a microconductivity sensor (Fer 2006; Fer and Sundfjord 2007). All MSS profiles were collected from the research vessels moored to the ice floe, whereas TIC deployments were made from hydroholes in the ice, more than two ship lengths away from the vessels. The MSS profiler was deployed using a manual winch equipped with 200 m of cable in total, limiting the total profiling depth.

A location map of the drifts, together with a sketch of the WSC, is shown in Fig. 1. In the Fram Strait, two drift stations (FSD) were occupied in deep water in 2005 (FSD1 and FSD2) by R/V Lance. In total, 51 and 26 MSS profiles were collected for about 13- and 10-h durations of FSD1 and FSD2, respectively. The Yermak Plateau Drift (YPD) was conducted in 2003 during the ARK XIX/1 cruise of R/V Polarstern. The sampling included two TIC systems, deployed 4 m apart vertically at 1 and 5 m below the ice over a duration of 6 days. On the shelf north of Svalbard, the Whaler’s Bay Drift (WBD) in 2003 (ARK XIX/1) was on the Svalbard branch of the WSC, whereas the two drifts on the Norwegian Bank (NBD1 in 2005 and NBD2 in 2007) were over shallow and colder water, both occupied by R/V Lance. Sampling consisted of TIC measurements at 1 and 5 m at WBD, at 1 m at NBD2, and MSS profiles at NBD1. A summary of the sampling at each drift is tabulated in Table 1.

Ancillary data include atmospheric measurements and navigation data from the vessels. In 2003 and 2007, TIC measurements were supplemented by shipboard CTD profiles. As a part of field work exercise, University Centre in Svalbard (UNIS) students conducted various sampling during the R/V Lance 2005 cruise. Here we use data from ice cores and ship ADCP, collected at FSD1 and FSD2 (F. Nilsen 2005, personal communication).

b. Data reduction

TIC time series are segmented into 15-min intervals and velocity data are rotated into a streamline coordinate system. The chosen averaging period of 15 min for the turbulence statistics preserves the covariance in the turbulent eddies (typically of several minutes in the under-ice layer) while excluding contamination from variability at longer time scales. To calculate turbulent fluxes, deviatory velocity \( \langle u', v', w' \rangle \), temperature \( T' \), and salinity \( S' \) are obtained by removing the linear trend fitted to the actual time series at each 15-min interval. Fluxes are obtained by zero-lag covariances assuming eddies advected past the...
sensors over the averaging time are representative of the ensemble of instantaneous turbulent fields (Taylor’s frozen turbulence hypothesis). Turbulent heat flux is calculated as \( F_H = \rho C_p (w' T') \), in units of watts per meters squared, where \( \rho \) is the seawater density and \( C_p \) is the specific heat of seawater. In the following, deviatory turbulent quantities are denoted by primes and angle brackets denote averaging in time or space. The Reynolds stress per unit mass is \( \tau = \langle u' v' \rangle + i \langle v' w' \rangle \), expressed in complex notation, where \( i = -1^{1/2} \), and the local friction speed is \( u_f = |\tau|^{1/2} \). The term “local” indicates fluxes obtained at the measurement depth and must be distinguished from the theoretically calculated ice–ocean interface fluxes, which are denoted with subscript 0.

MSS microstructure profile data are processed as described elsewhere (Fer 2006; Sundfjord et al. 2007). Initial 1024-Hz sampling is averaged over four data points (to 256 Hz) to reduce noise. The dissipation rate of turbulent kinetic energy (TKE) per unit mass \( \varepsilon \) in units of watts per kilogram is calculated using the isotropic relation \( \varepsilon = 7.5 \nu \langle u'^2 \rangle \), where \( \nu \) is the viscosity of seawater, \( \langle u'^2 \rangle \) is the shear variance of the horizontal small-scale velocity, and angle brackets denote appropriate averaging. The instrument fall speed (typically \( -0.7 \, \text{m s}^{-1} \)) is used to convert from frequency domain to vertical wavenumber domain using Taylor’s hypothesis, and the shear variance is obtained by iteratively integrating the reliably resolved portion of the shear wavenumber spectrum of half-overlapping 1-s segments. Narrowband noise peaks induced by the probe guard cage are above the wavenumber range chosen for the analysis. Dissipation data in the upper 8 m are unreliable owing to disturbances from the ship and the initial adjustment to free fall. The profiles of precision CTD and \( \varepsilon \) are produced as 10- and 50-cm vertical averages, respectively.

The diapycnal eddy diffusivity of mass is calculated assuming a balance between the production of TKE, its dissipation, and the buoyancy flux as \( K_\rho = \Gamma \varepsilon N^{-2} \) (Osborn 1980), using the flux coefficient (the ratio of work done against gravity to energy dissipated by friction) \( \Gamma = 0.2 \). The choice of \( \Gamma \) is a major uncertainty; however, it follows common practice and allows for direct comparison with other studies. Both observational evidence (Moum 1996) and direct numerical simulation results (Smyth et al. 2001) show significant variability. According to a parameterization that accounts for the evolution over the lifetime of an overturn (Smyth et al. 2001), to employ \( \Gamma = 0.2 \) will underestimate the median \( K_\rho \) by a factor of 2, whereas according to the studies by St. Laurent and Schmitt (1999) and Arneborg (2002) it will overestimate \( K_\rho \) by about a factor of 2 in shear-induced and patchy turbulence. The buoyancy frequency, \( N = \sqrt{-g/\rho (\partial \rho / \partial z)} \), is approximated using Thorpe-ordered \( \sigma_\theta \) profiles (Thorpe 1977), derived from the precision sensors. The density gradient is obtained as the slope of the linear regression of depth on \( \sigma_\theta \), over 4-m-length moving segments (i.e., typically over 40 data points). The background temperature and salinity gradients are derived similarly from the precision temperature and salinity records as the slope of the linear regression of depth on \( T \) and \( S \). As inferred from the Osborn model \( K_\rho \) is ill defined in the absence of stratification. We therefore excluded the segments where the temperature gradient or the density gradient was not significantly different from zero (identified as twice the uncertainty of the slope inferred from the regression described above).

An independent estimate of eddy diffusivity for heat is obtained using \( K_T = 3k_T C_x \) (Osborn and Cox 1972), using the temperature microstructure sampled by the pre-emphasized FP07 signal. Here, \( C_x = \langle (\partial T' / \partial z)^2 \rangle / \langle (\partial T / \partial z)^2 \rangle \) is the Cox number, \( k_T = 1.4 \times 10^{-7} \, \text{m}^2 \text{s}^{-1} \) is the molecular diffusivity for heat, \( \langle (\partial T' / \partial z)^2 \rangle \) is the small-scale temperature gradient variance, and \( \langle (\partial T / \partial z)^2 \rangle \) is the background temperature gradient. We estimate the dissipation rate of thermal variance, \( \chi = 2k_T (3 \langle (\partial T' / \partial z)^2 \rangle) \), by fitting the Batchelor’s form (Dillon and Caldwell 1980) constrained by the measured \( \varepsilon \) to the resolved wavenumber band of the temperature gradient spectrum over 4-m-length segments. The vertical wavenumber spectrum of the temperature gradient \( \partial T / \partial z \) is obtained as \( S_{Tz}(m) = WSTz(f) \), where \( S_{Tz}(f) = \omega^2 S_T(f) \), \( \omega = 2 \pi f \), \( W \) is the fall rate of the profiler, \( m = \beta W \) is the cyclonic wavenumber, and the temperature spectrum \( S_T \) is obtained using the transfer function for the ideal amplification of the circuit. The resolved wavenumber band corresponds to 1–30 Hz [1.4–43 cycles per meter (cpm) using the nominal fall rate of 0.7 m s\(^{-1}\)].

CTD profiles, either from the microstructure profiler with 0.5-m vertical resolution or from the ship CTD system with 1.0-m vertical resolution, are used to determine the mixed layer depth \( (D_m) \) and the pycnocline depth \( (D_p) \). The \( D_m \) is determined using the split-and-merge algorithm of Thomson and Fine (2003), where linear fits are made to individual segments of the density profile and the range of the uppermost, vertical fit defines the extent of the mixed layer; \( D_p \) is determined as the depth of maximum \( N^2 \) using 5-m smoothed buoyancy frequency profiles.

c. Ice drift velocity and under-ice surface friction speed

Ice drift velocity is derived from the 5-min averages of navigation data collected at high temporal sampling (10 s in R/V Lance, 60 s in R/V Polarstern). For each drift station, latitude–longitude pairs were converted to
a polar stereographic grid centered at the mean position of the drift, with a positive $y$ axis aligned with north. The ice drift velocity vector is obtained as the first difference of drift positions.

At the TIC stations, direct measurements of friction velocity $u^*$ in the boundary layer are available at 1 m below the ice. However, $u^*$ can be different from the under-ice surface (ice–ocean interface) friction velocity $u^*_0$, and comparable measurements are not available for MSS drifts. To be consistent between all drifts, we estimate the surface friction velocity $u^*_0$ from a Rossby similarity drag law (McPhee et al. 1999),

$$\frac{U}{u^*_0} = \frac{1}{k} \log \left( \frac{Ro^*}{A + iB} \right),$$

where $U$ is the ice velocity relative to the upper ocean velocity, $k = 0.4$ is the von Kármán constant, $Ro^*_0 = |u^*_0|/fz$, and $A$ and $B$ are similarity constants. For the TIC stations, $U$ is found from the measured velocity at 1 m, whereas for MSS stations FSD1 and FSD2, $U$ is found by subtracting the ADCP velocity at 25 m from the ice drift velocity. At NBD1, no current measurements are available and $u^*_0$ is calculated from ice drift velocity alone. For all stations, the surface roughness is set to $z_0 = 2.0 \times 10^{-3}$ m, which is obtained from the WBD data using the logarithmic law of the wall combined with velocity and friction velocity measurements at 5 m below the ice. The measurements from the deeper TIC are more representative for the overall ice floe and ice in WBD, because the instruments were deployed in an area of relatively smooth first-year ice within the surrounding, more heavily ridged multiyear ice and the measurements at 1 m from the interface may only reflect the local ice topography. The average value of $z_0 = 2.0 \times 10^{-3}$ m is slightly smaller than that of the multiyear floe.
3. General description of drifts

a. Hydrography and water masses

The sampling is limited to the upper 150 m and deep water masses are not observed. We therefore adopt a relevant subset of the water mass categories defined in Aagaard et al. (1985) and summarized in Table 2. Representative temperature and salinity profiles and $T$–$S$ diagrams for each drift are shown in Fig. 2. FSD and NBD1 profiles are time-averaged MSS profiles over the duration of the drift, whereas YPD and WBD profiles are single casts of ship CTD. No prominent cold halocline water is observed, and the mixed layer depth varied by one order of magnitude between a very shallow 6 m at FSD2 and about 70 m at YPD. The warm core of the WSC is composed of warm and saline AW. The Lower Arctic Intermediate Water (LAIW) is of similar salinity but colder, and at this location, can be a result of cooling AW by heat loss to the atmosphere. Arctic Surface Water (ASW), on the other hand, is both colder and fresher and can be a result of cooling AW by a combination of heat loss to the atmosphere and ice melting, which accounts for the freshening (Boyd and D’Asaro 1994).

On average, profiles from FSD1 and FSD2 show a very small fraction of LAIW (owing to the shallow sampling) and we include LAIW in the definition of AW when averaging properties over water masses (section 4d).

b. Fram Strait Drifts

Fram Strait Drift 1 was established west of the WSC on the morning of 28 April 2005 (day 118). Ice thickness during the Surface Heat Budget of the Arctic Ocean (SHEBA) campaign (McPhee 2002).
was about 120 cm over a water depth of about 2500 m near 79°N, close to the prime meridian. The conductive heat flux in the ice is calculated as \( F_{Hi} = -k_i(\partial T/\partial z) \), where \( k_i \) is the thermal conductivity of sea ice and \( (\partial T/\partial z) \) is the temperature gradient in the ice. Temperature gradient measured from an ice core yielded \( F_{Hi} = 10 \text{ W m}^{-2} \) for FSD1. Microstructure profiling started at 1254 UTC, and 54 casts were made at a typical sampling interval of 10–15 min.

The patch of open water on the side of the vessel where the profiling was made froze throughout the drift. Occasionally, profiling was interrupted because of ridging and rafting, and after about 14 h the sampling was terminated as the slack tether was unable to penetrate the young thin ice, making it impossible to conduct free-fall profiling. A malfunction of the automated logging of the ship’s meteorological and navigation data went unnoticed throughout the drift; however, positions at the start of each cast and occasional descriptive meteorological information were hand-logged by the authors. Air temperature was around \(-13^\circ\text{C}\) throughout the drift. The first couple of hours into the profiling, the wind was calm with a speed of 2 m s\(^{-1}\), then decreased practically to nil, and followed with a light breeze of about 3 m s\(^{-1}\) from day 118.0. Despite the calm wind, the station drifted gradually toward the northwest at an average speed of 0.24 m s\(^{-1}\) over warmer water (Fig. 3a).

Mean profiles of hydrography, dissipation rate of TKE, and diapycnal eddy diffusivity \( K_p \) averaged over all casts are shown in Fig. 4. On average, the mixed layer was 21 m deep, with temperature 0.19 K above the freezing point. Dissipation rate, in excess of \( 3 \times 10^{-7} \text{ W kg}^{-1} \) in the mixed layer, reduced to the noise level (\( 2 \times 10^{-8} \text{ W kg}^{-1} \)) below about 60-m depth. Corresponding eddy diffusivity \( K_p \) inferred from the Osborn model decreases from \( O(10^{-2}) \text{ m}^2 \text{ s}^{-1} \) at the base of the mixed layer to \( O(10^{-4}) \text{ m}^2 \text{ s}^{-1} \) below the pycnocline. The time-averaged velocity profile derived from the ship ADCP showed that the speed was about 15 cm s\(^{-1}\) down to 40 m, gradually decreased to about 8 cm s\(^{-1}\), and stayed nearly constant below 70 m (Fig. 5a). The angular shear (change of current direction with depth) was significant in the upper 70 m, and despite the strong stratification, the 4-m gradient Richardson number \( [\text{Ri} = N^2/\text{Sh}^2] \) was often below unity, suggesting that the enhanced \( e \) between 40 and 70 m (Fig. 4c) can be due to local shear instabilities.

Fram Strait Drift 2 was established on the morning of 30 April 2005 (day 120) on a relatively thin floe of about 30 cm. In contrast to the freezing conditions at FSD1, repeated ice core measurements and mixed layer temperatures indicated that the ice at FSD2 was melting rapidly. The melting rate could not be determined with confidence because of the thin ice floe and scarcity of ice...
Conductive heat flux, averaged over two ice cores, was 15 W m$^{-2}$. Air temperature gradually warmed from −12° to −3°C throughout the drift when wind averaged 6.4 m s$^{-1}$ (Figs. 6a,b). The mixed layer was shallow and the upper layer temperatures gradually increased. The time-averaged velocity profile derived from ship ADCP was nearly unidirectional at 240° in the upper 25–170 m with the speed between 35 and 40 cm s$^{-1}$. Dissipation rate and eddy diffusivity were significantly larger when compared to FSD1, however they similarly reached the noise level below about 60 m (Fig. 7). Compared to FSD1, the stratification was weaker; nevertheless, owing to the lack of shear, the 4-m Ri was one order of magnitude larger (Fig. 5). Relatively enhanced mean dissipation at around 50 m was largely a result of the event that occurred toward the end of the drift, following the pinching of the isotherms and isohalines (Figs. 6c,d).

Using a free-drift model (McPhee 1986; see also Boyd and D’Asaro 1994) and representative values of the air–ice ($C_a$) and ice–water ($C_w$) drag coefficients, relative ice speed will be about 2.5% of the wind speed, $U_r = U_u \rho_a C_a / (\rho_u C_w)$, where $U_u$ is the wind velocity and $\rho_a$ and $\rho_w$ are the air and water densities, respectively. The actual ice motion can be obtained by adding the average upper ocean velocity. In FSD2, the residual between average ice velocity inferred from the navigation (absolute ice velocity) less the 2.5% of the mean wind speed yields a 0.35 m s$^{-1}$ average upper ocean velocity, which is consistent with ship ADCP measurements. Also, applying the same model and equating the ice–water stress and the air–ice stress (calculated from absolute ice and wind velocities) yields a friction velocity at the ice–ocean interface that corresponds well in magnitude with that derived from the Rossby similarity drag law [Eq. (1)].

c. Yermak Plateau Drift

The Yermak Plateau Drift, established on day 100 in 2003, had the largest duration with 6 days of sampling. Camp was established on a large floe of 2–3-m thick, multiyear ice; however, TIC instrumentation was deployed through thinner ice (∼0.5 m) at the edge of the larger floe. The upper 70 m was well mixed in temperature and salinity (Fig. 2, green curve). The mixed layer temperature was the coldest and the temperature maximum below the thermocline was the lowest of all stations. The floe drifted at a slow pace of 7 cm s$^{-1}$, comparable to NBD1 but about $\frac{1}{2}$–$\frac{1}{3}$ the speed of FSD1 and NBD2. The location is slightly southeast of the site where upward
turbulent heat fluxes reaching 30 W m\(^{-2}\), due to energetic mixing events as a result of proximity to internal wave sources, were reported by Padman and Dillon (1991). It is also the same area where an automated buoy measured surface heat fluxes of 22 W m\(^{-2}\) over the elevated topography of Yermak Plateau, 50 days prior to our deployment (McPhee et al. 2003).

d. Drifts on the North Spitsbergen shelf

Whaler’s Bay Drift, north of Svalbard, was located where the continental slope branch of warm and relatively salty AW transported by the West Spitsbergen Current enters the Arctic. WBD was established on 1 April 2003 on an area of 1.1-m-thick first-year ice within about twice as thick and more heavily ridged multiyear ice. An ice core was extracted and ice temperatures measured at 10-cm vertical resolution yielded a temperature gradient corresponding to an upward conductive heat flux of about 41 W m\(^{-2}\). The relatively high temperature gradient reflects the low air temperature, which averaged to -29.1°C over the drift, while the average wind was northeasterly at 9 m s\(^{-1}\). Water depth was about 180 m with a mixed layer depth of 19 m with temperature 0.93 K above freezing.

Norwegian Bank Drift 1, established 1 May 2005, was closer to land and drifted across a thermohaline front that separated the well-mixed shallows of polar water (PW) characteristics from the deep water of Arctic Surface Water properties. Air temperature during NBD1 was relatively warm, varying between seawater temperatures and zero. The drift started over about 90-m-deep water and crossed the thermohaline front halfway into the drift.

![Figure 6](image-url)
The shallow side of the front was characterized by unusually high dissipation rates (Fig. 8), enhanced both in the upper and the bottom layers, suggesting strong tidal currents (section 3e) generating turbulence in the boundary layer near the seafloor and under the ice. The $T$–$S$ profiles and stratification averaged on either side of the front are contrasted in Fig. 9. Shallows are nearly homogeneous and well mixed. The dissipation rate is at least one order of magnitude larger on the shallow side of the front (Fig. 9c) and decreases with distance from the boundaries toward the midwater. Throughout NBD1, measured dissipation rates are well above the noise level.

Norwegian Bank Drift 2 was established on 24 April 2007 in the same area as NBD1. Compared to NBD1, the NBD2 drift was 5 times as fast and over a mixed layer that was colder and twice as deep (Table 1). The hydrography from the CTD casts shows a tongue of AW below about 60 m and a relatively strong salinity gradient at the front (Fig. 10). Average air temperature was $-13^\circ$C and an average northwesterly wind of 9 m s$^{-1}$ accompanied the drift across the shallow topography of the Norwegian Bank.

e. Tides

Tidal currents can be strong and influence the oceanic turbulent fluxes at the Yermak Plateau (Padman and Dillon 1991; Padman et al. 1992) and over the shelves north of Svalbard (Sundfjord et al. 2007). We estimate and contrast the tidal current ellipse parameters for each drift using the 5-km-resolution Arctic Ocean Tidal Inverse Model (AOTIM) (Padman and Erofeeva 2004). Results predicted for the start location of each drift for the two main constituents $M_2$ and $K_1$ are summarized in Table 3. Because NBD1 extends into significantly shallow water, ellipse parameters for the end location of NBD1 are also calculated. Generally, model water depths interpolated to the drift positions agree with the actual water depth. At the shallow end of NBD1, however, the model depth is 58 m, about double the actual depth, and the modeled tidal currents shown in Table 3, which scale inversely by the water depth, may therefore underestimate the actual current by a factor of 2. Except for FSD2, the tidal ellipses have clockwise rotation at all drifts for both the diurnal and semidiurnal constituents (negative semiminor axis length). Typically, the semidiurnal tide dominates, except from at YPD where the dominant constituent is $K_1$ and at FSD2 where both constituents are of comparable magnitude. Tidal currents at WBD are significant whereas those toward the end of NBD1 are the largest and will lead to significant stirring both in the under-ice boundary layer and the bottom boundary layer near the seabed.

4. Turbulent fluxes

a. Heat flux at the under-ice surface

The under-ice surface (ice–seawater interface) heat flux can be estimated using the parameterization

$$F_{h0} = \rho C_p c_h |\mathbf{u}_n| \Delta T,$$

Fig. 7. Same as in Fig. 4, but for FSD2 and averaged over 26 casts.
where $c_H$ is the turbulent exchange coefficient (Stanton number), $u_{*0}$ is the surface friction velocity, and $\Delta T$ is the mixed layer temperature above freezing. The exchange coefficient has been shown to be nearly constant, $c_H = 0.0057$, for the under-ice boundary layer (McPhee et al. 2003). Although the local turbulent heat flux in the under-ice boundary layer can be directly calculated from the TIC data at 1 m below the ice (section 2b), the actual surface flux can be different. The surface friction velocity can be estimated using the Rossby similarity [Eq. (1)] or it can be approximated using the local $u_*$ measured close to the ice undersurface (typically 1 m) when available. To be consistent, we calculated $u_{*0}$ via the Rossby similarity for all drifts. When compared to 1-m TIC measurements, $|u_{*0}|$ is 0.99, 0.30, and 1.11 times the local $|u_*|$ for WBD, YPD, and NBD2, respectively. The under-ice surface heat flux results are compared for all drifts in Table 4. Also tabulated are the measured heat fluxes at 1 m below the ice for the TIC drifts and mixed layer averages for the MSS drifts. The largest interfacial heat flux occurred at FSD2, consistent with the observations of rapid ice melting throughout the drift. Significantly large upward $F_{H0}$ was also observed at WBD on the Svalbard branch of the WSC. Although the drift speed during NBD2 was 5 times that during NBD1, $u_*$ was only 1.4 times as large, and the mixed layer temperature above freezing in NBD2 was 0.4 times that during NBD1. Consequently, according to
Eq. (2), the interfacial heat flux in NBD2 is only 60% of that in NDB1 (Table 4). For NBD1, $F_{H0}$ is calculated assuming stagnant upper ocean, which, according to the tidal estimates in section 3e, is probably not the case.

For the drifts where the conductive heat flux was calculated using the measured gradient in the ice (FSD1, FSD2, and WBD), an estimate of melting rate is made by assuming that any discrepancy between ocean heat flux and conductive heat flux at the ice–ocean interface is balanced by melting or freezing (e.g., Steele and Morison 1993). From this method, melting rates of 0.3, 9, and 5 cm day$^{-1}$ are estimated for FSD1, FSD2, and WBD, respectively. Although freezing was observed in open-water areas during FSD1, the ice in general was slowly melting as the surface heat flux was larger than the conductive heat flux in the ice. Accordingly, the salt fluxes in the mixed layer were positive (but small), which corresponds to an input of meltwater at the surface (Table 5). For FSD2, the inferred melting rate of 9 cm day$^{-1}$ is comparable to the observed rapid melting. The upward (positive) salt fluxes in the mixed layer were significantly larger than during FSD1, which qualitatively supports significant melting. For WBD, the melting rates are in line with the observed salt fluxes reported in Sirevaag (2009).

**b. Heat and salt flux in the mixed layer**

The local turbulent heat flux in the under-ice boundary layer is directly calculated from the TIC data as $F_H = \rho C_p (w^T)$ (section 2b). At TIC stations, values from the cluster at 1 m are averaged for WBD and NBD2, and those from both levels (1 and 5 m) are averaged for YPD to obtain representative heat flux values in the mixed layer. Although WBD had another cluster at 5 m below the ice, the mean heat flux was more than double that measured at 1 m (620 and 259 W m$^{-2}$, respectively), with significantly large differences after day 92.1 (Fig. 12d). Large values at 5 m at WBD are likely the consequence of pressure ridge keels and might be representative of variability in the mixed layer for irregular under-ice topography. For MSS stations, turbulent heat and salt fluxes ($=10^{-3}\rho \times$ salinity flux) are obtained using the eddy diffusivity for heat from the Osborn–Cox model and the property gradients (section 2b), both evaluated at 4-m vertical segments:

$$F_H = -\rho C_p K_T \left( \frac{dT}{dz} \right) \quad (W \text{ m}^{-2})$$

$$F_S = -\frac{\rho}{1000} K_S \left( \frac{dS}{dz} \right) \quad (kg \text{ s}^{-1} \text{ m}^{-2}).$$

![Figure 9. Average profiles for NBD1. Thin traces are the profiles averaged on the deep side of the thermohaline front (casts 1–14) whereas the thick traces are on the shallow (well mixed) side of the front (casts 19–29). (a) Temperature (black) and salinity (gray), (b) potential density anomaly $\sigma_0$ (black) and buoyancy frequency $N$ (gray), (c) 2-m bin-averaged dissipation of TKE $\epsilon$, and (d) eddy diffusivity using $0.2e^{-N^2}$, ignoring segments with $N < 1.5$ cph.](image-url)
In the calculations we assume turbulent flow leading to eddy diffusivities \( K_T \approx K_S \).

Mixed layer fluxes are obtained by averaging the values for \( z, D_{ml} \) for each cast. MSS-derived eddy diffusivity is available below 8-m depth [i.e., center of the first 4-m-long segment used in Eq. (3) is at 10 m], hence we only retain values at casts where \( D_{ml} > 10 \) m. TIC measurements, on the other hand, are either within the upper 5 m or 1 m below the ice. Therefore, the values inferred from MSS (TIC) are biased toward the lower (upper) part of the mixed layer.

A summary of the mixed layer turbulent heat flux as measured by the TIC is given in Figs. 11–13 for YPD, WBD, and NBD2, respectively. At YPD, the mixed layer temperature had a pronounced diurnal signal (Fig. 11a), consistent with the strong diurnal tides from AOTIM5 (Table 3) and previous observations of strong diurnal tides in the region (Padman et al. 1992). Friction velocity decreased from 1 to 5 m below the ice, suggesting a mechanical source for TKE generation at the interface. The average mixed layer heat flux of 2 W m\(^{-2}\) was comparable to the typical Arctic values and compared favorably with the heat flux at the ice–water interface of \( F_{H0} = 3 \) W m\(^{-2}\) obtained from the parameterization in Eq. (2). At the WBD, the variation in the measured \( F_H \) was large, both in time and between the two measurement levels. The average value at 1 m of 259 W m\(^{-2}\) was larger than, but comparable to, the interface value of 203 W m\(^{-2}\) (Table 4). The friction velocity was slightly larger at 5 m, particularly in the late part of the drift, suggesting TKE contribution from distant sources. At NBD2, TIC measured through a gradually less saline mixed layer and the front seen between the first two CTD stations in Fig. 10 was absent in the TIC data (Fig. 13a). Heat flux in the mixed layer increased from ~30 to over 80 W m\(^{-2}\) as the floe drifted over warmer water, as much as 0.45 K above freezing.

The mixed layer was only occasionally deeper than 10 m at FSD2 and NBD1, hence flux estimates in the mixed layer using Eq. (3) with profiler data from these sites are scarce and may not be representative of the drift average. The flux estimate from FSD1, on the other hand, is well resolved. Despite a mean upward heat flux of 28 W m\(^{-2}\), the open water at the surface was freezing as a result of low air temperatures increasing the ocean/air temperature difference and leading to heat loss to the atmosphere that is greater than the oceanic heat flux.

Larger fluxes of heat at the surface compared to fluxes across the pycnocline should reduce the total amount of heat in the mixed layer. Heat content in the mixed layer \( H_{ml} \) is calculated from the temperature above freezing and the mixed layer depth for each MSS profile, and the rate of change of the heat content \( dH_{ml}/dt \) is found as the slope of a line fitted to the evolution of \( H_{ml} \) in time. For FSD1 and FSD2, heat content in the mixed layer is reduced at rates equivalent to heat fluxes of \(-387 \) and \(-496 \) W m\(^{-2}\), respectively, where the negative fluxes mean that heat is lost from the mixed layer. As in Steele

### Table 3. Tidal current ellipse parameters derived at the drift start positions using the AOTIM5 model (Padman and Erofeeva 2004). Those at the final position of the NBD1 are also shown.

<table>
<thead>
<tr>
<th></th>
<th>WBD</th>
<th>YPD</th>
<th>FSD1</th>
<th>FSD2</th>
<th>NBD1 start</th>
<th>NBD1 end</th>
</tr>
</thead>
<tbody>
<tr>
<td>( M_2 )</td>
<td>Ma/mi (cm s(^{-1}))( ^\ast )</td>
<td>8.7/(-2.1)</td>
<td>3.9/(-0.7)</td>
<td>3.2/(-0.4)</td>
<td>4.3/0.4</td>
<td>8.4/(-0.1)</td>
</tr>
<tr>
<td></td>
<td>Pha/inc ((^\circ))( ^\ast)</td>
<td>272/53</td>
<td>249/51</td>
<td>53/85</td>
<td>60/87</td>
<td>30/45</td>
</tr>
<tr>
<td>( K_1 )</td>
<td>Ma/mi (cm s(^{-1}))( ^\ast )</td>
<td>4.8/(-1.4)</td>
<td>8.2/(-5.1)</td>
<td>2.0/(-0.1)</td>
<td>4.1/1.6</td>
<td>2.4/(-0.4)</td>
</tr>
<tr>
<td></td>
<td>Pha/inc ((^\circ))( ^\ast)</td>
<td>48/51</td>
<td>36/56</td>
<td>47/90</td>
<td>37/77</td>
<td>7/23</td>
</tr>
</tbody>
</table>

\( ^\ast \) Ma/mi are the semimajor/semiminor axis lengths.

\( ^\ast\) Pha/inc are the phase and the inclination.
and Morison (1993), we can estimate the contribution to changes in mixed layer heat content from horizontal advection and diffusion relative to the drift as the residual \( A = dH_{\text{ml}}/dt - F_{\text{H}} - F_{\text{HPC}} \), where \( F_{\text{HPC}} \) is the heat flux at the base of the mixed layer. Average values of \( A \) for FSD1 and FSD2 are \(-374\) and \(-282\) W m\(^{-2}\), respectively. Values for FSD1 are biased as the ice drifted past a weak thermohaline front in the mixed layer and into gradually colder (and fresher) water until day \(-118.7\). The contribution from horizontal advection and diffusion was below \(-2\) kW m\(^{-2}\) in this period, whereas \( A = 27\) W m\(^{-2}\) for the period after day 118.7. During the first period, the horizontal advection and diffusion is far more important than surface fluxes in determining the mixed layer heat content. Significant contribution from horizontal advection and diffusion was also inferred northeast of Svalbard by Steele and Morison (1993). For the latter period, \( A \) is of the same order as the surface fluxes. In the case of FSD2, both the surface fluxes and the advective fluxes are one order of magnitude larger than the last period of FSD1, due to the proximity of the WSC branch. However, both the advective effects and vertical fluxes are of the same magnitude and seem to be equally important in modifying the mixed layer heat content, which decreased steadily throughout the drift.

c. Fluxes across the pycnocline

Time-averaged profiles of turbulent heat and salt fluxes are shown in Fig. 14 for the profiling depth at each drift. In this section, fluxes averaged between the base of the mixed layer and the stratification maximum \((D_{\text{ml}} < z < D_{\text{p}})\) are presented. Average turbulent parameters \(\epsilon\), \(K_p\), and \(K_T\), the buoyancy frequency, and the resulting fluxes are summarized in Table 5. The stratification varies between 3 and 5 eph, with the weakest at NBD1, under the influence of strong mixing with exceptionally large eddy diffusivities.

The variability in the dissipation rate and the eddy diffusivity at FSD1 is not significant, leading to a narrow 95% confidence interval on the parameters; however, the standard deviation on resulting fluxes is comparable to their mean, owing to the variability in the 4-m property gradients. The mean values of \(\epsilon\) and \(K_T\) for FSD1, which was located a considerable distance away from the core of the WSC, are one order of magnitude larger than typical low-latitude open-ocean pycnocline observations (Gregg 1987). This increases by one order of magnitude at FSD2. The dissipation rate is 3–5 times larger at FSD2. The average current profile at FSD2 lacks mean shear (Fig. 5b). The pycnocline is shallower and the boundary mixing penetrates to the pycnocline, leading to entrainment of relative warm deeper water, which also explains the slowly decreasing heat content of the mixed layer. In contrast to FSD2, several layers of stratification maxima are observed at FSD1 (cf. Figs. 4, 6). In these layers where densely spaced isotherms are separated by relatively mixed layers, the dissipation rate has to work against the finescale stratification.

d. Fluxes in water masses

Mean profiles of heat and salt fluxes at FSD1, FSD2, and NBD1 (Fig. 14) show large vertical variability with a layering structure. In all drifts the fluxes decrease with distance from the surface in the upper \(-20\) m. On the shallow side of the front at NBD1, the heat flux increases toward the bottom whereas the salt flux remains small in magnitude but changes sign. Here, horizontal (possibly isopycnal) exchange with significantly warmer water of comparable salinity from the deeper side of the front can

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**Table 4. Comparison of the average mixed layer oceanic conditions and heat flux for all drifts.**

<table>
<thead>
<tr>
<th>Layer</th>
<th>WBD</th>
<th>YPD</th>
<th>FSD1</th>
<th>FSD2</th>
<th>NBD1</th>
<th>NBD2</th>
</tr>
</thead>
<tbody>
<tr>
<td>(T_{\text{ml}}) (°C)</td>
<td>-0.56</td>
<td>-1.81</td>
<td>-1.68</td>
<td>-0.30</td>
<td>-1.03</td>
<td>-1.49</td>
</tr>
<tr>
<td>(S_{\text{ml}})</td>
<td>34.53</td>
<td>34.16</td>
<td>34.19</td>
<td>34.54</td>
<td>34.49</td>
<td>34.37</td>
</tr>
<tr>
<td>(\Delta T) (K)</td>
<td>0.93</td>
<td>0.05</td>
<td>0.19</td>
<td>1.58</td>
<td>0.86</td>
<td>0.37</td>
</tr>
<tr>
<td>(F_H) (W m(^{-2}))</td>
<td>259</td>
<td>2</td>
<td>28</td>
<td>206(^*)</td>
<td>240(^*)</td>
<td>69</td>
</tr>
<tr>
<td>(</td>
<td>u_{\text{H}}</td>
<td>\times 10^{-2}) (m s(^{-1}))</td>
<td>0.94</td>
<td>0.23</td>
<td>0.44</td>
<td>0.99</td>
</tr>
<tr>
<td>(F_{\text{T}}) (W m(^{-2}))</td>
<td>203</td>
<td>3</td>
<td>21</td>
<td>303</td>
<td>116</td>
<td>69</td>
</tr>
</tbody>
</table>

\(^*\) Few data points.

**Table 5. Average turbulent parameters in the mixed layer and in the layer between the base of the mixed layer and pycnocline. All values are inferred from microstructure profiling. The buoyancy frequency, heat flux, and salt flux are given as mean values ± one standard deviation. Values of the dissipation rate and the eddy diffusivities are the maximum likelihood estimator and the 95% confidence limits are given in brackets.**

<table>
<thead>
<tr>
<th>Layer</th>
<th>FSD1</th>
<th>FSD2</th>
<th>NBD1</th>
<th>FSD1</th>
<th>FSD2</th>
<th>NBD1</th>
</tr>
</thead>
<tbody>
<tr>
<td>(N) (cph)</td>
<td>1.6 ± 1.0</td>
<td>3.6 ± 1.3</td>
<td>0.7 ± 1.0</td>
<td>4.8 ± 0.6</td>
<td>4.4 ± 1.4</td>
<td>3.0 ± 1.0</td>
</tr>
<tr>
<td>(\epsilon) \times 10(^{-3}) (W kg(^{-1}))</td>
<td>23.6 [16.54]</td>
<td>598 [18.19713]</td>
<td>353</td>
<td>5.3 [4.8 5.9]</td>
<td>23.5 [15.1 36.5]</td>
<td>917.3 [290 2900]</td>
</tr>
<tr>
<td>(K_T) \times 10(^{-4}) (m(^2) s(^{-1}))</td>
<td>80 [37 175]</td>
<td>61 [22 169]</td>
<td>276</td>
<td>8 ± 8</td>
<td>89 ± 103</td>
<td>301 ± 267</td>
</tr>
<tr>
<td>(F_H) (W m(^{-2}))</td>
<td>28 ± 44</td>
<td>206 ± 29</td>
<td>240</td>
<td>8 ± 8</td>
<td>89 ± 103</td>
<td>301 ± 267</td>
</tr>
<tr>
<td>(F_S) \times 10(^{-3}) (kg s(^{-1}) m(^{-2}))</td>
<td>1.86 ± 5.4</td>
<td>15.9 ± 8.2</td>
<td>30.9</td>
<td>0.44 ± 0.4</td>
<td>4.2 ± 5.8</td>
<td>33.1 ± 36.1</td>
</tr>
</tbody>
</table>
lead to the observed flux profiles. To assess the turbulent exchange between different water masses, we average the fluxes over the extent of each water mass. All 4-m segments from the MSS casts with heat and salt fluxes inferred from the Osborn–Cox model (section 2b) are used to identify the water masses (Table 2) using the segment mean temperature and salinity. Average values of stratification, dissipation rate, inferred diffusivity, and turbulent fluxes are given for the MSS drifts (Table 6). Figure 15 shows the average heat fluxes in different water masses for the duration of the drifts, also including those in the mixed layer for completeness. The uppermost water mass [PW and Polar Intermediate Water (PIW)] typically extends below the mixed layer. FSD1 mixed layer heat fluxes compare well with the Stanton number–based estimates for the surface fluxes, except for day 118.63 when downward fluxes were observed when crossing a thermal front in the mixed layer and toward the end of the drift when significant convection led to enhanced upward fluxes. In FSD2, there are only four casts that resolve the mixed layer, hence the heat flux estimates here are uncertain and are slightly smaller than the Stanton number–based estimate. An increase in heat flux toward the end of the drift is observed in all water masses, corresponding to the mixing event at around day 120.81. During the NBD1, heat fluxes within the PW, PIW, and ASW water masses increase toward the end of the drift as the floe approaches shallower topography. Generally for all drifts below the mixed layer, layer-averaged heat fluxes decrease with depth as a consequence of reduced mixing and thermal diffusivity.

A summary of drift average heat fluxes and salt fluxes within the three classes of water masses is given in Fig. 16. All three drifts show a decreasing magnitude of heat and salt fluxes with increasing depth, somewhat counterintuitive to the general understanding that the main source of heat and salt is the layers of Atlantic water. This is mainly because the AW (and LAIW) has relatively weak vertical temperature gradient when compared to ASW, although the eddy diffusivity for heat in these water masses is comparable (even larger $K_T$ in AW at FSD1). Furthermore, the average value in the AW is reduced because the temperature gradient reverses below the depth of the AW temperature maximum, and thus the direction of the heat flux is opposite in the upper and lower parts of this water mass, reducing the average (see, e.g., Fig. 14a). Heat is likely entrained from the deeper AW into PW, as observed at FSD2; however, advective heat flux is also significant (section 4b). The vertical variability of the heat fluxes within the AW layer is significant at FSD2 (Fig. 14a). At the upper side of the AW at about 40 m where there is a subsurface
temperature maximum, the heat flux is in excess of 100 W m\(^{-2}\), an order of magnitude larger than the layer-averaged value. The nonzero convergence of the vertical heat flux above the AW will heat up the overlying layer. Water mass transformations in this region can be affected by ice melting. Following Cokelet et al. (2008), lines with a constant ratio of atmospheric cooling to ice–ocean interface melting of ice with a salinity of 5 psu are...

**FIG. 12.** Same as in Fig. 11, but for WBD on the northern shelf of Svalbard. Because the time series is relatively short, all 15-min segments are shown using 1-h running mean.

**FIG. 13.** TIC measurements at 1 m below the ice during NBD2 in 2007. All 15-min segments are shown using 1-h running mean. (a) Salinity (solid) and temperature (dashed), (b) elevation above freezing (shaded area, left axis) and the local friction velocity (solid), and (c) the heat flux.
plotted as yellow dashed lines in the $T$–$S$ diagram in Fig. 2. The ratio indicates that a water mass with initial properties $T = 3.2^\circ C$, $S = 35.1$ is modified by melting only ($R = 0$), by equal contributions of atmospheric cooling and interface melting ($R = 1$), or more by atmospheric cooling ($R = 2$). The limit $R \to \infty$ (a vertical line on $T$–$S$ space) indicates modification by atmospheric cooling alone. Generally, the upper water mass characteristics are not produced along the melting line, but both the atmospheric cooling and ice melting are important (approximately equally). This is a consequence of the proximity to the marginal ice zone and the variable ice concentration. From Fig. 2, the water column at FSD1 has a gentler slope than FSD2, WBD, and NBD1. A slope with $R < 1$ indicates that melting has played a more dominant role in modifying the water column. FSD1 was situated westward of the other drifts, where it is more likely to encounter water exported from the Arctic through the Fram Strait, which has been modified by melting under continuous ice cover. The FSD2, WBD, and NBD1 water columns in the $T$–$S$ diagram are along lines that indicate that atmospheric cooling has made a larger contribution to modification than melting, which is appropriate since water masses are advected along the WSC and drifts are within the marginal ice zone where air–sea–ice interaction is intense.

5. Summary and concluding remarks

In total, six drift stations north of Svalbard were occupied in early spring covering years 2003–07. Turbulent characteristics were observed using a microstructure profiler or eddy correlation instrumentation. The drift stations represent three different conditions: 1) within close proximity to the West Spitsbergen Current (WSC) with shallow and warm mixed layers and subsequent large heat fluxes [Whaler’s Bay Drift (WBD) and Fram Strait Drift 2 (FSD2)]; 2) away from the WSC where the mixed layer is deeper, colder, and displays relatively moderate vertical heat fluxes [Yermak Plateau Drift (YPD) and Fram Strait Drift 1 (FSD1)]; and 3) over the continental shelf in the Svalbard branch where cooling and mixing of the WSC is enhanced by the ice drift and tidal effects related to topography [Norwegian Bank Drifts (NBD1 and NBD2)].
Table 6. Water mass–averaged properties. Values are the drift averages for \( T \) and \( S \), average + one standard deviation for depth \( z \), thickness \( h \), buoyancy frequency \( N \), and the turbulent surface heat fluxes \( F_{\text{sh}} \) and \( F_{\text{sd}} \), and the maximum likelihood estimator with the 95% confidence interval in brackets for the dissipation rate and the eddy diffusivities.

<table>
<thead>
<tr>
<th>Layer</th>
<th>PW and NBDI</th>
<th>PW and ASW</th>
<th>PW and LAIW</th>
<th>PW and ASW</th>
<th>PW and LAIW</th>
</tr>
</thead>
<tbody>
<tr>
<td>FSD1</td>
<td>( 11.9 )</td>
<td>3.49</td>
<td>( 53.93 )</td>
<td>( 10.4 )</td>
<td>( 2.2 )</td>
</tr>
<tr>
<td></td>
<td>( 10.4 )</td>
<td>( 1.22 )</td>
<td>( 76.1 )</td>
<td>( 10.6 )</td>
<td>( 1.98 )</td>
</tr>
<tr>
<td></td>
<td>( 10.6 )</td>
<td>( 2.14 )</td>
<td>( 48.7 )</td>
<td>( 2.14 )</td>
<td>( 1.98 )</td>
</tr>
<tr>
<td></td>
<td>( 2.14 )</td>
<td>( 0.23 )</td>
<td>( 119 )</td>
<td>( 0.23 )</td>
<td>( 119 )</td>
</tr>
<tr>
<td></td>
<td>( 0.23 )</td>
<td>( 366 )</td>
<td>( 7.9 )</td>
<td>( 0.23 )</td>
<td>( 366 )</td>
</tr>
<tr>
<td></td>
<td>( 366 )</td>
<td>( 17 )</td>
<td>( 1.2 )</td>
<td>( 366 )</td>
<td>( 17 )</td>
</tr>
<tr>
<td></td>
<td>( 17 )</td>
<td>( 1.8 )</td>
<td>( 100 )</td>
<td>( 1.8 )</td>
<td>( 100 )</td>
</tr>
<tr>
<td>Surface heat fluxes ( F_{\text{sh}} ) (W m(^{-2}))</td>
<td>( 10.9 \times 10^{-3} )</td>
<td>( 1.2 \times 10^{-5} )</td>
<td>( 0.77 \times 10^{-3} )</td>
<td>( 0.77 \times 10^{-3} )</td>
<td>( 0.77 \times 10^{-3} )</td>
</tr>
</tbody>
</table>

Measurements from drifts within or close to the WSC show surface and mixed layer heat fluxes of 200–300 W m\(^{-2}\) and estimated melting rates of 5–9 cm day\(^{-1}\). Heat fluxes are comparable to and provide support for earlier studies that inferred vertical heat fluxes from heat loss observed between subsequent cross sections of the WSC (e.g., Aagaard et al. 1987; Saloranta and Haugan 2004; Cokelet et al. 2008). During FSD2, turbulence in the upper layers is one order of magnitude larger than that observed away from the WSC, and extends deep into the pycnocline, entraining warmer water into the mixed layer. Changes of the mixed layer heat content, however, indicate that the contribution from horizontal advection and diffusion is equally important in the heat balance within the mixed layer.

Outside the WSC, turbulent fluxes in the mixed layer are small. Over the northeastern flank of Yermak Plateau where the mixed layer is near freezing point and ice drift is slow, the average heat flux is 2 W m\(^{-2}\) at YPD, but varies significantly due to solar radiation and tides. At FSD1, vertical heat fluxes and contributions from horizontal advection and diffusion account equally for changes in mixed layer heat content. In contrast to FSD2, mixing at FSD1 is typically confined to the mixed layer with localized enhanced mixing between 40 and 70 m due to strong velocity shear.

On the shelf area, observed elevated levels of mixing are due to tidal and topographic effects and lead to surface heat fluxes of \( O(100) \) W m\(^{-2}\). As the shallows are approached, an increasing fraction of the water column is covered by bottom and upper boundary layers and mixing by tides and topography gets increasingly important.

Profiles of heat flux typically decrease with depth when averaged in water mass classes. Average fluxes in the Atlantic water (AW) and the Lower Arctic Intermediate Water layers are considerably lower than the surface heat fluxes (e.g., only 10% of the surface flux at FSD1). This also shows that a considerable fraction of the mixed layer heat is entrained from deeper layers or advected into the area. Near the WSC the vertical variability is large and heat flux on the crucial upper side of the AW exceeds 100 W m\(^{-2}\), one order of magnitude larger than the layer-averaged value.

The slope of temperature against salinity properties compared with that expected from a given ratio of melting/atmospheric cooling shows that upper-layer water masses in all our drifts north of Svalbard are slightly more affected by atmospheric cooling than by melting of ice. This is in line with Cokelet et al. (2008), who show that the contribution from atmospheric cooling decreases along the path of the WSC as the ice cover gets more solid.
A quiescent Arctic interior with weak vertical diffusion is crucial for the maintenance of the cold halocline layer (Fer 2009) as well as the circulation of the AW layer (Zhang and Steele 2007). In the absence of storm events and eddies, oceanic heat fluxes in the deep basin cannot contribute significantly to the ocean surface heat budget. Krishfield and Perovich (2005), using basinwide drifting buoy observations, estimate an annual oceanic heat flux of 3–4 W m$^{-2}$ to the Arctic pack ice, with significantly larger values in summer. In contrast, ice-tethered buoy observations in the Canada Basin thermocline yield diffusive heat fluxes about one order of magnitude less than the annual mean heat flux to the ice (Timmermans et al. 2008). Although relatively elevated turbulence is observed over deep topography (Rainville and Winsor 2008), most of the mixing and removal of heat from the AW will occur along the Arctic perimeters, and possibly dominate basin-averaged heat flux. The warm and saline AW layer propagates into the Arctic Ocean as a boundary current flowing along the continental slope. Recently, Lenn et al. (2009) reported that measured vertical mixing of the boundary current on the east Siberian continental slope could not account for the freshening and cooling of the AW, suggesting the influence of shelf waters and lateral mixing.

This study presents directly measured surface turbulent fluxes in addition to turbulent fluxes and mixing inferred from high-resolution turbulence profiling at sites where similar datasets are scarce. The results show that mixing and heat transfer within the branches of the WSC and over the shelf areas close to the Fram Strait are several orders of magnitude larger than those reported by Lenn et al. (2009) farther upstream. Outside the branches of the WSC and off the shelves, however, we observe vertical mixing and oceanic heat fluxes similar to values found in the central Arctic. The large spatial variability and the lack of turbulent mixing in the AW boundary current along the continental slope farther upstream suggest that significant cooling of AW occurs in local regions north of Svalbard with strong currents and tidal forcing over topography.
mixing is important for the regional heat budget, hence the ice cover and cooling of the WSC in the gateway to the Arctic north of Svalbard. In contrast, mixing by processes including lateral mixing, shelf–basin exchange, and intrusions of cold shelf waters sinking downslope may be more typical of the Arctic shelves. Although our measurements are characterized by significant lateral and temporal variability, they represent an important contribution to the understanding of processes by which the core of the WSC loses heat, and provide support for previous hydrography-based estimates of vertical heat fluxes.

Acknowledgments. Constructive comments and suggestions of the two anonymous reviewers are appreciated. This study partly received support from the Research Council of Norway through the project CuNoS: Current measurements north of Svalbard (178919/S30). We thank the captains and the crews of R/V Polarstern and R/V Lance, the oceanographic group of the ARK XIX/1 cruise, and the UNIS students of the R/V Lance cruises.

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