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14. ABSTRACT This project supported the doctoral research of Ms Laura Crews, who focused on atmosphere-ice-ocean interaction using observations collected as part of the ONR Stratified Ocean Dynamics of the Arctic (SODA) DRI. The dissertation is composed of three distinct studies. The first explores the impact of sea ice melt water on timing and small-scale lateral structure of sea ice freeze-up, demonstrating that the advected melt waters can accelerate freeze-up. The second study explores the impact of large lead systems on the generation of near-inertial oscillations in the upper ocean, while the third explores the role of sea ice in modulating radiation of near-inertial energy out of the mixed layer.						
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Stratification at Ocean Fronts

Student Support

N00014-18-1-2694

Craig M. Lee,

Applied Physics Laboratory, University of Washington

craiglee@uw.edu

Long-term Goals

The decreasing trend in minimum sea-ice extent in the Arctic Ocean has been a topic of concern with far reaching impacts. At least seasonally, there are good reasons to believe that the Arctic Ocean will become a more active ocean, with larger surface waves and larger area exposed to summer solar and wind forcing. Particularly near the Marginal Ice Zone (MIZ), the mix of ice and open water, combined with variations in ice surface and keel roughness, lead to a complex physics that vary over short spatial and temporal scales. Ocean properties and air-ice-ocean fluxes are therefore highly heterogeneous near the ice. This project links several data sets and approaches from recent ONR projects, focusing on the air-ice-ocean interactions near the ice edge. The fine and sub-mesoscale processes considered here are important for the evolution of the Arctic system, but are often poorly resolved in observations and ignored in numerical models.

This project supports Ms. Laura Crews, a graduate student in the University of Washington Oceanography program. Her work involves a combination of the analysis of existing observational data and numerical modeling to study sea ice and upper ocean dynamics in the Arctic Ocean.

Note that this represents a change in focus from the original mid-latitude submesoscale dynamics. The scales of interest remain the same but are now couched in a high-latitude setting.

Work Completed

Ms. Crews received Master of Science degrees in Applied Mathematics (December 2020) Oceanography (June 2021). Her MS thesis used high-resolution measurements of the upper ocean collected by ship-based profiling and autonomous vehicles (Seagliders and Wavegliders operated as part of the Stratified Ocean Dynamics of the Arctic Ocean [SODA] program) and one-dimensional mixed layer models to investigate the role of sea ice melt water on freeze-up in the central Beaufort Sea. Meltwater advection can dramatically alter upper ocean structure, shallowing the mixed layer and thus reducing the cooling required to bring mixed layer temperature to freezing. Meltwater advection can also drive mixed layer cooling, which the SODA measurements find to be nearly as large as heat lost to the atmosphere.

Comparisons of model results and observations show that sea ice forms in areas impacted by melt water advection several days prior to freeze-up in nearby areas

that are uninfluenced by meltwater, pointing to the importance of characterizing small-scale, three-dimensional upper ocean structure when forecasting sea ice evolution. The resulting paper (Crews et al., 2022) has been published in the Journal of Geophysical Research: Oceans.

Ongoing work focuses on understanding how the presence of sea ice impacts the generation and propagation of near-inertial waves. Year-long timeseries of upper ocean velocity, temperature and salinity, collected as part of SODA, are being used to explore the role of sea ice internal stresses in modulating generation of upper ocean near-inertial oscillations. A series of numerical experiments using a simple ice-ocean slab model modified to simulate the impact of sea ice internal stress show a damped response. These studies suggest that large scale lead systems may modulate sea ice response to wind-forcing. When leads are present, sea ice is free to respond across a range of time scales. In the absence of such leads, sea ice motion is constrained, damping response at inertial time scales while still allowing longer timescale motions. A manuscript describing these results has been drafted.

Ms. Crews participated in the September-October 2021 NORSE cruise, focused on mesoscale and submesoscale dynamics in the Nordic Seas. She attended a Winter School on Atmosphere-Ocean-Ice Interaction Processes, taught at the University Centre in Svalbard, and presented her work at the 2021 AGU fall meeting and at the 2022 Gordon Conference on ocean mixing. She has also participated in the 2023 MPOWIR Patullo Conference, presented at the final Nansen Legacy Symposium (Tromsø, 2023) and at the Norwegian Polar Institute.

Products

Crews, L., Lee, C. M., Rainville, L., & Thomson, J. (2022). Direct observations of the role of lateral advection of sea ice meltwater in the onset of autumn freeze up. *Journal of Geophysical Research: Oceans*, 127, e2021JC017775.
<https://doi.org/10.1029/2021JC017775>.

Crews, L., Lee, C. M., Rainville, L., & S. Brenner (2023). Sea ice leads increase ice and ocean inertial oscillation strength during and Arctic winter storm. Manuscript in preparation.

Direct Observations of the Role of Lateral Advection of Sea Ice Meltwater in the Onset of Autumn Freeze Up

Laura Crews^{1,2} , Craig M. Lee² , Luc Rainville² , and Jim Thomson² 

¹School of Oceanography, University of Washington, Seattle, WA, USA, ²Applied Physics Laboratory, University of Washington, Seattle, WA, USA

Key Points:

- High spatial and temporal resolution observations by ship-based and autonomous instruments in the ice-free Beaufort Sea are presented
- Modified meltwater advected about 100 km over several weeks, cooling and shoaling the mixed layer and hastening freeze up by several days
- Meltwater advection caused nearly as much mixed layer heat loss as was caused by seasonally integrated heat loss to the atmosphere

Correspondence to:

L. Crews,
lcrews@uw.edu

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Abstract In seasonally ice-free parts of the Arctic Ocean, autumn is characterized by heat loss from the upper ocean to the atmosphere and the onset of freeze up, in which first year sea ice begins to grow in open water areas. The timing of freeze up can be highly spatially variable, complicating efforts to provide accurate sea ice forecasting for marine operations. While melt season anomalies can be used to predict freeze up anomalies in some parts of the Arctic, this one-dimensional view merits further examination in light of recent work demonstrating the importance of three-dimensional flows in setting mixed layer properties in marginal ice zones. In this study, we show that horizontal advection of sea ice meltwater hastens freeze up in areas distant from the ice edge. We use nearly 800 temperature and salinity profiles along with satellite imagery collected in the central Beaufort Sea in autumn 2018 to document the roughly 100 km advection of a cold and fresh surface meltwater layer over several weeks. After the meltwater arrived, the mixed layer was cooler and shallower than the mixed layer in adjacent areas unaffected by the meltwater. The cooler and shallower meltwater-influenced mixed layer promoted earlier ice formation. Within the meltwater-affected area, advection was nearly as important as heat loss to the atmosphere for seasonally integrated mixed layer heat loss.

Plain Language Summary In large parts of the Arctic Ocean, sea ice melts completely in summer and reforms in autumn in a process known as “freeze up.” The timing of freeze up may affect energy exchanges between the ocean and atmosphere, such as occur during storms, as well as impact the sea ice ecosystem. Warming ocean temperatures mean freeze up is occurring later, lengthening the season in which vessels that are not ice-rated can engage in operations like shipping, resource extraction, and supplying local communities with fuel and cargo. Accurate freeze up forecasting helps ensure these operations are conducted safely. This study uses ocean temperature and salinity data from autonomous vehicles as well as data from satellites to demonstrate how areas that were affected by recent sea ice melt can freeze up early. We observed a cold and fresh footprint of meltwater near sea ice in the Beaufort Sea that flowed away from its original location and altered the upper ocean in neighboring areas. A simple model that simulates the atmosphere's influence on the ocean demonstrates that this meltwater caused the ocean to freeze several days earlier than it otherwise would have by cooling the upper ocean and limiting the depth of ocean mixing.

1. Introduction

The autumn freeze up, during which sea ice begins to expand in seasonally ice-free seas, is preceded by an autumn cooling period in which heat fluxes from the ocean to the atmosphere cool the ocean mixed layer toward the freezing temperature, a prerequisite for ice formation. Once the mixed layer has cooled to the freezing temperature, additional heat loss to the atmosphere results in the growth of ice crystals floating on the ocean surface. How freeze up proceeds from this point depends on atmospheric conditions, the sea state, and the stratification below the mixed layer. In calm conditions, the frazil crystals freeze together into a thin sheet called nilas (Wadhams, 2000). When surface waves are present, the frazil crystals instead freeze into slushy pancakes which eventually coalesce into floes (Wadhams, 2000). Ice thickens if the upward heat flux through the ice exceeds the heat flux from the underlying ocean (Wadhams, 2000). However, mixed layer deepening caused by, e.g., brine rejection or storms may liberate heat from the ocean pycnocline and slow or temporarily reverse the freeze-up process (Smith et al., 2018).

The autumn ice advance date is occurring later throughout the Arctic Ocean (Stroeve et al., 2014; Thomson et al., 2016), with the most pronounced shifts in phenology found in the Chukchi and Beaufort Seas (Thomson et al., 2016). Later freeze up is due to warmer sea surface temperatures (SSTs), which are caused by increased solar

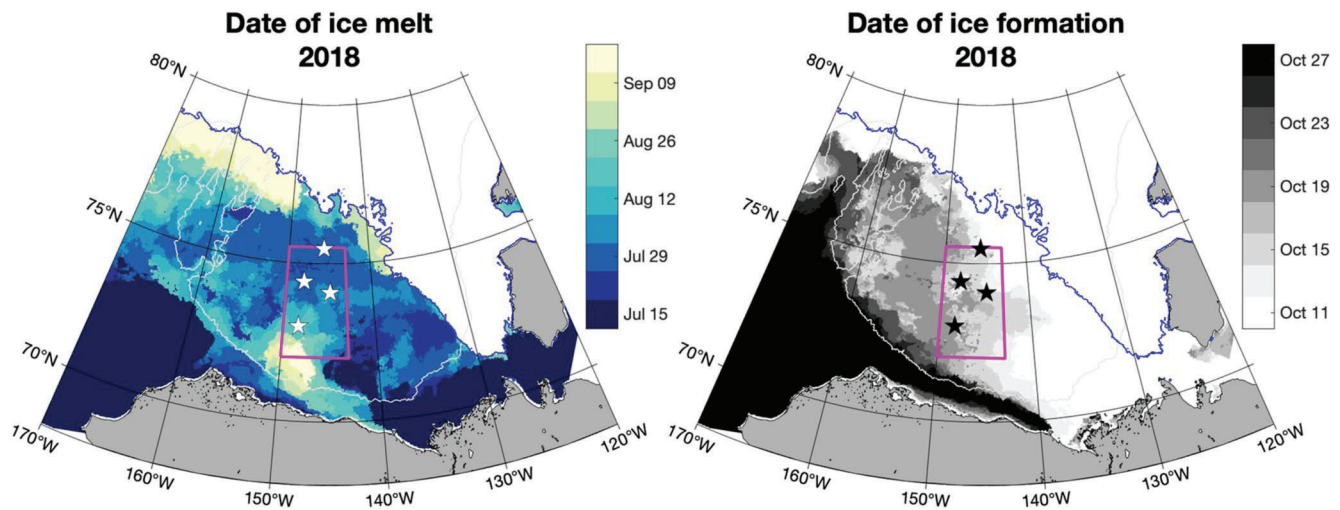


Figure 1. Spatial variability of (left) melt out and (right) freeze up dates in the Beaufort Sea in 2018. Ice retreat was defined as the first day AMSR2 ice concentration fell below 0.15 for five consecutive days and freeze up was defined as the first day AMSR2 ice concentration exceeded 0.15 for five consecutive days. Ice never retreated in the area bounded by the blue contour. Solid gray regions delineate landmasses, light gray contours mark the 1,000-m isobath, and white stars indicate mooring locations. The pink box marks the region depicted in Figure 2.

absorption throughout the summer (Stroeve et al., 2014). A variety of definitions for freeze up have been used in the literature, and the timing of freeze up can depend on which definition is chosen (Johnson & Eicken, 2016). Thomson et al. (2016) defined the ice advance day as the first day that sea ice concentration derived from passive microwave satellites exceeded 15% for 5 days. Freeze up in the Beaufort Sea in autumn 2018, visualized using this metric, was patchy on smaller spatial scales (10–100 km; Figure 1) than those over which the atmospheric forcing is expected to vary (100–1,000 km), pointing to the importance of ocean processes in generating freeze up timing variability on these smaller scales.

The timing of freeze up is important for several reasons. First, ice inhibits heat exchange between the ocean and atmosphere once it develops sufficient thickness (e.g., Maykut, 1978), so earlier or later ice formation would be expected to limit or extend the period in which meaningful heat exchange can occur. In addition, the timing of freeze up matters because the presence of ice alters the efficiency of atmosphere-ocean momentum transfer (e.g., Brenner et al., 2021) and air-sea gas exchange (Islam et al., 2017). Furthermore, the abundance of protists incorporated into sea ice decreases with later freeze up, which in turn decreases ice protist abundance at the beginning of the spring bloom (Niemi et al., 2011). Finally, accurate freeze up forecasting at operational time scales is essential for vessels to safely navigate these waters.

The mixed layer of the Beaufort Sea at the end of summer is shallow, fresh and stratified primarily due to ice melt, and warm due to solar heating (Jackson et al., 2010; Toole et al., 2010). The continuous effect of cooling and freshening associated with summer ice melt competes with the solar heating, which can result in a temperature maximum below the mixed layer (near-surface temperature maximum or “NSTM,” Jackson et al., 2010; Steele et al., 2011). Beneath the NSTM is a temperature minimum marking the remainder of the previous winter's mixed layer (Jackson et al., 2010). Below this minimum are the warm Pacific Summer Water layer, the cold Pacific Winter Water layer, and the warm Atlantic Water layer, all of which were advected from Arctic Ocean boundaries. These boundary-derived layers do not typically influence the surface in the Beaufort Sea on seasonal time scales because of the strong near-surface stratification (Toole et al., 2010).

Studies of upper ocean processes in the Arctic often emphasize one-dimensional mixed layer evolution forced by momentum and buoyancy exchanges with the overlying sea ice or atmosphere (e.g., Dewey et al., 2017; Toole et al., 2010). However, mixed layer instabilities at submesoscale fronts have been shown to play a significant role in setting ocean mixed layer properties (e.g., Boccaletti et al., 2007; Thomas et al., 2008). In the Beaufort Sea, restratification under sea ice was not replicated by a one-dimensional model (Toole et al., 2010), and Timmermans et al. (2012) argue that this restratification was likely caused by submesoscale processes on the basis of frontal structure observations and $O(1)$ balanced Richardson number estimates.

Subsequent observations and modeling studies have shown that submesoscale fronts are found in marginal ice zones (MIZs) where surface freshening due to ice melt causes strong horizontal surface buoyancy gradients over length scales of 1–10 km (e.g., Brenner et al., 2020; Horvat et al., 2016; Lu et al., 2015; Manucharyan & Thompson, 2017). Fronts in the Arctic are also found at the confluence of different water masses, e.g., in Fram Strait (von Appen et al., 2018), or river plumes (Alkire et al., 2019; Macdonald et al., 1995). Observations demonstrate an active submesoscale field in the Beaufort Sea (Mensa et al., 2018), in the neighboring Chukchi Sea (Timmermans & Windsor, 2013), as well as in Fram Strait (von Appen et al., 2018).

Basin-scale observations in the Beaufort Sea indicate that month-to-month changes in mixed layer temperature and salinity are largely due to one-dimensional processes (Dewey et al., 2017). Here, we expand on this regionally one-dimensional view using mesoscale-resolving observations of ocean temperature and salinity in the weeks preceding freeze up. We find that the horizontal advection of a shallow meltwater front cooled and restratified the mixed layer. These lateral effects allowed freeze up to occur earlier in the meltwater than in adjacent waters, highlighting an important consequence of mixed layer heterogeneity and three-dimensional processes.

This study is organized as follows. Section 2 outlines the ocean observations and remote sensing products used in this study and quantifies differences between in situ and remote sensing observations. Section 3 describes the meltwater and its advection. Section 4 develops a mixed layer heat budget to quantify the importance of meltwater advection to mixed layer heat loss. Section 5 describes an additional instance of freshwater advection and its effect on freeze up. Section 6 summarizes and discusses the results.

2. Methods

2.1. Overview

This study uses observations collected in the Beaufort Sea in September–October 2018, as part of the Stratified Ocean Dynamics of the Arctic (SODA) experiment, to investigate mechanisms that drive freeze up heterogeneity. The SODA experiment, sponsored by the U.S. Office of Naval Research, aims at understanding stratification changes in the Arctic Ocean and includes many other components (Lee et al., 2016; Rainville et al., 2020). The upper ocean's evolution was documented by mesoscale-resolving upper ocean temperature and salinity profiles from four autonomous Seagliders. The persistent Seaglider sampling was supplemented by episodic sampling from four Wave Gliders and a ship-based underway CTD (uCTD). The locations of these measurements are shown in Figure 2. Conservative temperature and absolute salinity were calculated using the Gibbs SeaWater (GSW) Oceanographic Toolbox Version 3.0.11. All data collected as part of the field campaign (i.e., not by satellites) and discussed in this paper are reported as conservative temperature and absolute salinity unless otherwise noted. Potential density was calculated for a reference sea pressure of 0 dbar. Additional context was provided by remote sensing products and a one-dimensional mixed layer model.

2.2. Upper Ocean Temperature and Salinity Observations

2.2.1. Seagliders

Seaglider autonomous vehicles sampled ocean temperature and salinity between the surface and 1,000 m. Seagliders employ a saw-tooth dive pattern with ~5-km horizontal spacing between profiles at the surface. Temperature and salinity data were averaged into 1-m depth bins and profiles lacking data above 15 m were not used. If the remaining temperature or salinity profiles did not extend to the surface, the shallowest measured temperature or salinity was extrapolated as a constant value to the surface because we assume that these depths were well mixed.

Five Seagliders were deployed in open water on 21 September 2018; two of these were recovered after a few days due to technical issues. The three remaining Seagliders sampled two quasi-parallel lines spanning roughly 73.6°N 147.4°W to 74.9°N 145.3°W, (Figure 2). Each line was sampled by Seagliders operating in tandem, with one or two Seagliders traversing half a line. This sampling strategy provided complete occupation of the line at a time scale of a few days. Data used in this study span 21 September to 15 October, with the end of the study period characterized by active freeze up (Figure 1). Two Seagliders continued sampling intermittently after freeze up and were lost in the ice while trying to overwinter, returning data from as late as February 2019. During the study period, one of the Seagliders made a one-time excursion north of the sampling line to assess conditions near the MIZ. Sampling by an additional Seaglider away from the main Seaglider line occurred on 26 September.

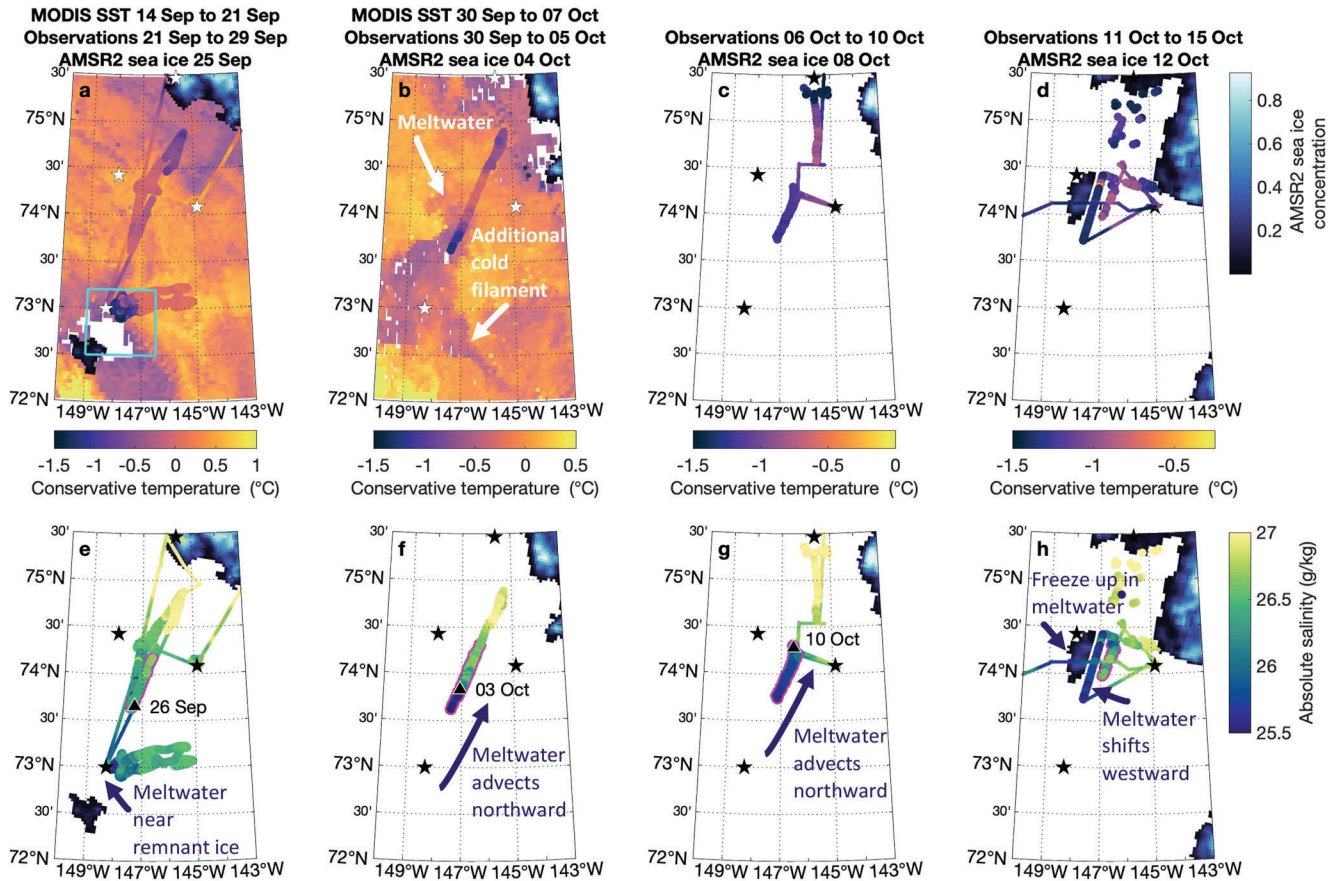


Figure 2. Upper ocean observations and remote sensing show the meltwater's advection in the weeks preceding freeze up. (a–d) Conservative temperature and (e–h) absolute salinity observations, including Seaglider and underway CTD (uCTD) data averaged in the upper 5 m as well as Wave Glider data at 5 or 9 m (large scatter points) and *Healy* underway data at 2.7 m (thin line), at the time periods specified in the column titles. MODIS-Terra sea surface temperature (SST) is included in the first two panels and the AMSR2 ice concentration is shown where it exceeded zero. The black triangles are the locations of the 26 g/kg isohaline outcrop, a proxy for the edge of the meltwater. Seaglider profiles outlined in pink were used in the mixed layer heat budget developed in Section 4. Stars indicate mooring locations.

2.2.2. Wave Gliders

Four Liquid Robotics SV-2 Wave Gliders sampled near 73°N 147°W from 19 to 23 September as part of the SODA Alaska North Slope process study (MacKinnon et al., 2021). The Wave Gliders sampled temperature, conductivity, and pressure using Sea-Bird “Glider Payload” CTDs. The low autumn sun angle at this latitude limited solar recharging so sampling was performed at 30-min intervals to increase vehicle endurance, resulting in ~1-km horizontal sampling resolution. The nominal measurement depths varied by vehicle; the most common configuration was temperature sampling at 0.2 and 9 m as well as conductivity sampling at 9 m. One vehicle also measured conductivity at 0.2 m, using an Aanderraa 4319B sensor. The exact depth varied slightly for each measurement as the vehicle moved.

2.2.3. Ship-Based Measurements

Temperature and salinity profiles were collected by uCTD at ~1-km horizontal spacing on 13 October. The uCTD recovery line was routed through the extended arm of USCGC *Healy*'s starboard crane to minimize sampling in the ship's wake. The uCTD was deployed in “tow-yo mode,” meaning it was not recovered between casts, and data above 1 m were excluded to eliminate errors associated with identifying the start of the cast. Data from the down cast were used and sampling depths were determined by the probe's free-fall time and the ship's speed; casts were typically to 130–160 m with maximum cast depths of about 180 m.

Near-surface ocean temperature and salinity were also collected by *Healy* while underway. The seawater intake was 2.7 m below the ship's waterline. Underway salinity data are known to drift over time due to biofouling of

the conductivity cell (Alory et al., 2015) and temperature data are biased because seawater warms while traveling through the ship system before its temperature is measured (Alory et al., 2015). To correct the underway data, we computed the difference between uCTD conservative temperature and absolute salinity profiles averaged in the upper five meters and concurrent underway data. The underway data used in the comparison were the median value in 1-hr windows centered on the times of the uCTD profiles. To correct underway absolute salinity, a linear fit of the differences between the uCTD absolute salinity and colocated underway absolute salinity was used to compute a continuous correction time series that was added to the underway absolute salinity (Alory et al., 2015). To correct underway conservative temperature, the median difference between the uCTD conservative temperature and colocated underway conservative temperature (-1.3°C) was added to the underway conservative temperature.

2.3. Mixing Layer and Mixed Layer Definitions

Using data collected from Ice-Tethered Profilers, Timmermans et al. (2012) distinguished between the Beaufort Sea surface layer, defined as a density change of 0.25 kg/m^3 relative to the shallowest measurement, and the actively mixing layer, defined as a density change of 0.01 kg/m^3 relative to the shallowest measurement. They found that mixing layers were common and attributed them to lateral restratification. Inspection of the density profiles in this study showed that mixing layers were common in open water as well, so we defined the mixing layer using a density change of 0.01 kg/m^3 and the mixed layer using a density change of 0.25 kg/m^3 (a scheme also used in Cole et al. (2014)). We note that using a density difference criterion to define the mixing layer will not necessarily indicate active turbulence and may bias results by including mixing that has already occurred.

2.4. Ocean Heat Content

Ocean heat loss in an Eulerian framework was quantified by calculating the change with time in ocean heat content. The ocean heat content was calculated from each conservative temperature profile as

$$\text{HC} = \int_h^0 c_p \rho (T(z) - T_{fr}(z)) dz \quad (1)$$

where ρ is the measured potential density, c_p is the heat capacity, and T_{fr} is the freezing temperature calculated from the observed absolute salinity and pressure. Section 2.2.1 stated that missing surface data in the measured temperature profile $T(z)$ were filled with a constant value, but this interpolation is not expected to impact the results related to ocean heat content as it only affected two profiles used in the heat content analysis (Section 4; the shallowest available measurements were at 2 and 3 m). Heat was integrated using the trapezoidal method from the surface to the depth h . In this paper, we will discuss the mixed layer heat content, for which h is the depth at which density increased by 0.25 kg/m^3 relative to the surface, as well as the upper ocean heat content including the mixed layer and top of the pycnocline for which h is a specified constant depth.

2.5. Remote Sensing and Reanalysis

MODIS-Terra nighttime SST 8-day composite images were used to supplement in situ ocean temperature observations. Only data with quality flag values of 0 or 1 were used.

Dynamic ocean topography (DOT) collected by CryoSat-2 from early September to early October was used to calculate surface geostrophic velocity (Scharroo et al., 2013).

Daily sea ice concentrations from AMSR2 (a passive microwave instrument) were used to map sea ice extent and assess the timing of freeze up (Spreen et al., 2008).

Synthetic Aperture Radar (SAR) satellite imagery provided occasional high resolution ($\sim 50\text{ m}$) representation of the ice conditions. Backscatter characteristics can be used to clearly distinguish open water, sea ice floes, sea ice ridges, leads, and ice type with a much higher level of detail than passive microwave sea ice products (e.g., Kwok et al., 1999).

In situ salinity observations were compared to the weekly SMOS ARCTIC sea surface salinity product (Supply et al., 2020a, 2020b) which attempts to correct for biases caused by cold water (Köhler et al., 2015) and the presence of sea ice (Tang et al., 2018).

To force the mixed layer model (Section 2.7), hourly mean net longwave, net shortwave, sensible, and latent heat fluxes, eastward and northward turbulent surface stresses, and evaporation and precipitation were downloaded from ERA5 reanalysis at 0.25° resolution (Hersbach et al., 2018), at this latitude equal to roughly 8-km resolution in the zonal direction and 28-km resolution in the meridional direction.

ERA5 combines observations and model output, and in the data-sparse Arctic the number of independent observations is limited. Uncertainty in ERA5 is estimated from the ensemble spread, which depends in part on the availability of observations (Hersbach et al., 2018). This ERA5 ensemble spread accounts for random uncertainties in the observations, SST, and model physical parameterizations, but not for systemic errors in the model. The ERA5 ensemble spread is not meant to estimate error at a specific point in space and time, but rather to show where the reanalysis is more or less reliable (<https://confluence.ecmwf.int/display/CKB/ERA5%3A+uncertainty+estimation>). Examining the ERA5 ensemble spread from 21 September to 21 October 2018 in the Beaufort Sea, the Bering Sea, and Gulf of Alaska for all the ERA5 parameters (heat fluxes, surface stresses, evaporation, and precipitation) used to force the mixed layer model showed transiently heightened uncertainty coinciding with passing storms. There were some regions of persistently high uncertainty as well, such as for latent heat flux near the Alaskan coast south of Bering Strait. Overall, it does not appear that our study area suffers from particularly large ERA5 ensemble spread compared to other regions, except for somewhat elevated ensemble spread for sensible heat flux.

Graham et al. (2019) compared January to May meteorological observations collected north of Svalbard to six atmospheric reanalyses, including ERA5. They found that the reanalyses were highly correlated with the observations for wintertime surface meteorological parameters and downwelling longwave radiation, but that all the reanalyses showed poor skill in simulating turbulent heat fluxes over sea ice (Graham et al., 2019). As our study relies on data from Seagliders which do not collect meteorological data, we proceed using the reanalysis, acknowledging ERA5 is an imperfect product that may not accurately capture local conditions, especially in the vicinity of the ice edge.

2.5.1. A Comparison of Observed and Remotely Sensed Sea Surface Temperature and Salinity

The observations presented in this paper provide an opportunity for comparison with concurrent remote sensing. Each conservative temperature observation was compared to MODIS-Terra nighttime SST averaged within a 15 km box centered on that observation (Figure 3a). The closest daily MODIS image in time with any data in the spatial averaging box was used, with candidate MODIS images occurring up to 3 days before or 3 days after the observation. Salinity observations were compared to weekly SMOS ARCTIC sea surface salinity (Supply et al., 2020a). The observed salinity on the Practical Salinity Scale PSS-78, rather than the observed absolute salinity, was used to match the units used by the satellite product. Observations from the Seagliders (averaged in the upper 3 m), Wave Gliders (at 0.2 m), uCTD (averaged in the upper 3 m), and *Healy* underway system (at 2.7 m) were distributed unevenly in time, space, and by platform (i.e., all Wave Glider measurements were from a short period of time in a limited area and the *Healy* underway system's rapid sampling meant there were far more underway samples). To account for this, the *Healy* underway and Wave Glider data sets were subsampled at 15-min and 150-min intervals, respectively, to create a comparison data set more representative of the entire region and time period. Plots of in situ data from individual platforms versus satellite data showed the same general patterns as the multiplatform comparison.

The MODIS SSTs tended to be warmer than observations, with this difference exacerbated at temperatures of $<0^{\circ}\text{C}$ (i.e., as the ocean approached freeze up, Figure 3a). MODIS measures the skin SST, which is generally lower than the underlying ocean temperature due to the temperature gradient resulting from heat flow from the ocean to the atmosphere (Minnett et al., 2019). The temperature change across the thermal skin averages 0.17°C , with larger differences expected for low wind conditions (Minnett et al., 2019). The thermal skin effect undoubtedly contributed to the temperature difference between the observed and MODIS SST, but as the MODIS SSTs were generally warmer than the observed SSTs, with differences of $\sim 1^{\circ}\text{C}$ at lower temperatures (Figure 3a), we do not expect the thermal skin effect to be the primary reason for disagreement between the MODIS and observed SSTs.

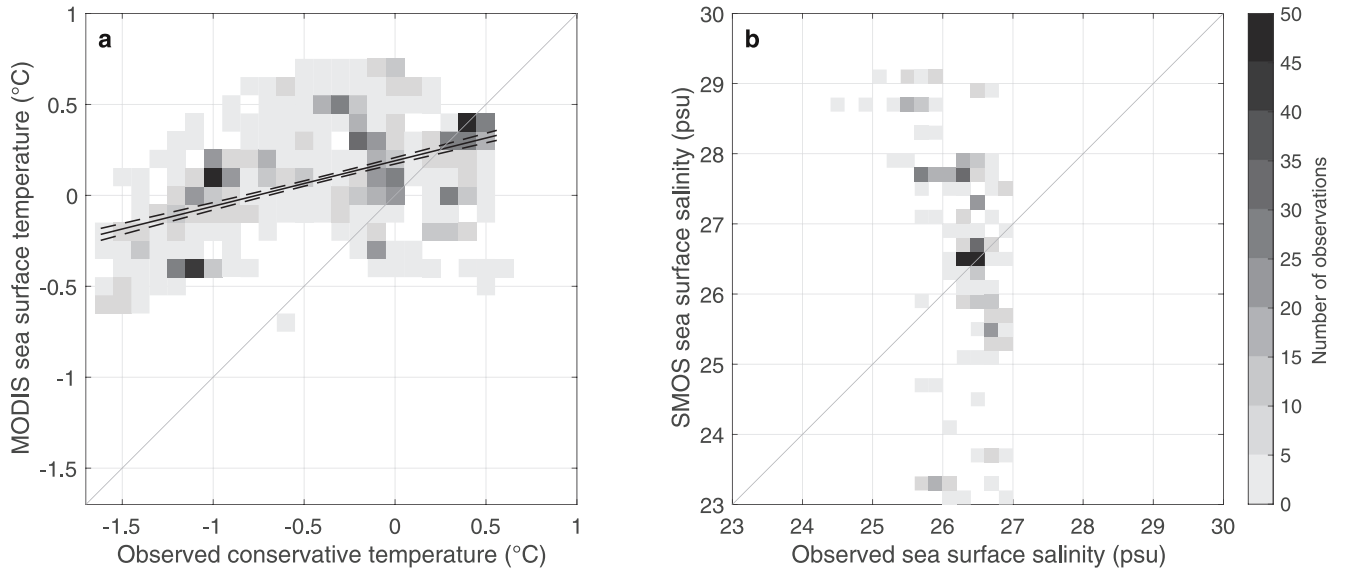


Figure 3. Two-dimensional histograms comparing (a) conservative temperature and (b) practical salinity from Seagliders, Wave Gliders, uCTD, and the *Healy* underway system ("observed") to concurrent satellite data. Satellite data were from (a) daily MODIS-Terra and (b) weekly SMOS ARCTIC (Supply et al., 2020a). The solid line in (a) is the linear regression and the dashed lines are the 95% confidence bounds.

The standard deviation of the differences between in situ sea surface salinity (SSS) and concurrent SMOS SSS was 1.66 psu, with larger differences common at in situ salinities typical of the meltwater (Figure 3b). The L-band radiometer used to derive satellite SSS only measures the top centimeter of the ocean, so near-surface stratification contributes to differences between observed and SMOS SSS, with saltier observations expected when the stratification is stronger or the salinity data are collected deeper in the water column (Supply et al., 2020b). However, the observed salinities in the meltwater were *fresher* than the satellite SSS (Figure 3b), meaning stratification cannot explain the disagreement. Previous comparisons between satellite SSS and in situ Arctic Ocean data show relatively large discrepancies in regions with cold water and sea ice contamination (Fournier et al., 2019), in line with our finding of larger discrepancies in meltwater conditions.

These comparisons demonstrate the importance of in situ observations, particularly for smaller scale features or conditions typical of cold and stratified meltwater. As a further consideration, this study will demonstrate the importance of mixed layer depth variability to freeze up timing variability (Section 4), the characterization of which requires in situ observations. Despite these shortcomings, the MODIS data in particular provided useful regional-scale context in visualizing the extent of the meltwater (Figure 2).

2.6. Surface Velocities

To calculate geostrophic velocity, DOT data were linearly interpolated to a regularly spaced latitude/longitude grid with ~15-km spacing and smoothed with a 2-D Gaussian kernel using the MATLAB `imgaussfilt` function with a standard deviation of 0.5. Geostrophic velocities were calculated from the smoothed $\text{DOT} = \eta$ as

$$u_g = -\frac{g}{f} \frac{\partial \eta}{\partial y}, v_g = \frac{g}{f} \frac{\partial \eta}{\partial x} \quad (2)$$

The average velocity in the Ekman layer was calculated from ERA5 horizontal turbulent surface stresses τ^x and τ^y as

$$v_{Ek} = -\frac{\tau^x}{f \rho_0 D_{Ek}}, u_{Ek} = \frac{\tau^y}{f \rho_0 D_{Ek}} \quad (3)$$

with the gravitational acceleration g , the Coriolis parameter $f = 2\Omega \sin(\text{latitude})$, and the reference density $\rho_0 = 1026 \text{ kg/m}^3$. Ekman theory describes steady flow that develops over time scales longer than an inertial period, so the time series of each component of surface stress at each ERA5 grid point was low-pass filtered with

a fifth-order Butterworth filter to retain stresses with periods of at least 16 hr, slightly longer than the 12.5-hr inertial period at this latitude, before applying Equation 3.

2.7. Mixed Layer Modeling

The atmospheric influence on the ocean mixed layer was examined using the one-dimensional upper ocean mixing model of Price et al. (1986), hereafter called the Price-Weller-Pinkel (PWP) model. In each PWP time step, heat and freshwater are added to (for ocean heating/precipitation) or subtracted from (for ocean cooling/evaporation) the water column. The water column overturns if these subtractions make the density structure unstable, entraining successively deeper depth bins until the stratification is stable. Momentum imparted by surface stress is then added to the mixed layer and additional depth bins are entrained into the mixed layer until the bulk Richardson number R_b and gradient Richardson number R_g exceed critical values

$$R_b = \frac{g\Delta\rho h}{\rho_0(\Delta V)^2} \geq 0.65 \quad (4)$$

$$R_g = \frac{g \frac{\partial \rho}{\partial z}}{\rho_0 \left(\frac{\partial V}{\partial z} \right)^2} \geq 0.3 \quad (5)$$

where h is the mixed layer depth and Δ denotes the difference between the (homogenized) mixed layer and the depth bin just below the mixed layer. The bulk Richardson number requirement simulates entrainment at the base of the mixed layer; if R_b falls below the critical value, pycnocline depth bins are entrained and their properties distributed throughout the mixed layer until the R_b criterion is satisfied. The gradient Richardson number requirement simulates shear flow instabilities in the stratified transition layer below the mixed layer; if R_g falls below the critical value at any point in the stratified part of the profile, adjacent depth bins are mixed until the R_g criterion is satisfied throughout the profile. The theoretical requirement for shear instability is $R_g < 0.25$, but here we follow Price et al. (1986) and mix for $R_g < 0.3$.

We forced the PWP model with surface heat fluxes, surface stresses, and evaporation and precipitation from ERA5 reanalysis (Section 2.5). For each observed profile, ERA5 atmospheric forcing was averaged in a 30-km box centered at the profile's location; most model runs used data from four ERA5 grid cells though some used data from as many as eight ERA5 grid cells, depending on how the observation's location aligned with the ERA5 grid. The net shortwave heat flux was distributed throughout the water column according to an extinction coefficient of 0.38 (Price et al., 1986), an appropriate value for clear, mid-ocean Type 1A water (Paulson & Simpson, 1977). Phytoplankton and CDOM absorption are neglected as their effects on ocean optical properties in the deep Beaufort basin are small (Antoine et al., 2013). The sum of the net longwave, sensible, and latent heat fluxes was added to the shallowest depth bin. The PWP model was used to explore the ocean's evolution as it cooled to freezing, so we assumed that the ocean was ice-free and did not include freshwater/salt additions due to ice melt/growth. The PWP depth bin size was 1 m and the deepest bin was 50 m.

Surface forcing was interpolated to 10-min temporal resolution to run the PWP model with a 10-min time step. Use of a short time step was necessary because when ocean heat loss integrated over a longer time step was extracted from the uppermost model cell, the water in that cell could momentarily fall below the freezing point. Distributing cooling over a larger number of short time steps caused static overturn without the surface cell falling below freezing.

The momentum and scalar diffusivities were set to $10^{-6} \text{ m}^2/\text{s}$ following Dewey et al. (2017). The mixing layer depth in the PWP experiments was the depth at which density was at least 0.01 kg/m^3 greater than the surface density, consistent with the Beaufort Sea PWP experiments of Toole et al. (2010).

3. Observations

3.1. Overview

The following sections show how horizontal advection of sea ice meltwater hastened freeze up. Melting sea ice that lingered in the southern Beaufort Sea created a cold, fresh, and shallow mixed layer (Section 3.2). In situ

Seaglider and ship-based observations documented the northward advection of this feature during the 3 weeks preceding freeze up (Section 3.3, Figure 2). The meltwater advection can be explained by the combination of Ekman and geostrophic velocities (Section 3.4).

3.2. Ice and Ocean Conditions Before Meltwater Advection

Ice retreated from the central Beaufort Sea in late July and early August 2018 (Figure 1). The study area was roughly delineated by four subsurface moorings installed as part of the SODA program; while data from those moorings are not used in this study, mooring locations are marked by star symbols on maps as “landmarks” to orient the reader. The primary ice edge was northeast of the study area, and an additional area of ice separate from the main ice pack persisted in the southern Beaufort Sea. This remnant ice became a distinct patch in the fourth week of August 2018 (Figure 1). At that time, its extent was about $5 \times 10^4 \text{ km}^2$, with an average sea ice concentration of about 50% (not shown). Ice extent and concentration then declined to about $1 \times 10^4 \text{ km}^2$ and 15%, respectively, by mid-September (not shown). The remnant ice then moved northwestward and continued to shrink, leaving the study region ice-free after 28 September. The local decrease in the remnant ice extent (Figure 1) and concentration indicates most of it melted locally in the southern Beaufort Sea.

Initially, the upper ocean was relatively cold only near the ice edges. MODIS SST data from 14 to 21 September showed a band of cold water extending roughly 100 km from the northeastern MIZ as well as a localized cold patch surrounding the melting remnant ice (Figures 2a and 2e). In situ observations agreed with this MODIS SST pattern. Seagliders observed cooler mixed layer temperatures at the northward limit of their transect, aligning with the SST gradient seen in the satellite imagery adjacent to the northern MIZ (Figure 2a). Wave Gliders observed a near-surface (9 m) horizontal conservative temperature gradient that roughly aligned with that in the satellite imagery near the remnant ice (Figures 2a and 4a). Some differences between the MODIS and observed conservative temperatures in Figure 2a could be because the observations were from later than the MODIS period (Wave Gliders 19–23 September, Seagliders 22–26 September; the concurrent MODIS data were too limited by clouds to discern SST gradients). However, a careful comparison of observed near-surface conservative temperature and concurrent MODIS SST showed that MODIS temperatures were warmer than the observations (Figure 3a, additional discussion in Section 2.5.1).

Away from the ice edges, the late summer mixed layer had relatively uniform properties. In the central part of a 22–26 September Seaglider transect the mixed layer was about 25 m deep with conservative temperature of $\sim 0^\circ\text{C}$ and absolute salinity of $\sim 26.5 \text{ g/kg}$ (Figures 5b and 5c). Below the mixed layer, isopycnals sloped downward with increasing latitude (Figures 5b and 5c). The northern MIZ's influence was apparent at the northern end of the transect, where the mixed layer was somewhat colder, saltier, and deeper (depth $\sim 30 \text{ m}$, conservative temperature of $\sim -0.5^\circ\text{C}$, absolute salinity of $\sim 26.9 \text{ g/kg}$, not shown).

Seaglider conservative temperature-absolute salinity profiles collected on 26 September in the cold footprint of the remnant ice show surface conditions qualitatively consistent with modification by local ice melt (Figures 4e and 4f). Ice melt would be expected to cool, freshen, and stratify near-surface waters, and in the Seaglider observations the mixed layer was near-freezing (Figure 4e), relatively fresh, and stratified (Figures 4a and 4f), and was about 10 m deep (Figures 4e and 4f). Absolute salinity increased by about 1 g/kg between the surface and 20 m where an NSTM was $\sim 1\text{--}1.5^\circ\text{C}$ warmer than the relatively cold sea surface.

Sea ice melt releases very low salinity water which can mix downward, increasing stratification in the upper ocean and creating a seasonal pycnocline referred to by Randelhoff et al. (2017) as a meltwater layer, which is distinct from the isolated lenses of pure sea ice melt that can accumulate in quiescent areas under sea ice. Such a stratified surface layer is not pure ice melt and might more accurately be described as “modified meltwater,” but we follow Randelhoff et al. (2017) and call this phenomenon meltwater. We now discuss whether the observed layer of cold, fresh surface water in the remnant ice area (Figures 4e and 4f) could have been caused by local ice melt mixing with ambient water.

Gade (1979) developed an equation to estimate the product properties of ice melt in ocean water which accounts for the latent heat necessary for the change of state and assumes a turbulent boundary between the ocean and the ice. The Gade Line slope $\frac{d\theta}{dS}$, reformulated by Herraiz-Borreguero et al. (2013), is given by

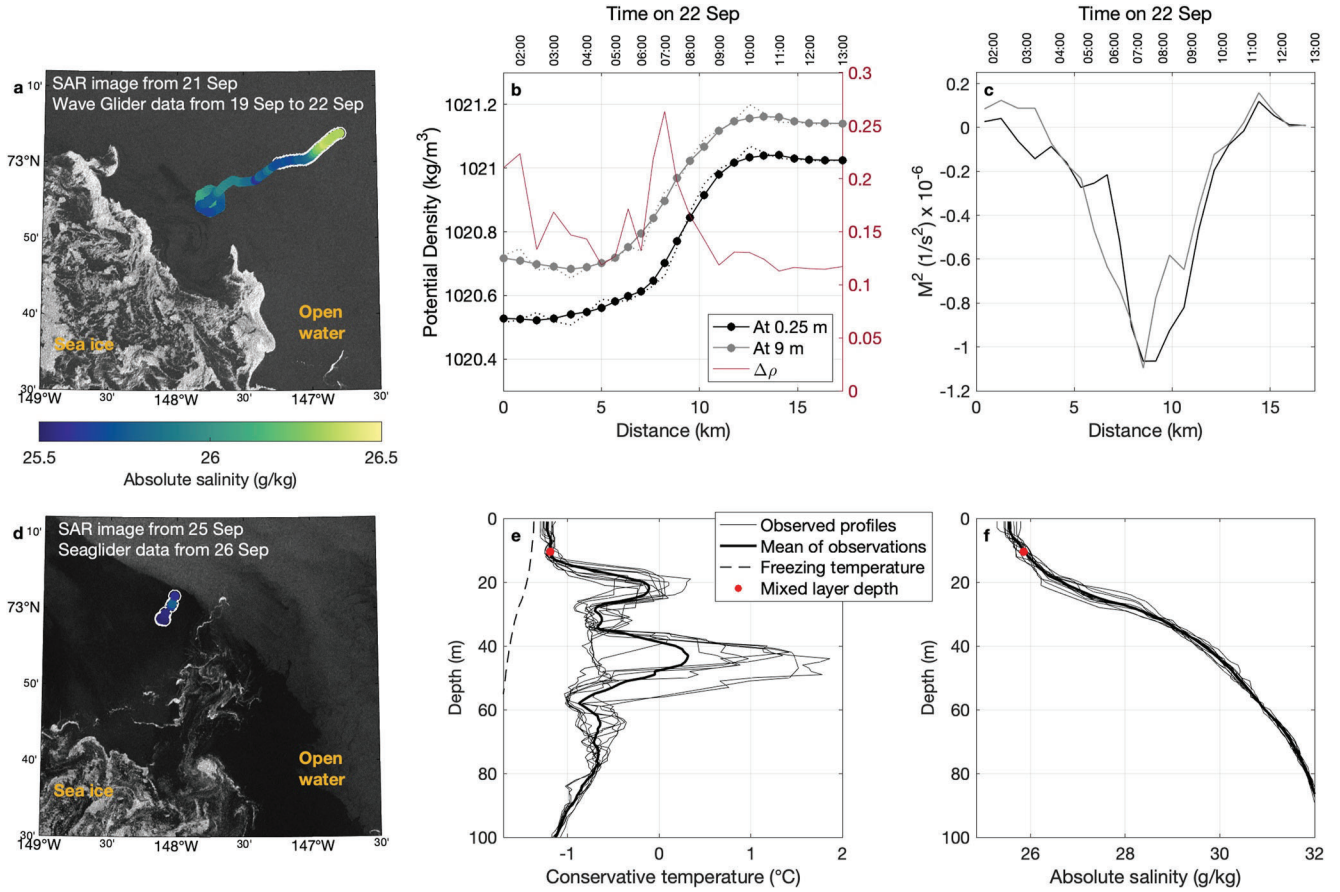


Figure 4. Meltwater conditions near the remnant sea ice. (a) Wave Glider salinity at 0.25 m and sea ice from Synthetic Aperture Radar (SAR) on 21 September in the cyan box in Figure 2, with data from the white-outlined track shown in (b) and (c). (b) Wave Glider potential density at 0.25 m (black lines) and 9 m (gray lines) including the raw data (dotted lines) and data smoothed with a moving mean with a window width of ~ 4 km (solid lines). The density difference between the shallow and deep measurements is shown in red. (c) Along-track horizontal buoyancy gradient M^2 calculated from the smoothed density. (d) Seaglider absolute salinity averaged in the upper 5 m and sea ice from SAR on 25 September. Seaglider profiles, including the average profiles (bold lines) of (e) conservative temperature and (f) absolute salinity. The red dots show the mixed layer depth, defined as a density change of 0.25 kg/m^3 relative to the surface.

$$\frac{d\theta}{dS} = \frac{L}{S_0 c_w} + \frac{(\theta_f - T_i) c_i}{S_0 c_w} + \frac{(\theta_0 - \theta_f)}{S_0} \quad (6)$$

where L is the latent heat of fusion for ice, θ_0 and S_0 are the potential temperature and salinity of the ambient water, θ_f is the ocean freezing temperature, T_i is the ice temperature, and c_w and c_i are the specific heat capacities of water and ice, respectively. As the remnant ice in our study was melting during summer, we assume it was already close to its melting temperature and use $T_i = -2^\circ\text{C}$. The first term on the right-hand side dominates so this assumption has little impact on the results.

Gade Lines show that ice melt by the ambient water observed outside the meltwater would generate product water with salinities between 25.8 and 26 g/kg (Figure 6). The temperature and salinity profiles in the meltwater (black and purple points on Figure 6) approximately follow the Gade Lines from the ambient water conditions (blue points in Figure 6) until around 10-m depth, at which point they freshen further. This suggests an additional source of liquid fresh water was necessary to produce the observed salinities in the upper 10 m.

To estimate the volume of ice melt necessary to generate the observed salinity profiles, we calculate the upper ocean salt deficit. The salt deficit quantifies the thickness of ice growth required to increase the mixed layer salinity to match the salinity at a reference depth (Wilson et al., 2019), but can also be thought of as the ice melt necessary to freshen the surface layer relative to the salinity at a reference depth. For the average Seaglider salinity profile in the remnant ice area (Figure 4f), we can infer that 20 m is roughly the base of the pycnocline formed

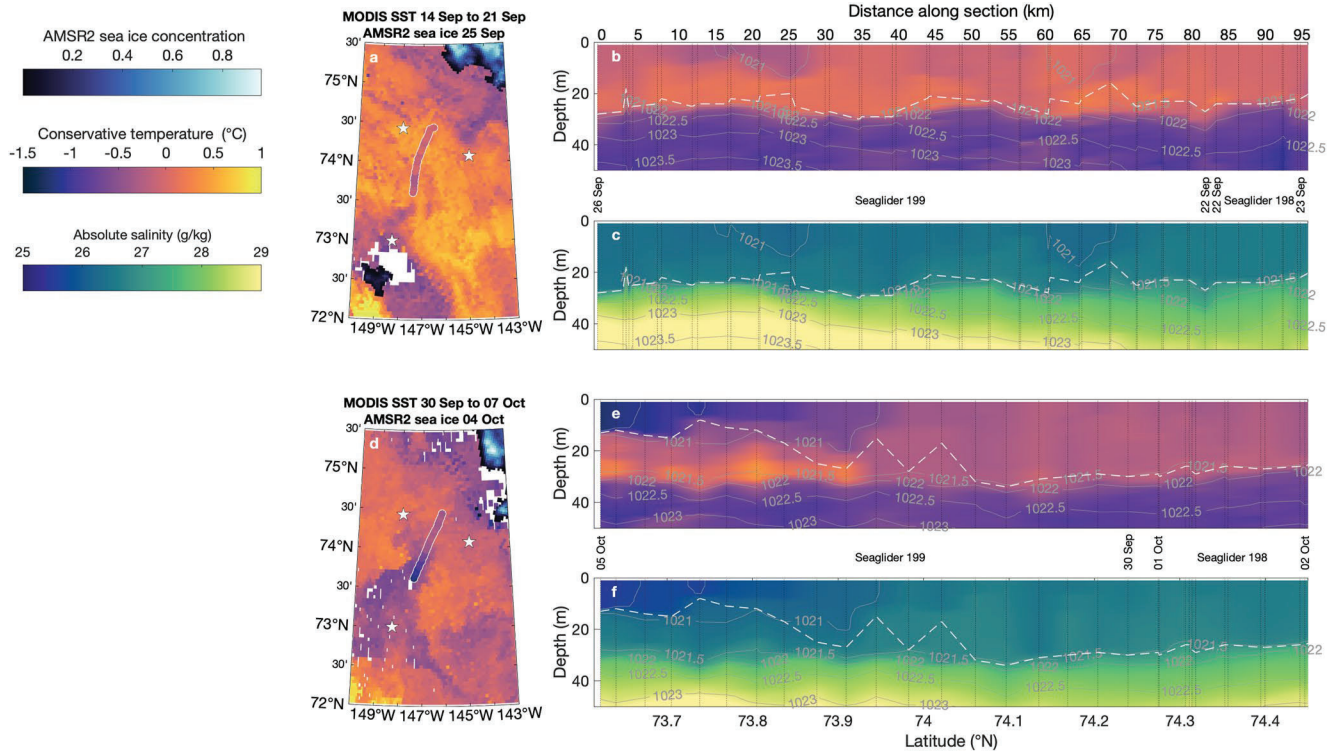


Figure 5. Conservative temperature and absolute salinity in latitude-depth sections measured by two Seagliders from (a–c) 22–26 September and (d–f) 30 September to 5 October. (a) and (d) show profile locations as white-outlined dots, with dot color indicating observed conservative temperature averaged in the upper 5 m. Stars indicate mooring locations. Sampling began partway through the transect with one Seaglider sampling outward in each direction, indicated by the dates. The gray lines on the section plots mark potential density contours at 0.5 kg/m^3 intervals and the dashed white lines mark mixed layer depth defined as a density difference of 0.25 kg/m^3 relative to the surface. Sections show only the upper 50 m of the 1,000-m profiles, with vertical lines indicating individual profiles.

by the ice melt based on the presence of an NSTM below this depth (Figure 4e). The salt deficit SD^* , which has units of ice thickness, is defined as (Wilson et al., 2019)

$$SD^* = \frac{1}{\Delta S_i} \int_0^{z_c} (S(z_c) - S(z)) dz \quad (7)$$

where z_c is the reference depth and ΔS_i is the salinity difference between the ocean surface and the sea ice. Assuming an ice salinity of 6 g/kg , $SD^* = 0.7 \text{ m}$ for $z_c = 20 \text{ m}$. However, the value of SD^* for our salinity profile is quite sensitive to the choice of z_c as the water column remains strongly stratified below 20 m (Figure 4f). Increasing z_c to 30 m to include all of the water column above the temperature minimum associated with the previous winter's mixed layer increases SD^* to nearly 3 m, an unreasonably large amount of local ice melt. However, it seems unlikely that ice melt could have mixed to 30 m while the subsurface temperature maximum between 20 and 30 m persisted. Above 10 m, the depth at which the observed profiles departed from the Gade Line estimates, $SD^* = 0.1 \text{ m}$. These calculations assume 100% ice concentration; attaining the necessary ice melt volume from patchy ice cover requires thicker ice than the SD^* values reported above.

In summary, the salt deficit calculation suggests that the freshwater equivalent of roughly 1–1.5 m of patchy ice melt could result in the fresher surface layer observed in the remnant ice area while the departure of the observed temperature-salinity profiles from the Gade Lines in the upper 10 m suggests that an additional source of liquid freshwater was necessary to fully explain the temperature-salinity properties observed in the meltwater. Precipitation is unlikely as the freshwater front was sharp, requiring highly localized rainfall, and a plume of river water is unlikely at this distance from the coast. Ekman pumping could have accumulated fresher surface water in this area, increasing the apparent freshening beyond that attributable to one-dimensional ice melt. The previous water in this area could have been anomalously fresh before ice melt occurred; as we lack earlier ocean observations, we cannot determine the role of the initial conditions. We proceed with the assumption that local ice melt was

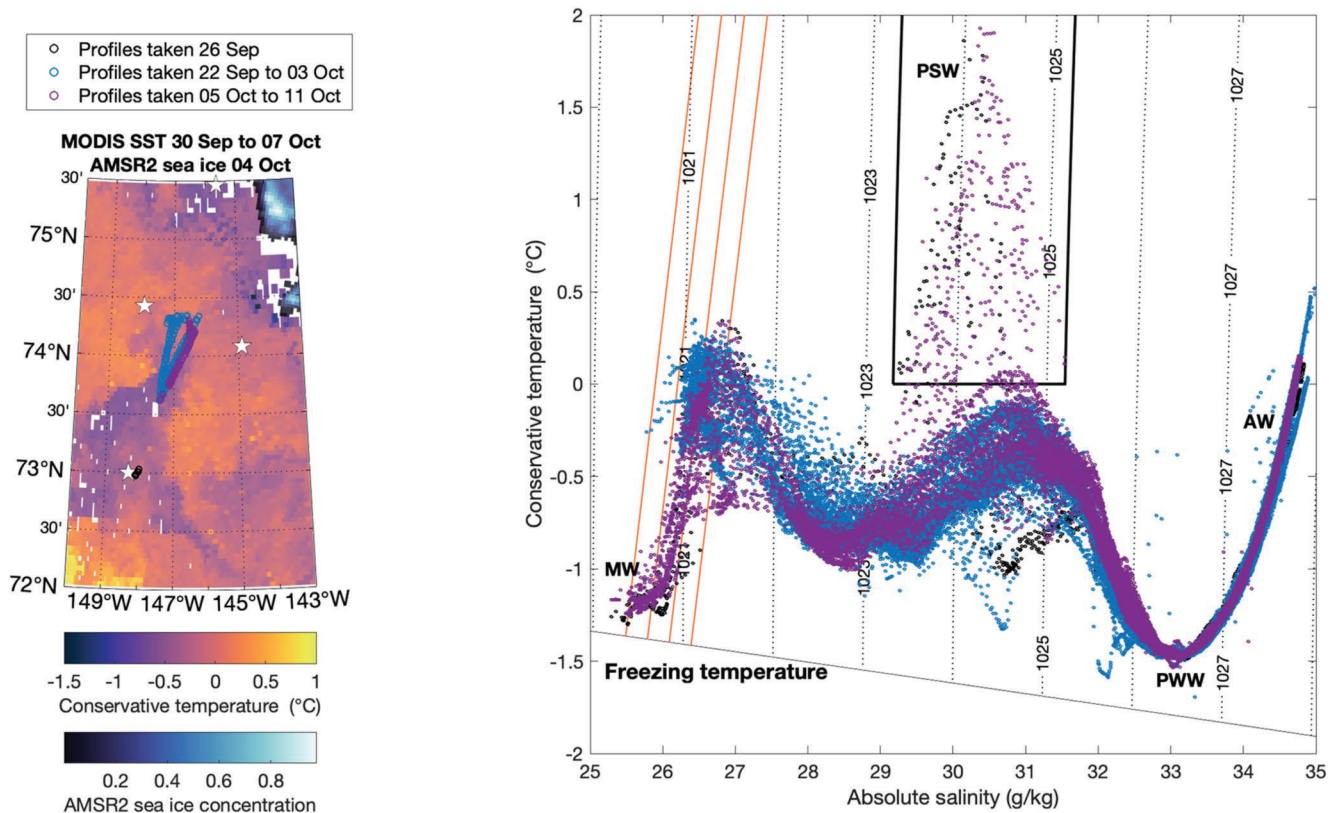


Figure 6. Conservative temperature and absolute salinity in the remnant ice area (black) and farther north before (blue) and after (purple) the meltwater arrived. (left) Colored dots show profile locations and white stars show mooring locations. (right) Seaglider profiles from 0 to 1,000 m along with potential density contours (black dotted lines) and Gade Lines (orange solid lines). The freezing temperature at the surface is indicated. MW, meltwater; PSW, Pacific Summer Water; PWW, Pacific Winter Water; AW, Atlantic Water. The black box defines PSW.

primarily responsible for the formation of the cold, fresh, and shallow surface layer, which will be called the meltwater, acknowledging that the temperature and salinity profiles suggest a minor contribution from an additional liquid freshwater source. The lateral transition from fresher to saltier surface salinities will be called the meltwater front.

Wave Gliders observed a strong salinity gradient at the edge of the meltwater front (Figure 4a) near where satellite-derived SAR imagery showed wave-like and filamentous structures at the ice edge that evolved substantially over several days (Figures 4a and 4d). Density measured at 0.25 and 9 m by a Wave Glider that crossed the meltwater front increased by about 0.5 kg/m^3 over the 10-km transect (Figure 4b). The horizontal buoyancy gradient $M^2 = \frac{\partial b}{\partial x'}$ reached $10^{-6}/\text{s}^2$ (Figure 4c) where the buoyancy $b = -\frac{g\rho}{\rho_0}$ was calculated using the reference density $\rho_0 = 1026 \text{ kg/m}^3$. M^2 was calculated in the along-track direction x' between pairs of successive Wave Glider density measurements so does not include contributions from cross-track gradients. The distance between Wave Glider measurements along this transect averaged 750 m and ranged between 590 and 890 m. The roughly six-hour cross front transit (Figures 4b and 4c) was fast enough, compared to the local inertial period, that the observations can be viewed as a snapshot of the front.

3.3. Meltwater Adveacts Northward Over Several Weeks

In the 30 September to 7 October MODIS SST image, a cold filament-like structure extended northward from the meltwater around the remnant ice (Figure 2b). Consistent with the SST imagery, Seagliders observed a colder and fresher surface layer spreading northward along their transect from 3 to 10 October (Figures 2b, 2c, 2f, and 2g). Because of this layer, mixed layer depth decreased to about 15 m (Figures 5e and 5f). Mixed layer conservative temperature and absolute salinity (conservative temperature around -1°C , absolute salinity around 25.7 g/kg) were similar to those observed earlier in the area of the remnant ice (Figure 6). Similarity in upper

ocean conservative temperature and absolute salinity properties along with the apparent coherence of the cold filament in the SST imagery strongly suggest northward advection of the meltwater. The 3–10 October Seaglider data show that the meltwater front had weakened, with M^2 reduced to 20% of that measured earlier by the Wave Glider (roughly 80 km to the south).

Satellite SST imagery was too limited by cloud cover to fully visualize the meltwater after 7 October, but the in situ observations suggest it moved northward and westward in subsequent days. The meltwater occupied the entirety of the next Seaglider transect, sampled from 6 to 11 October, indicating continued northward meltwater advection (Figures 2c and 2g). Observations by Seagliders, uCTD, and *Healy's* underway system from 11 to 14 October show that warmer and saltier waters replaced those found in the meltwater's former location, but that cold and fresh surface conditions were present further west (Figures 2d and 2h). These observations suggest westward advection of the meltwater along with a westward shift of meltwater-adjacent waters. An isolated ice patch formed on 12 October coincident with the meltwater location (Figures 2d and 2h). Heterogeneity in surface temperature and salinity persisted until the onset of freeze up, and freeze up occurred earliest in the meltwater. The mechanisms by which the meltwater hastened freeze up are discussed in Section 4.

3.4. Simulated Advection by Geostrophic and Ekman Velocities

The observed northward advection of the meltwater can be explained by a combination of geostrophic and Ekman advection. Simulated trajectories were created using geostrophic velocity (Equation 2), the average velocity in the Ekman layer (Equation 3), and the combination of the two. For each trajectory, “tracking” was initialized on 21 September near where the Wave Gliders showed the edge of the meltwater (Figure 7), then the location was updated at hourly time steps (corresponding to the times at which ERA5 surface stress was available) using the appropriate velocity. Average velocities in the Ekman layer were calculated using the ERA5 surface stress that was closest in space at the given time step and geostrophic velocity was taken as the spatially closest DOT geostrophic velocity. To explore the advection attributable to each of these processes, separate trajectories were advanced with the average Ekman layer velocity only, the geostrophic velocity only, or the sum of the two.

Ekman depth was estimated as $D_{Ek} = 10$ m to match the average mixed layer depth in the remnant ice area Seaglider profiles (Figures 4e and 4f). The theoretical Ekman depth is $D_{Ek} = \sqrt{\frac{2A}{f}}$ where A is the eddy viscosity. A typical value of $A = 0.1$ m²/s (e.g., Vallis, 2017, p. 204) at the latitude of our study area suggests an Ekman depth of about 37 m, considerably larger than the value that we use. However, Ekman layer theory assumes the ocean has a constant density, and observations indicate that stratification tends to inhibit vertical momentum flux and shallow the Ekman depth (e.g., Price et al., 1987). It is therefore reasonable to expect the Ekman depth to be less than the theoretical value in the strongly stratified meltwater we discuss.

Previous observations under sea ice in the Beaufort Sea from October to March indicate a median Ekman depth of 11 m, with most Ekman depths shallower than mixing depths (Cole et al., 2014). While our study is in ice-free waters so these results might not apply directly, they do demonstrate the shoaling effect of stratification on the Ekman depth and indicate that the mixed layer depth is not an unreasonably shallow estimate for the Ekman depth. The sensitivity of the simulated trajectories to the choice of Ekman depth is discussed shortly.

Meltwater advection can be tracked in the observations using the 26 g/kg isohaline outcrop as a proxy for the meltwater edge (triangle symbols on Figures 2 and 7). Ongoing upper ocean cooling altered surface density slightly, motivating the use of an isohaline rather than an isopycnal. The simulated trajectories are evaluated by their ability to reproduce the observed advection.

The ocean geostrophic velocity calculated from DOT showed cyclonic circulation with a length scale of ~100 km located north of the remnant ice area (Figure 7). This geostrophic circulation agrees with the sloping isopycnals below the mixed layer seen in the southern half of the first Seaglider transect (Figures 5b and 5c), which crossed the northern part of the cyclonic circulation seen in the DOT. This observed isopycnal slope also agrees qualitatively with the large-scale *anticyclonic* circulation of the Beaufort Gyre, but the substantial weakening of this subsurface meridional density gradient between Seaglider transects (compare Figures 5b, 5c, and 5e, 5f) suggests that it was attributable in part to transient circulation like the cyclonic circulation derived from the DOT.

Trajectories calculated from the geostrophic velocity alone followed the cyclonic circulation northward but fell short of the observed extent of northward advection (not shown). Although the full DOT survey of the study area

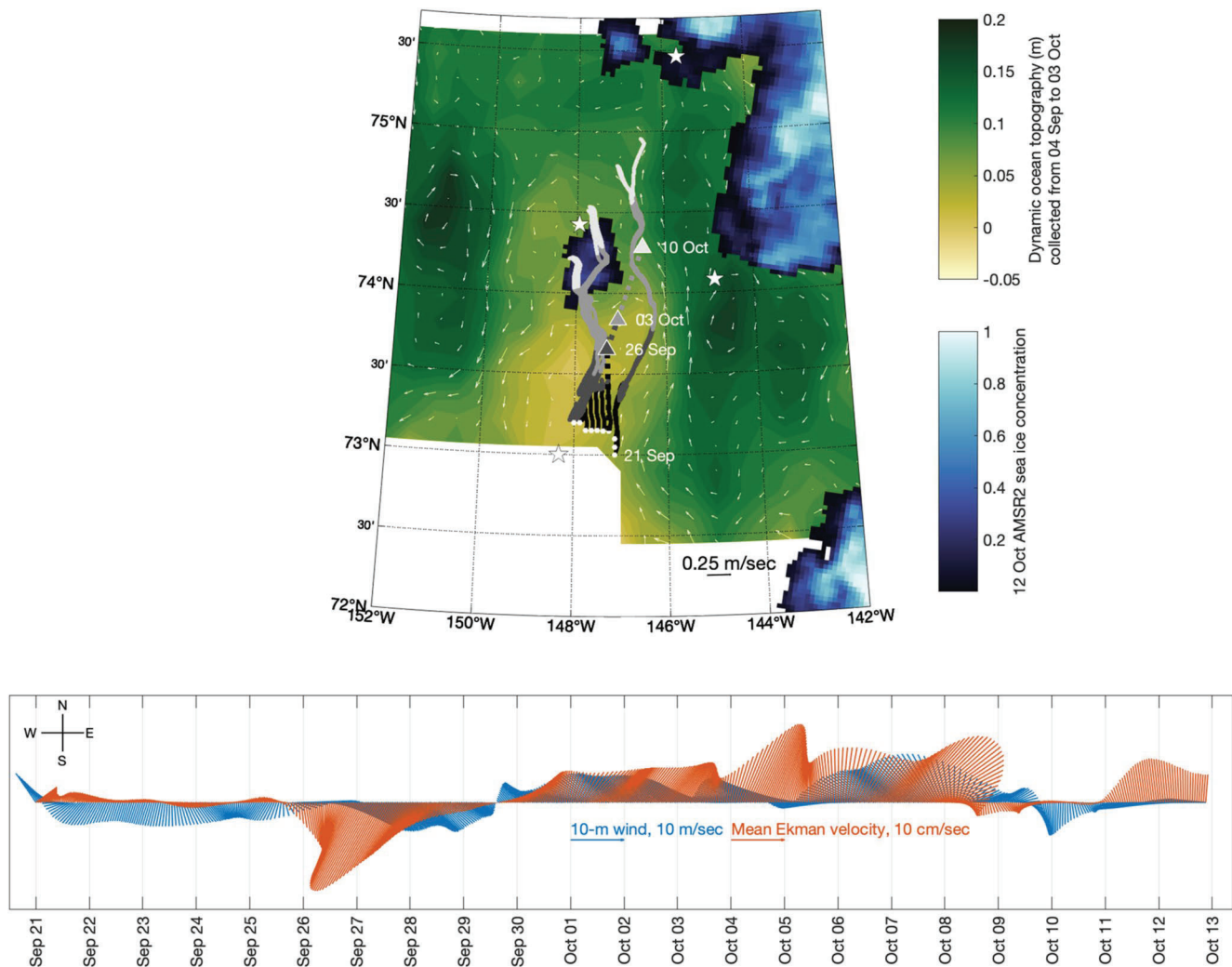


Figure 7. Simulated advection trajectories driven by geostrophic and Ekman velocities compared to the observed meltwater advection. (top) Dynamic ocean topography (background colors) and surface geostrophic velocity (white arrows) along with the simulated trajectories advected by the combined geostrophic and Ekman velocities from 21 September (white dots) to 15 October (ends of solid tracks). The locations of the advected tracks at key times (track color changes) correspond to the times of the observed locations of the 26 g/kg isohaline outcrop (colored triangles), which are connected to illustrate the inferred advection path (dotted line). Stars indicate mooring locations. (bottom) ERA5 10-m wind and associated Ekman velocities averaged in the region bounded by 145°–149°W, 73°–74.5°N. The length of the lines shows the speed, with scaling lines shown below the time series. The direction of the lines shows the cardinal direction of the velocity, with a compass rose provided for reference.

took more than a month and consisted of a series of parallel, roughly meridional satellite tracks, the orientation of these tracks meant that the zonal DOT gradients were captured on time scales of only a few days, with the region north of the meltwater surveyed from 16 to 22 September. The shorter time scale needed to resolve the zonal gradients than the meridional gradients along with the fact that the meltwater area was surveyed shortly before the observed northward advection of the meltwater began give us relatively more confidence in the northward geostrophic circulation responsible for northward meltwater advection than in other aspects of the calculated geostrophic velocity field.

Ekman transport advected the meltwater northward, but Ekman-only advection was also insufficient to match the observed northward advection (not shown). For a 10-m Ekman depth, simulated Ekman trajectories from 21 September to 3 October went southwestward and then northeastward back to near their original locations. From 3 to 11 October, Ekman trajectories extended 50 km to the northeast (regionally averaged Ekman velocity shown in Figure 7).

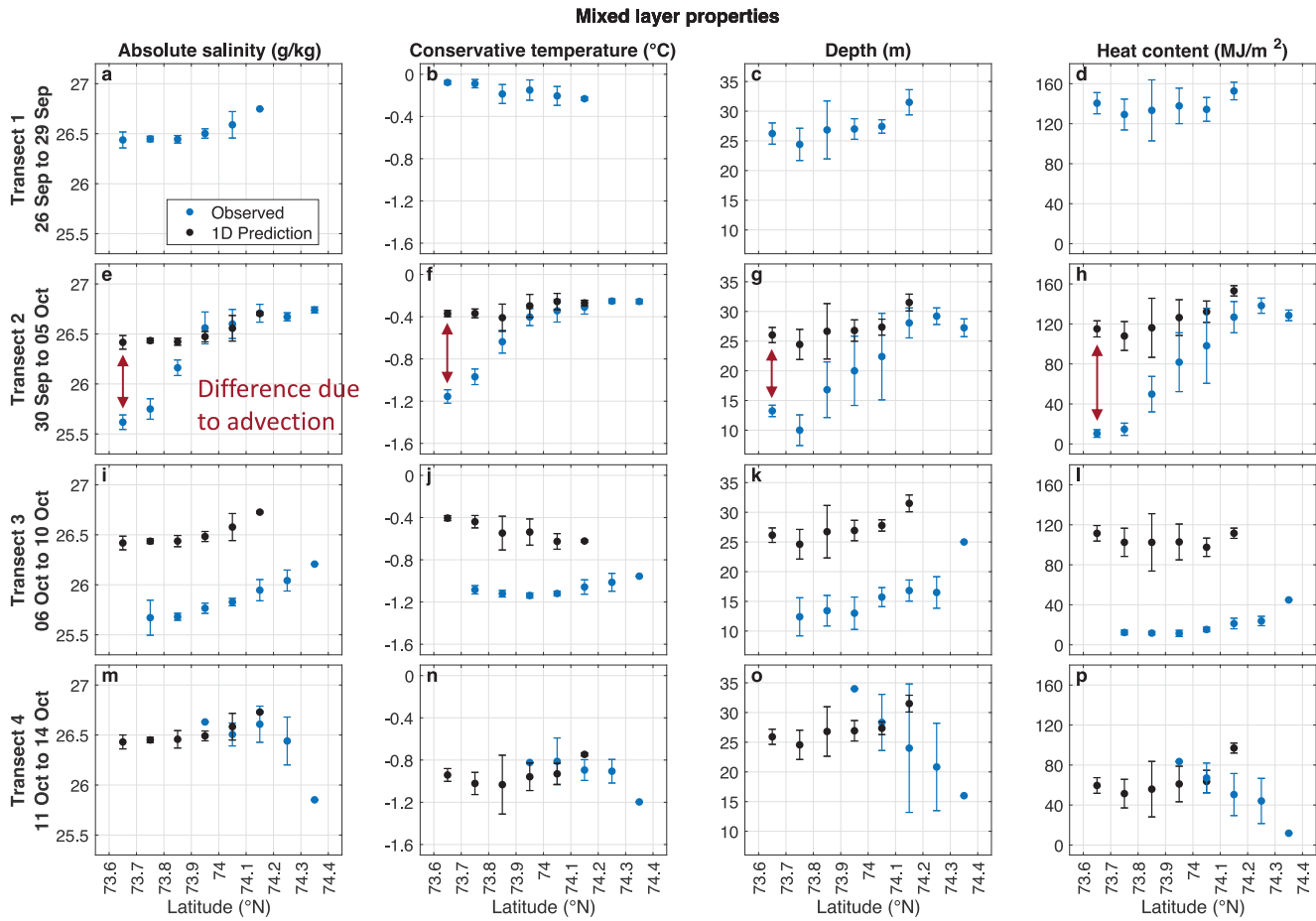


Figure 8. Differences between observations and model predictions demonstrate that advection altered mixed layer properties. Observed (blue) and concurrent Price-Weller-Pinkel (PWP) model-predicted (black) mixed layer properties binned by latitude and averaged. Observation locations are shown in Figures 2e–2h. Each row shows mixed layer properties on the transect labeled to the left while each column shows the property labeled at the top. Meltwater presence can be inferred from observed absolute salinities below 26 g/kg. There are no model predictions for Transect 1 as these profiles were used to initiate the model. The error bars show one standard deviation.

Neither Ekman-only nor geostrophic-only advection was sufficient to explain the filament's northward advection, but the simulated trajectories driven by the combined velocities did transport surface water far enough to be in coarse agreement with the observations (Figure 7). Additionally, the more western simulated trajectories, initiated in the core of the meltwater, aligned well with the early ice formation patch.

The average mixing layer depth in the remnant ice area Seaglider profiles was 6 m (whereas the average *mixed* layer depth used as the Ekman depth previously was 10 m). When tracking was initiated in the same start positions as before and advected using the combined velocities calculated with a 6-m Ekman depth, most trajectory endpoints clustered in the same region as in the 10-m Ekman depth simulation though the easternmost trajectories terminated somewhat further north (not shown). Even with a thin, 6-m Ekman layer, the trajectories advected by Ekman velocity only did not travel far enough north to match the observations, affirming that the geostrophic component of the circulation was necessary to explain the observed meltwater advection (not shown). The average mixed layer depth increased to 14 m on a later Seaglider transect through the meltwater (Figure 8k), indicating that the Ekman depth within the meltwater could also have increased with time. When tracking was initiated in the same start positions as before and advected using the combined velocities calculated with a 15-m Ekman depth, some trajectories terminated between 146°–148°W and 74°–75°N, in qualitative agreement with the observations and with the 10-m Ekman depth results (not shown). The main difference from the 10-m Ekman depth results was that some of the western trajectories were more strongly affected by the geostrophic circulation and were advected southward after 3 October.

These simple surface current advection simulations are not meant to stand in for a fully coupled three-dimensional ocean model, which could be used to better understand the evolution of the meltwater. For example, our use of the vertically averaged velocity in the Ekman layer neglects the expected Ekman velocity spiral which could shear meltwater over ambient water at the edges of the meltwater patch. The simulations do however demonstrate that, within a range of reasonable Ekman depths, the surface currents can explain the observed northward advection of the meltwater.

4. Mixed Layer Heat Budget

4.1. Overview

Ocean observations described in the previous section show a co-occurrence of meltwater presence and early sea ice advance seen by passive microwave satellites. This co-occurrence suggests meltwater advection influenced the timing of freeze up. This section quantifies that link, demonstrating that decreases in mixed layer temperature and mixed layer depth due to meltwater advection hastened freeze up by several days. Additionally, a mixed layer heat budget shows that, in the area affected by the meltwater, advection was a leading cause of mixed layer heat loss.

The mixed layer heat budget was developed for a subset of the study area, along a transect that was affected by meltwater and was repeatedly sampled by Seagliders (see pink highlighting in Figures 2e–2h). Focusing on a repeatedly sampled transect eliminates differences attributable to pre-existing heterogeneity in upper ocean structure. We name the four Seaglider occupations Transects 1–4, described next. Transect 1, sampled from 26 to 29 September, was completed before the meltwater entered the study area; the mixed layer conditions observed on this first transect are the initial conditions against which subsequent observations are compared (Figures 8a–8d). Transect 2, sampled from 30 September to 5 October, crossed into the meltwater, with meltwater present at the southern end of the transect but absent from the northern end (Figures 8e–8h). Meltwater filled Transect 3, sampled from 6 to 10 October (Figures 8i–8l). Transect 4, sampled from 11 to 14 October, did not contain meltwater (Figures 8m–8p).

Changes in mixed layer heat content are due to horizontal and vertical processes. Prior to freezing, vertical (or one-dimensional) processes include net surface heat exchanges with the atmosphere as well as heat exchanges with the underlying pycnocline. Horizontal processes include horizontal advection of colder (or warmer) waters, directly bringing the mixed layer closer to (or further from) its freezing temperature, as well as horizontal advection of waters with different salinities, which impacts the mixed layer depth and stratification.

The definition of mixed layer heat content (Equation 1) means the mixed layer heat content goes to zero as the mixed layer temperature approaches the freezing temperature. Once the mixed layer has lost its heat, frazil ice begins forming if further heat is lost to the atmosphere. Thus, the freeze-up process initiates when the mixed layer reaches the freezing temperature, which will be referred to as the onset of freeze up. If atmospheric and oceanic conditions remain favorable for ice growth, sea ice will mature to a stage at which it is visible in remote sensing. This growth period requires time, so the onset of freeze up defined by freezing temperature mixed layers is not the same as the freeze up timing inferred from satellite products (i.e., in Figure 1).

Our goal is therefore to create mixed layer heat budgets in the presence and absence of meltwater to quantify the importance of meltwater advection to mixed layer heat loss integrated throughout the autumn season.

4.2. Heat Budget Development and Results

As described in Section 4.1, the mixed layer heat budget contains three terms: net surface heat flux, which we infer from ERA5, heat exchange with the pycnocline, which we estimate using the PWP model, and advection. Heat exchange between the pycnocline and the mixed layer driven by entrainment due to mixed layer deepening, and by turbulent diffusion, are combined into a bulk pycnocline heat exchange. The differences between observed changes in mixed layer heat content, calculated in an Eulerian reference frame using the repeat transects, and changes simulated by the one-dimensional PWP model, which excludes advection, are taken as an estimate of horizontal advection.

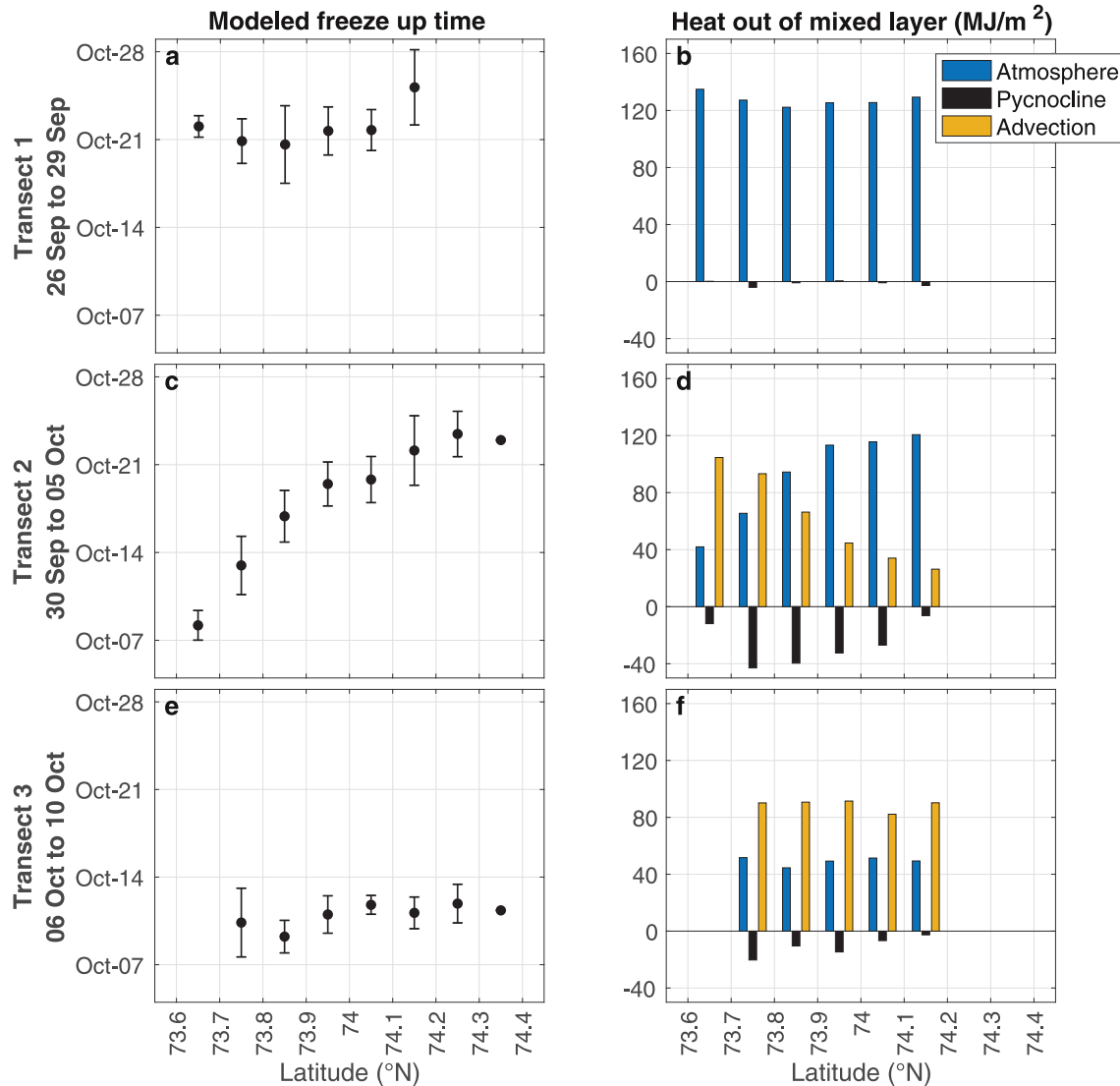


Figure 9. Freeze up times and mixed layer heat budgets varied with the presence or absence of meltwater. Observation locations are shown in Figures 2e–2h. (a, c, e) Price-Weller-Pinkel (PWP) model-predicted freeze up binned and averaged by latitude. The error bars are one standard deviation. (b, d, f) Heat budget components for each latitude bin on each transect.

For the Transect 1 heat budget, the PWP model was initialized with observed conservative temperature and absolute salinity profiles collected on the transect. Separate model runs for each profile were forced with ERA5 heat flux, surface stress, and surface freshwater flux, and run until the modeled mixed layer reached the freezing temperature, interpreted as the modeled onset of freeze up. Mixed layer properties observed on the meltwater-free Transect 4, which was sampled after ambient water replaced meltwater on the transect (Figures 2d and 2h), agreed with properties predicted by the PWP model runs initiated on the meltwater-free Transect 1 (Figures 8m–8p). Agreement between observations and concurrent model predictions demonstrates that one-dimensional processes adequately described mixed layer evolution outside of the meltwater and indicates that advection was not an important component of the mixed layer heat budget for Transect 1. The modeled onset of freeze up for Transect 1 on average occurred on 21 October (Figure 9a). This modeled freeze up onset is comparable to the satellite-observed 18 October ice advance southwest of the meltwater (Figure 1). Differences in the definition of freeze up would be expected to drive differences in timing between these estimates, but their rough agreement supports the validity of the modeled mixed layer dynamics.

The PWP model results for Transect 1 profiles were used to create a heat budget representative of mixed layer evolution without meltwater advection. Net surface heat loss for each profile was estimated by integrating the ERA5 heat flux from the time of the initial observation to the time of the modeled onset of freeze up. Time-integrated mixed layer heat exchange with the pycnocline was calculated as the change in pycnocline heat content over the same time period. The pycnocline heat content was the depth integrated heat content between the base of the mixed layer and 50 m, the deepest depth simulated in the PWP model. As PWP does not account for advection, the *change* in modeled pycnocline heat content was due almost entirely to heat exchange with the mixed layer. Diffusion altered the temperature profile within the pycnocline somewhat, meaning the choice of lower integration bound could influence the modeled pycnocline heat content if heat diffused into or out of the integration volume from below, but as diffusion was intentionally small in our model runs and we integrated the heat content of the entire modeled pycnocline, this effect is not important to the calculated change in pycnocline heat content.

Average surface heat fluxes throughout the PWP model runs were between -45 and -65 W/m^2 , with positive fluxes directed downward and higher values for later-freezing profiles. This pattern makes sense because atmospheric heat loss accelerated as autumn progressed, so later-freezing profiles were exposed to larger, late-season heat fluxes for longer periods of time. The largest surface heat flux during our study period was around -150 W/m^2 and occurred on 7 October.

Mixed layer heat content changes on Transect 1 were almost entirely due to heat loss to the atmosphere, which, on average, totaled 128 MJ/m^2 between the initial observations and modeled freeze up (Figure 9b). Heat gained by the mixed layer from the pycnocline averaged 1.4 MJ/m^2 (Figure 9b), small compared to heat loss to the atmosphere. Agreement between model predictions and later observations outside of the meltwater indicates that advection was negligible for Transect 1 profiles. Transect 1 results can be thought of as simulating mixed layer evolution to the onset of freeze up in the absence of meltwater advection, and therefore approximating the mixed layer heat budget for waters adjacent to the meltwater, which were not sampled directly.

Mixed layer properties observed on the meltwater-containing Transects 2 and 3 deviated substantially from those predicted by PWP simulations initiated on the meltwater-free Transect 1. Transect 2 crossed into the meltwater and thus included a mix of profiles with and without meltwater; we focus on profiles that contained meltwater as we assume that the previously discussed Transect 1 results are representative of meltwater-absent conditions. The observed meltwater mixed layer (defined to include observed profiles with surface absolute salinity below 26 g/kg) was about 0.55 $^{\circ}\text{C}$ colder and 13 m shallower than the model predictions (Figures 8e–8l). These differences indicate a strong role for advection in setting the mixed layer properties observed on Transects 2 and 3. For model runs initiated with meltwater-containing Transect 2 and 3 observations, the modeled onset of freeze up on average occurred on 11 October (Figures 9c and 9e). This timing agrees with the satellite-observed ice patch that formed on 12 October adjacent to the previously observed meltwater (Figures 2d and 2h). The agreement in satellite-observed and modeled freeze up timing in the meltwater implies that while the meltwater's presence was due to advection, the evolution within the meltwater was one-dimensional.

Developing a meltwater-influenced heat budget using observations on Transects 2 and 3 requires two additional steps beyond those followed when meltwater was absent. The first problem is that Transects 2 and 3 were sampled later than Transect 1. It is the time-integrated loss of heat from the mixed layer throughout the autumn cooling season, rather than instantaneous heat fluxes into or out of the mixed layer, that controls the onset of freeze up. Thus, it is important that the integrals of mixed layer heat fluxes for meltwater-present and meltwater-absent heat budgets begin at roughly the same time.

To extend the Transect 2 and 3 heat budgets, we used Transect 1 model results to estimate the vertical heat budget terms integrated from the time Transect 1 was sampled to the times when Transects 2 and 3 were sampled. In other words, Transect 1 model results, run until the time when observations were collected on Transects 2 and 3, were added to the vertical heat budget terms derived from model runs initiated on those transects and run until the modeled onset of freeze up. The vertical heat budget terms for Transects 2 and 3 are thus the sum of two sets of model runs spanning two different time periods: Transect 1 model results prior to the Transect 2 or 3 observations, and additional model runs initiated with the Transect 2 or 3 observations and run until the modeled onset of freeze up. Heat budget terms were combined by averaging Transect 1 model results in latitude bins and then adding them to Transect 2 and 3 model results binned in the same manner.

The second difference between the meltwater-absent Transect 1 heat budget and the heat budgets for Transects 2 and 3 is that meltwater advection played a major role in mixed layer heat loss for the later observations. We assume that differences between observed mixed layer properties and concurrent modeled mixed layer properties can be attributed to horizontal advection. Therefore, the advective component of the mixed layer heat budget for Transects 2 and 3 was estimated as the difference between the observed mixed layer heat content and the concurrently predicted mixed layer heat content from PWP model runs initiated on Transect 1 (see Figures 8h, 8l, and 8p).

Heat loss to the atmosphere for meltwater-containing profiles on Transects 2 and 3 was about 50 MJ/m² (Figures 9d and 9f). Heat gained from the pycnocline was variable, with maximum values of up to 44 MJ/m² (Figure 9d). The larger pycnocline heat exchange within the meltwater relative to outside the meltwater was likely due to increased mixing caused by shear instabilities at the base of the meltwater mixed layer, whose shallower depth resulted in larger acceleration for a given surface stress. The differences between observed and concurrent modeled mixed layer heat content indicated an advective mixed layer heat loss of about 90 MJ/m² (Figures 9d and 9f).

Our observations do not span the entire autumn cooling period. Net surface heat flux in ERA5 became net negative around 10 September, and about 50 MJ/m² of ocean heat was lost to the atmosphere before Seaglider sampling began (not shown). Lacking ocean observations, we cannot estimate pycnocline and advective heat budget components during this intervening period. However, accounting for this earlier ocean cooling implies seasonally integrated heat loss to the atmosphere (50 MJ/m² prior to the earliest observations + 50 MJ/m² between observations and freeze up in the meltwater) was comparable to the mixed layer heat loss associated with meltwater advection (90 MJ/m²). This demonstrates that meltwater advection played a leading role in mixed layer heat loss integrated throughout the autumn. It also shows that, despite relatively higher values within the meltwater, heat exchanges with the pycnocline were of secondary importance to the mixed layer heat budget compared to heat loss to the atmosphere and advection.

5. Additional Advection: Restratification Hastens Freeze up in Northern Study Area

This section shifts focus to the northern part of the study area and briefly discusses a freshwater advection event separate from the meltwater advection described previously. Freeze up swept across the northern study area on 14 October (Figure 1), which was several days earlier than predicted by PWP model runs initiated by observations in the area on or before 7 October. Though not recreated in PWP model runs, shoaling mixing depth, which would confine heat loss to a smaller portion of the water column, offers a possible explanation. Consistent with this mechanism, Seaglider observations from 12 to 15 October show that the upper 15 m of the water column freshened, decreasing the average surface density by 0.1 kg/m³ compared to the earlier profiles (see the example profiles in Figure 10). PWP runs initiated with these later profiles had shallower cooling and an average onset of freeze up of 15 October, agreeing with the satellite-derived freeze up timing.

The observed restratification was likely due to lateral processes as it was not simulated by the one-dimensional PWP model forced by ERA5 surface heat and freshwater fluxes. The area was ice-free when restratification was observed (Figure 10), so local ice melt also cannot explain the surface freshening. Lateral processes could act to spread buoyant meltwater and thus create shallow stratification. The model results suggest that this restratification accelerated freeze up.

6. Summary and Discussion

This study demonstrates how freshwater advection alters upper ocean stratification and leads to spatial variability in freeze up timing. Nearly 800 ocean conservative temperature and absolute salinity profiles characterize the open water region in the central Beaufort Sea, immediately prior to freeze up in 2018, at high spatial and temporal resolution. Few previous studies capture ice-free Arctic Ocean conditions at these small scales during this critical time of year due to the risks to instrumentation when freeze up is imminent or in progress. The observed oceanic modulation of freeze up means forecasting efforts should consider the three-dimensional evolution of the mixed layer and its impacts on sea ice development.

In one case of freshwater advection, a cold and fresh surface meltwater layer caused by melting remnant ice in the southern Beaufort Sea advected roughly 100 km in several weeks. Meltwater hastened the onset of freeze up by

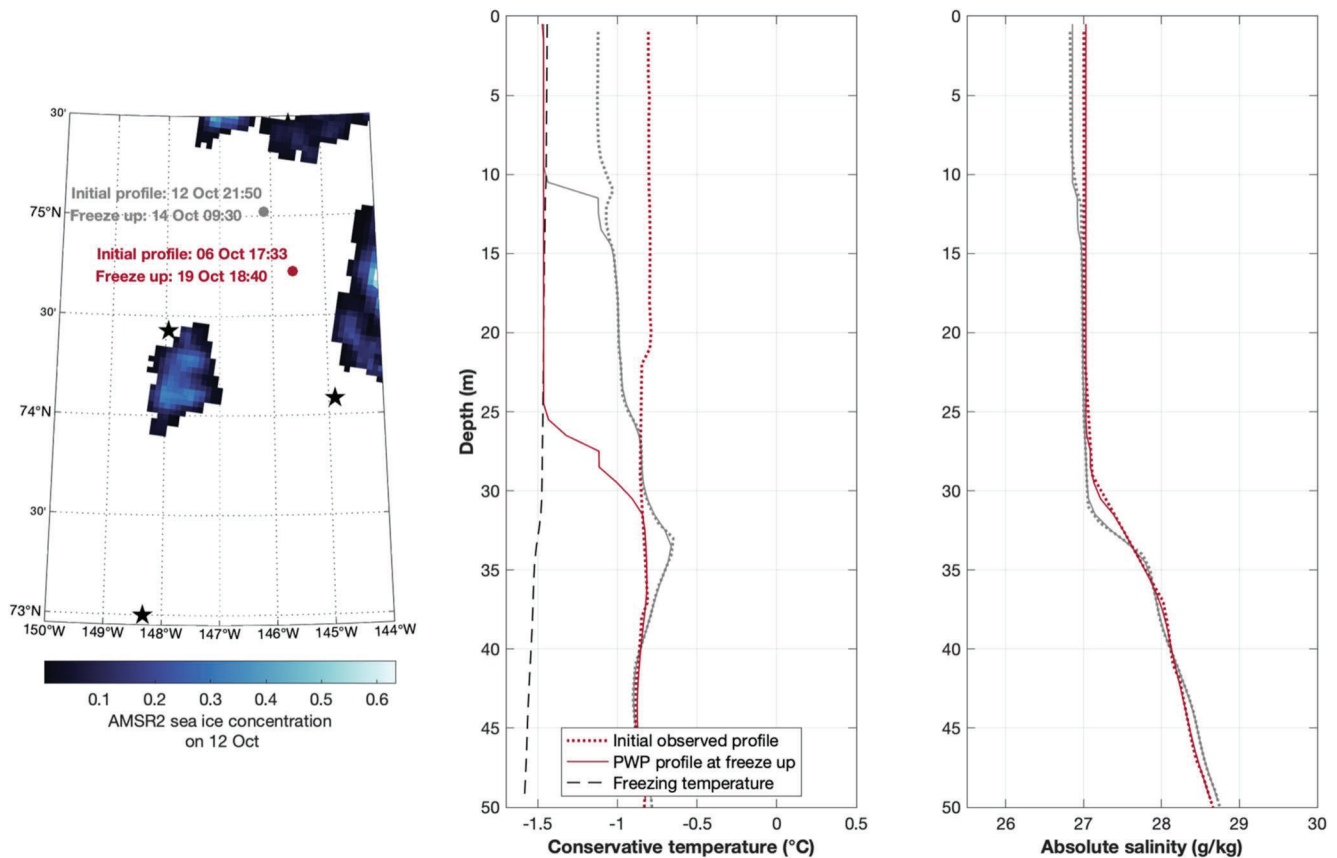


Figure 10. Surface freshening near the northern marginal ice zone (MIZ) restricted cooling to a shallower layer and resulted in earlier modeled freeze up. (left) Locations of example observed profiles used to initiate the Price-Weller-Pinkel (PWP) model along with AMSR2 ice concentration where it exceeded zero. Stars indicate mooring locations. (right) Observed conservative temperature and absolute salinity profiles (dotted lines) and modeled profiles at freeze up (solid lines). The profile colors correspond to the colored observation locations and times shown on the map.

cooling and shoaling the mixed layer, with the result that ice advanced in meltwater-affected areas several days earlier than in surrounding areas where meltwater was absent. Mixed layer shoaling limited the volume of surface water that had to cool to the freezing temperature before ice could form. Outside of the meltwater the weaker stratification distributed cooling over a deeper mixed layer and delayed ice formation.

The mixed layer heat budget for the autumn cooling period showed that, outside of the meltwater, regional-scale heat loss to the atmosphere was the dominant means of mixed layer cooling. Within the area affected by meltwater, advection was nearly as important as heat loss to the atmosphere for seasonally integrated mixed layer heat loss. Heat gained by the mixed layer from the pycnocline was of secondary importance.

While previous research concluded that advection plays a minor role in Beaufort Sea upper ocean freshening on seasonal time scales (Dewey et al., 2017), we note that the meltwater front documented in our study would be indiscernible in the SIZRS data set used in Dewey et al. (2017) which has temporal resolution of 1 month and spatial resolution of 1° of latitude. While determining the frequency of such advective events is outside the scope of this work, we note that a second filament of cold surface water emanating from the remnant ice area, unsampled in our study, was visible in the satellite SST imagery (Figure 2b). To the extent that anomalously cold surface temperatures were a tracer for meltwater, this SST pattern suggests that meltwater advection on $O(100\text{ km})$ scales may be an important source of upper ocean heterogeneity.

A separate freshwater advection event occurred near the MIZ when the mixing layer freshened by 0.1 g/kg in the days preceding freeze up. Our observations do not capture the origin of this freshwater or the exact timing of the restratification; however, the mixed layer model's inability to replicate the surface freshening suggests it was not due to one-dimensional surface forcing, and satellite imagery shows that it was not due to local ice melt. We thus

attribute the restratification to a lateral flow of freshwater. Our measurements do not resolve the ocean structure associated with the northern MIZ well enough to determine the dynamics contributing to the observed restratification; advection by Ekman, geostrophic, or submesoscale flows are all candidates.

Mixed layer instability associated with the nearby ice edge is one possible explanation for the restratification observed near the northern MIZ, as modeling studies of submesoscale meltwater fronts show that mixed layer eddies can spread meltwater laterally (Horvat et al., 2016; Manucharyan & Thompson, 2017; Manucharyan & Timmermans, 2013). However, the available potential energy is lower in fronts with shallower meltwater layers, and shallow meltwater layers take longer to develop eddies so their dynamics are more likely to be influenced by wind and surface buoyancy forcing (Manucharyan & Thompson, 2017). Timmermans et al. (2012) found that lateral restratification was common in profiles beneath sea ice and noted that the slight density increases below actively mixing layers isolated the rest of the underlying mixed layer and prevented pycnocline entrainment (Timmermans et al., 2012). We add to their results by showing that diverging mixing and mixed layers precondition freeze up, demonstrating a mechanism by which three-dimensional ocean circulation allows freeze up to advance more quickly than one-dimensional ocean cooling would predict.

While this study has focused on the mixed layer, the Seaglider profiles also captured the northward advection of Pacific Summer Water (PSW, defined by MacKinnon et al. (2021) as water bounded by the 1023.2 and 1025.2 kg/m³ isopycnals and warmer than 0°C) that was nearly 2°C (Figure 6). These PSW warm features were present in some Seaglider profiles collected in the remnant ice area on 26 September (Figures 4e and 6) and then similarly warm PSW was observed in *O*(20 km) boluses on the Seaglider transect north of the remnant ice area beginning on 5 October (not shown), arriving roughly in tandem with the meltwater (Figure 6). The top of the PSW was found at similar salinities at both locations but isopycnals deepened with latitude resulting in a roughly 20 m downward displacement of the PSW as it advected northward (not shown). The later northern profiles showed exceptionally warm PSW at deeper salinities as well (down to $S_A = 31$ g/kg, Figure 6) suggesting downward mixing had occurred. MacKinnon et al. (2021) described a very warm (up to 6°C) PSW jet impinging from the south on remnant sea ice (the same remnant sea ice discussed in this study) and breaking into intrathermocline eddies when observed from 14 to 17 September 2018. Our observations likely represent the continued subduction and advection of this PSW jet.

The stratification below the meltwater stored a relatively large amount of heat in an NSTM when ice formed. The modeled upper ocean heat content from 0 to 40 m at freeze up for profiles in the meltwater was 100 MJ/m² on average while the modeled upper ocean heat content at freeze up for profiles outside the meltwater was 67 MJ/m² on average. The 40 m heat content integration bound was chosen to represent a typical winter mixed layer depth in this region as Peralta-Ferriz and Woodgate (2015) found Canada Basin winter mixed layer depths averaged 33 m using a more restrictive density change threshold ($\Delta\rho = 0.1$ kg/m³ relative to the surface) than used here ($\Delta\rho = 0.25$ kg/m³). The fate of heat entrained from the NSTM by convection or wind-driven mixing depends on the maturity of the overlying ice. Most heat reaching the base of thick, insulating ice is expected to slow the growth of that ice (Timmermans, 2015), whereas Smith et al. (2018) show that a substantial fraction of the heat released by a storm early in the freeze-up process escaped directly to the atmosphere with only 30–40% going toward melting the thin pancake ice. In addition, downwelling due to Ekman pumping can cause the NSTM to deepen throughout the autumn and winter (Jackson et al., 2010). If surface mixing is relatively weak, this downwelling may be strong enough to sequester the NSTM below the mixed layer such that it persists through the winter (Steele et al., 2011).

While it is not possible with our data set to differentiate among these outcomes for the additional heat stored in the meltwater-present NSTM relative to the meltwater-absent NSTM, the larger NSTM heat content at freeze up below the meltwater is nonetheless an important result. At first glance, earlier ice formation in the meltwater might be expected to result in thicker ice at the end of winter due to a longer ice growth season. However, entrainment of the stronger (warmer and shallower) NSTM below the meltwater could counteract the “head-start” of earlier ice formation and inhibit seasonally integrated ice growth. Ice that thins during winter due to NSTM entrainment retreats earlier the following year, allowing the ocean to absorb more solar radiation the following summer (Jackson et al., 2012). The onset of the spring ice algal bloom is light limited (Leu et al., 2015), so earlier ice retreat also impacts the ice-based ecosystem.

Arctic sea ice forecasting is an area of substantial ongoing research (e.g., Guemas et al., 2016, and references therein), and the importance to stakeholders of predicting freeze up on regional spatial scales has recently been highlighted (Bushuk et al., 2019; Day et al., 2014; Stroeve et al., 2016). Positive sea ice area anomalies during the melt season result in cold SST anomalies which can persist throughout the summer, meaning the ocean retains a “memory” of previous ice conditions which preconditions the re-emergence of positive ice area anomalies during freeze up (Blanchard-Griggs et al., 2011; Day et al., 2014). This re-emergence mechanism means that the timing of ice retreat can be used to statistically predict the timing of ice advance in some parts of the Arctic, with moderate predictive skill in the Beaufort Sea (Stroeve et al., 2016). Modeling studies also show that the inclusion of SST information contributes to skillful freeze up forecasting at socioeconomically relevant spatial scales (Bushuk et al., 2019; Sigmond et al., 2016). Our results advance this forecasting paradigm by demonstrating that (a) SST anomalies associated with late ice retreat advect and precondition freeze up in neighboring areas that did not themselves experience anomalously late ice melt and (b) late ice melt can create persistent, anomalously shallow mixed layer depths that also promote early freeze up.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The ocean observations presented in this paper are available on the University of Washington Research Works archive at <http://hdl.handle.net/1773/47135>. ERA5 hourly data on single levels were downloaded from the C3S climate data store (<https://cds.climate.copernicus.eu#!/home>). The link to the reanalysis on surface/single levels is <https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels?tab=overview>. Sea ice concentration from AMSR2 was downloaded from the University of Bremen data archive at https://seaice.uni-bremen.de/data/amr2/asi_daygrid_swath/n3125/2018/. Level 3 Daily and 8-Day average MODIS-Terra sea surface temperature data were downloaded from the Physical Oceanography Distributed Active Archive Center at NASA/JPL from <https://podaac-tools.jpl.nasa.gov/drive/files/allData/modis/L3/terra/11um/v2019.0/4km>. Dynamic ocean topography (DOT) from CryoSat-2 was downloaded from <http://rads.tudelft.nl/rads/data/authentication.cgi>. Level 3 v1.1 weekly SMOS ARCTIC sea surface salinity data were downloaded from SEANOE at <https://www.seanoe.org/data/00607/71909/>. Satellite SAR ice images and analysis were provided via special support from the U.S. National Ice Center. RADARSAT-2 Data and Products are under a copyright of MDA Geospatial Services Inc. 2018—All Rights Reserved, obtained via the U.S. National Ice Center. RADARSAT is an official mark of the Canadian Space Agency. Sentinel-1 data were obtained from the Copernicus Data Hub, supported by the European Space Agency.

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1 **Sea ice leads increase ice and ocean inertial oscillation strength**
2 **during an Arctic winter storm**

3 Laura Crews,^a Craig M. Lee,^a Luc Rainville,^a Samuel Brenner,^b

4 ^a*Applied Physics Laboratory, University of Washington, Seattle, WA, USA*

5 ^b*Department of Earth, Environmental, and Planetary Sciences, Brown University, Providence,*
6 *RI, USA*

7 *Corresponding author: Laura Crews, lcrews@uw.edu*

8 ABSTRACT: Ocean mixed layer inertial oscillations are energetic currents that link energy input
9 by the wind to ocean mixing and dissipation. Sea ice properties and dynamics play a central
10 role in atmosphere-ocean momentum transfer in polar regions, and sea ice changes are altering
11 many aspects of the Arctic system, including ocean near-inertial currents. This study uses moored
12 velocity observations and remote sensing collected during a February 2019 storm in the Beaufort
13 Sea to demonstrate how sea ice lead opening permitted inertial oscillation development in the
14 sea ice and ocean surface mixed layer despite persistently high ice concentrations, whereas an
15 unfractured ice pack selectively filtered out near-inertial motions. At the northern mooring site,
16 which experienced minimal lead opening, internal ice stresses reduced inertial oscillation strength
17 by 82% relative to a slab model driven by the surface forcing. At the southern mooring site, in
18 contrast, lead opening relieved internal ice stresses and allowed relatively uninhibited ocean inertial
19 oscillation development. Our results suggest that near-inertial motion predictions in consolidated
20 ice packs should account for the lateral scales of ice fracturing and cannot rely on free-drift proxies
21 like ice concentration and wind factor. While inertial oscillations in the ice-free ocean are locally
22 driven, this study connects the local inertial response to regional ice dynamics, including the ice
23 pack's interaction with the distant coast.

1. Introduction

In polar regions, sea ice is fundamental to momentum transfer between the atmosphere and the ocean (e.g., McPhee 2012). The mechanical ice-ice interactions that comprise internal ice stresses, such as crushing during floe collisions, can interrupt this momentum transfer. Local internal stresses are not simply due to local ice deformation; instead, consolidated ice can transfer internal stresses from ice-coast interactions hundreds of kilometers into the ice pack (Richter-Menge et al. 2002) decreasing ice mobility and generating long and narrow cracks known as leads (e.g., Lewis and Hutchings 2019; Jewell et al. 2023).

Fluctuating wind stresses excite energetic currents in the ocean surface layer that oscillate at frequencies near the local inertial frequency f (Pollard and Millard 1970). These inertial oscillations can be strong enough to initiate shear-driven mixing and pycnocline entrainment at the mixed layer base (Plueddemann and Farrar 2006), and may propagate into the pycnocline as near-inertial internal waves (D’Asaro 1989). Accurately predicting or modeling near-inertial energy input to the ocean mixed layer is an essential step toward understanding mixed layer dynamics as well as deep ocean circulation, stratification, and mixing (e.g., Alford 2001; Wunsch and Ferrari 2004).

While simple models adequately describe the ice-free ocean’s response to surface wind stress (Pollard and Millard 1970; D’Asaro 1985), sea ice complicates this response. The sea ice and the ocean mixed layer can undergo tightly coupled inertial oscillations (Brenner et al. 2023a; McPhee 1978), but these oscillations are generally weaker during ice-covered periods than during open-water periods. This seasonality has often been attributed to damping by a strong and consolidated ice pack in which internal stresses are important (Kawaguchi et al. 2019; Martini et al. 2014; Rainville and Woodgate 2009; Gimbert et al. 2012b) but seasonal changes in mixed layer depth also may contribute (Brenner et al. 2023a).

Deviations from this ice-influenced inertial oscillation seasonality represent opportunities to explore how changes in ice dynamics or characteristics impact near-inertial motions. For example, despite high ice concentrations, the most energetic ocean inertial oscillation event observed throughout a year-long mooring deployment on the Beaufort slope occurred in December (Martini et al. 2014). This event was “preconditioned” by another wind event several days earlier that “loosen[ed]” the ice pack (Martini et al. 2014), demonstrating that strong and episodic near-inertial motions can be generated in winter if conditions are favorable. This intriguing observation suggests

that a more mechanistic understanding of how sea ice dynamics and characteristics impact ocean near-inertial energy will be needed to predict how this energy will respond to future conditions.

The Arctic has undergone considerable changes in sea ice extent, thickness, and drift speed in recent decades (Comiso et al. 2008; Kwok and Rothrock 2009; Rampal et al. 2009; Rothrock et al. 2008) with concomitant changes in broadband momentum transfer across the air-ice-ocean interface (Brenner et al. 2021) and faster sea ice inertial oscillations (Gimbert et al. 2012b). It has been hypothesized that Arctic sea ice loss could increase near-inertial internal wave energy and mixing by enhancing atmosphere-ocean momentum transfer through a weaker and more mobile ice pack (e.g., Fine and Cole 2022). Thus far, however, the impact of sea ice on ocean near-inertial energy remains equivocal.

In the mixed layer, thinning and weakening ice could upend the historical paradigm of inertial oscillation damping. Idealized slab model experiments suggest that relative to open-ocean conditions, sea ice can actually *increase* inertial oscillation amplitude when internal ice stresses are weak (Brenner et al. 2023a). This intensification arises because the rough topside of the ice relative to open water increases the efficiency of momentum transfer from the atmosphere (Brenner et al. 2023a; Martin et al. 2014). These conflicting effects of sea ice on the ocean mixed layer’s inertial response (observed inhibition by internal ice stresses, modeled amplification by air-ice drag) indicate that improved physical understanding is needed of the role sea ice plays in modifying inertial oscillation generation.

In the pycnocline, early studies found that near-inertial internal wave energy was far below mid-latitude levels (D’Asaro and Morehead 1991; Levine et al. 1987). More recent studies have not found an interannual trend in internal wave energy or mixing in observational data collected over the last several decades (Dosser et al. 2021; Guthrie et al. 2013), and internal wave-driven dissipation in winter has *declined* in the Canada Basin (Dosser et al. 2021). This wintertime decline, which contradicts the prediction that weaker ice would enhance internal wave-driven mixing, may be due to decreases in rough multiyear ice floes (Dosser et al. 2021) as rougher ice increased observed internal wave generation at high ice concentrations (Cole et al. 2018). Again, these varying results suggest that our physical understanding of near-inertial motions in ice-covered waters remains incomplete.

Recent observations showed that ice dynamics strongly decreased the ocean’s inertial response in the northern Beaufort Sea but had a modest impact in the southern Beaufort Sea (Brenner et al. 2023a). In this paper we leverage the presence of this gradient to examine how sea ice mobility and dynamics impacted inertial oscillation development during an Arctic storm in February 2019. Focusing our study on an individual storm allows a detailed analysis of ice and ocean observations from a moored array that spanned nearly 600 km of the Beaufort Sea and provided measurements of contrasting sea ice conditions. Our results demonstrate how air-ice-ocean interactions across a wide range of interacting spatial and temporal scales, including the ice pack’s interaction with the distant coast, impacted near-inertial energy input to the mixed layer.

This paper is structured as follows. Section 2 describes the mooring observations and an ice-ocean slab model used to interpret the observations. Section 3 describes the evolution of the coupled atmosphere-ice-ocean system during a winter storm with a particular focus on near-inertial frequency motions of the ice and ocean mixed layer. Sections 4 and 5 discuss and summarize these results.

2. Methods

a. Overview

This paper uses data collected by an array of three moorings deployed in the Beaufort Sea’s deep Canada Basin from autumn 2018 to autumn 2019 as part of the Stratified Ocean Dynamics of the Arctic (SODA) experiment (Lee et al. 2016; Rainville et al. 2020). These moorings were oriented roughly south to north and are referred to as SODA–A, SODA–B, and SODA–C (Figure 1a). The array spanned a wide range of sea ice conditions, from weak and seasonal ice at the southern site to strong, year-round ice cover at the northern site (Brenner et al. 2023a). Each mooring was equipped with three acoustic Doppler current profilers (ADCPs) and numerous temperature and salinity sensors in the mixed layer and upper pycnocline. Some of these data were previously presented in a series of papers that developed higher-level products, including observations of sea ice geometry and the ice-ocean drag coefficient (Brenner et al. 2021), time series of mixed layer depth (Brenner et al. 2023b), and an idealized ice-ocean slab model that can be used to explore near-inertial motions generated by surface forcing (Brenner et al. 2023a).

In addition to contrasting observed inertial motions across the mooring array, we also compare observed velocities to those predicted by a slab model with no internal ice stresses. The slab model is useful in this context because it simulates the storm’s *potential* to generate inertial oscillations at each mooring in the absence of internal ice stresses. Thus, the model helps elucidate the impact of regional differences in ice dynamics on the observed response by controlling for regional differences in wind forcing, which also impact the observed response.

b. Moored ADCP measurements

Data from two ADCPs at each mooring are used in this paper. First, each mooring had an upward-looking 5-beam Nortek Signature-500 ADCP surveying ice and ocean velocities along with sea ice draft (using the instrument’s fifth beam as an altimeter). These ADCPs were located at 45 m for SODA–A, 42 m for SODA–B, and 27 m for SODA–C. See Brenner et al. (2021) for a description of Signature-500 ADCP processing. Second, upward-looking Teledyne Workhorse-300 ADCPs were located approximately 45 m below the Signature-500 ADCPs. There was some overlap in the vertical range measured by both ADCPs. This overlap was used to calibrate the Workhorse-300 compasses from the Signature-500 stored magnetometer data (see Brenner 2022, Chapter 2.3.1). Data from both ADCPs were then combined to create ocean velocity profiles spanning the upper ~80 m of the water column in 2-m vertical bins at hourly temporal resolution (see Brenner 2022, Chapter 2.3.2). Data from the uppermost 6 m of the water column were discarded due to sidelobe reflections.

1) WKB SCALING

Propagating internal waves are refracted by varying stratification, altering wave amplitude and wavenumber. If changes in stratification are small compared to the vertical scale of the waves, Wentzel-Kramers-Brillouin (WKB) scaling, a technique used to account for this refraction, can be used to compare data at different depths as if stratification were constant (Leaman and Sanford 1975). Though this paper primarily considers mixed layer near-inertial motions, wave propagation in the pycnocline is also of interest. WKB-scaled complex velocities U_n were calculated from observed pycnocline velocities U using:

$$U_n = U \sqrt{\frac{N_0}{N(z)}} \quad (1)$$

Stretched depth coordinates z' were calculated by integrating the stratification scaling factor from the deepest mixed layer depth observed at each mooring during the study period:

$$z' = \int_{MLD_{max}}^z \frac{N(z)}{N_0} dz \quad (2)$$

Stratification N was from CTD casts collected when moorings were deployed in the autumn. As stratification below the winter mixed layer varies at interannual timescales, we assume these observed stratification values are representative of pycnocline stratification throughout the mooring deployments. N_0 was the average stratification between the deepest mixed layer depth and 100 m. Stretched and unstretched depths were equal above the deepest observed mixed layer depth, but velocity scaling (Equation 2) was applied below the time-varying observed mixed layer depth.

c. Remote sensing and reanalysis

Several remote sensing and reanalysis products provided context about regional atmospheric and sea ice conditions to aid in interpretation of the mooring observations. Because of the overlying ice, the moorings did not observe meteorological variables, so surface air pressure and 10-m wind from ERA5 (Hersbach et al. 2023) were used to understand the storm's structure and to force the slab model. ERA5 data were downloaded at 0.25° horizontal resolution and hourly temporal resolution. NSIDC's Polar Pathfinder product (Tschudi et al. 2019) provided regional sea ice motion at 25-km horizontal resolution and daily temporal resolution. Moderate Resolution Imaging Spectroradiometer (MODIS) Level-1B brightness temperature at 1-km horizontal resolution (MODIS Characterization Support Team 2017) was used to visualize sea ice lead formation and evolution. Multiple MODIS images per day were available in our study region though clouds limited the usefulness of some images. Daily maps of sea ice leads at 1-km horizontal resolution (Willmes and Heinemann 2016; Reiser et al. 2020) were used to calculate the lead fraction (the lead area divided by the total ice and lead area in a given region; Willmes et al. 2023) in the vicinity of the moorings. Daily passive microwave sea ice concentration from AMSR2 at 3.125-km horizontal

resolution (Spreen et al. 2008) was assessed as a potential proxy for the importance of ice dynamics in inertial oscillation damping.

d. Coupled ice-ocean slab model

1) MODEL OVERVIEW

The ocean’s response to wind forcing, including Ekman transport and inertial oscillations, can be modeled as a “slab” in which the wind stress is distributed uniformly throughout the mixed layer as a body force (Pollard and Millard 1970). These slab model formulations typically include a linear damping term to simulate inertial oscillation decay due to internal wave radiation, mixing, or turbulent dissipation (e.g., D’Asaro 1985; Pollard and Millard 1970).

We use a one-dimensional coupled ice-ocean slab model to explore the ice and ocean response to wind forcing. This model formulation was developed by Brenner et al. (2023a) and is similar to the model developed by Gimbert et al. (2012a). We adopt D’Asaro (1985)’s notation for the complex velocities $Z = u + iv$ and stresses $T = (\tau_x + i\tau_y)/\rho_o$. Briefly, the ice and ocean momentum equations are coupled by the ice-ocean stress and modeled as:

$$\frac{dZ_i}{dt} + ifZ_i = \frac{T_{ai} - T_{io}}{d} - r_i Z_i \quad (3)$$

$$\frac{dZ_o}{dt} + ifZ_o = \frac{T_s}{H - d} - r_o Z_o \quad (4)$$

where d is the sea ice draft, H is the mixed layer depth, and r_i and r_o are damping coefficients approximating internal ice stresses and stress at the mixed layer base, respectively. The subscripts $()_i$ and $()_o$ indicate ice or ocean. The ocean velocity Z_o is vertically averaged in the mixed layer.

The air-ocean, air-ice, and ice-ocean stresses were calculated using quadratic drag laws as:

$$T_{a\Box} = \left(\frac{\rho_a}{\rho_o} \right) C_{a\Box} Z_a |Z_a| \quad (5)$$

$$T_{io} = C_{io} (Z_i - Z_o) |Z_i - Z_o| \quad (6)$$

with ρ_a the density of air, $C_{a\Box}$ the atmosphere-ice (C_{ai}) or atmosphere-ocean (C_{ao}) drag coefficient, and C_{io} the ice-ocean drag coefficient. As the wind speed is typically much larger than the ice or ocean speed, the velocity of the underlying (ice or ocean) surface was neglected in Equation 5, but the ice-ocean relative velocity was used to calculate the ice-ocean stress (Equation 6). The total ocean surface stress T_s in mixed sea ice and open water conditions was calculated by scaling T_{ao} and T_{io} by ice fraction A :

$$T_s = AT_{io} + (1 - A)T_{ao} \quad (7)$$

2) MODEL IMPLEMENTATION

Equations 3 and 4 were solved using *ode15s*, MATLAB's variable order initial value problem solver for stiff ODEs, which uses adaptive time stepping for computational efficiency. The ice and ocean were initially at rest, and the model was spun up for several days before the period of interest. The observed hourly ice draft was from ADCP measurements (Brenner et al. 2021) and the observed daily mixed layer depth was derived from ADCP backscatter and near-inertial shear (Brenner et al. 2023b). Following Brenner et al. (2023a), a minimum sea ice draft of 0.10 m was imposed to stabilize Equation 3. Time-varying wind forcing $Z_a = u_{10m} + iv_{10m}$ and sea ice concentration A from ERA5 at hourly temporal resolution and 0.25° horizontal resolution were averaged in a 30 km radius of each mooring, consistent with Brenner et al. (2021) and Brenner et al. (2023a). The time-varying drag coefficients were calculated as follows: C_{ao} varied with wind speed following Large and Yeager (2004), C_{ai} varied with ice concentration following Hersbach et al. (2023) with maximum values at intermediate ice concentrations, and C_{io} was calculated weekly using a force-balance approach and observed ice and ocean velocities (Brenner et al. 2021).

The tuneable ice and ocean decay constants r_i and r_o approximate damping by internal ice stresses and internal wave radiation, respectively. Brenner et al. (2023a) chose r_i and r_o to maximize model fit to observed ice and ocean velocities, allowing r_i to vary with ice thickness and concentration. Our use of the slab model is motivated by a somewhat different experimental philosophy; here, we compare observed velocities, which are influenced by internal stresses, to modeled velocities for a stress-free ice pack, assuming that differences between observed and modeled velocities can be attributed primarily to the unmodeled internal stresses. Thus, we use $r_i = 0$ to simulate an ice pack

in free drift (i.e., with negligible internal ice stress divergence). The time invariant ocean damping coefficient r_o was determined for each mooring by minimizing the RMSE between the observed and modeled mixed layer near-inertial velocities when $A < 0.85$ (i.e., periods of the year-long mooring deployments when internal ice stresses were likely small, so using $r_i = 0$ in the fit was realistic). The calculated r_o values were $0.03f$ at SODA–A, $0.015f$ at SODA–B, and $0.025f$ at SODA–C, comparable to the Beaufort Sea value of $0.05f$ found by Fine et al. (2021) but significantly smaller than the commonly-used mid-latitude value of $0.15f$ (Alford 2001).

e. Near-inertial filtering and frequency determination

The near-inertial frequency band was defined as horizontal velocities with frequency magnitude between $0.9-1.1f$. At the latitudes of our study, this frequency band also includes semidiurnal tides, so barotropic tidal velocities from the Arc5km2018 tidal model (Erofeeva and Egbert 2020) were subtracted from the observed ice and ocean velocities prior to near-inertial filtering. Tides are weak in this region; SODA–C had the strongest tides with maximum speeds of 0.005 m s^{-1} . Following de-tiding, near-inertial velocities were extracted from mooring observations and slab model results by bandpass filtering with a fourth-order low-pass Butterworth filter that permitted frequencies from $0.9-1.1f$.

The frequency ω of the observed or slab modeled near-inertial response was determined by minimizing the difference between the observed or slab modeled velocity and a fitted sine wave with a prescribed test frequency and optimized amplitude and phase. This approach was preferred to identifying a spectral peak because the time periods of interest were relatively short so spectra computed from the hourly velocity data were not well resolved. The test ω values were fit to velocity data in a four-day window beginning on 25 February as this time period captures the near-inertial response to the storm considered in this study. The eastward and northward components of velocity were fit separately. The best-fit ω values are reported as frequency ranges for which R^2 between the observed or slab modeled velocity and fitted velocity exceeded 0.95 for both the eastward and northward components of velocity. While the frequency yielding the best fit depended somewhat on the temporal window of observed or slab modelled velocity used for fitting, the conclusions are not sensitive to shrinking, expanding, or shifting the temporal window as long as the focus remains on the storm response period.

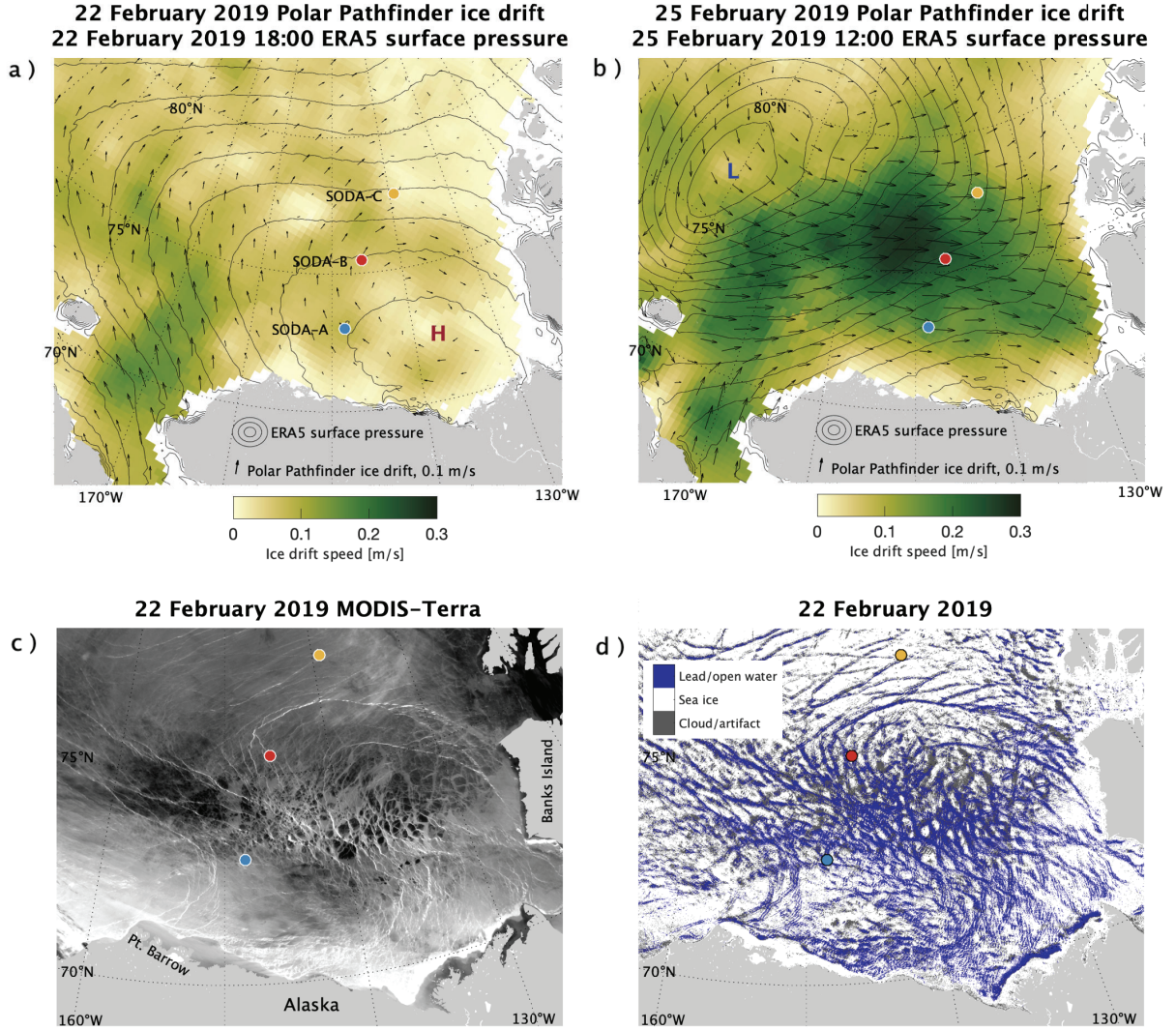


FIG. 1. Regional atmospheric and sea ice conditions. (a-b) ERA5 surface pressure (gray contours) and Polar Pathfinder ice drift (color and black vectors; Tschudi et al. 2019) on (a) 22 February and (b) 25 February. (c-d) MODIS-Terra (c) brightness temperature (Band 31) imagery and (d) identified leads (Willmes and Heinemann 2016; Reiser et al. 2020) on 22 February. The SODA-A, SODA-B, and SODA-C mooring locations are shown in all panels.

3. Results

a. Results overview

This paper examines the evolution of the atmosphere, sea ice, and ocean before, during, and after a storm in the central Beaufort Sea in February 2019. This series of events can be divided into

three phases: sea ice lead formation (Section 3b), storm passage (Section 3c), and the observed inertial response (Section 3d). Prior to the storm, from 22–23 February, winds pushed the ice pack against the Canadian Arctic Archipelago and the Canadian and Alaskan mainland east of Point Barrow (Figure 1a), opening a network of leads that traversed some of the study area (Figure 1c-d). Following lead formation, from 24–26 February, a low pressure system brought strong (up to 16 m s^{-1}), clockwise-rotating winds to the region (Figures 1b, 2a). This storm excited sea ice and ocean mixed layer inertial oscillations whose magnitudes peaked between 26–28 February (Figures 3e, 4e, 5e). The observed ice and ocean near-inertial velocities varied in accord with the history of lead activity, with stronger inertial oscillations at the fractured sites. In contrast to the differing near-inertial responses, the storm caused comparable low-frequency ice motion at the southern (SODA–A) and northern (SODA–C) moorings.

b. 22–23 February: Sea ice lead formation

High atmospheric pressure moved eastward along the northern Alaskan coast on 22 February 2019. Anticyclonic winds compacted ice against the coastline (Figure 1a), and leads opened in the ice pack. One lead system extended from Point Barrow northward toward SODA–A and a second lead system extended westward from Banks Island toward SODA–B (Figure 1c). These topographic points are known hotspots for lead initiation (Reiser et al. 2020, their Figure 10), with similar wind conditions favoring lead formation across the Beaufort Sea (Jewell and Hutchings 2023; Jewell et al. 2023). On 22 February the lead fraction increased to 35% in $200 \text{ km} \times 200 \text{ km}$ regions around SODA–A and SODA–B but remained below 5% in a $200 \text{ km} \times 200 \text{ km}$ region around SODA–C (Figure 1d, Figure 2b). AMSR2 ice concentration (Spreen et al. 2008) averaged in the same $200 \text{ km} \times 200 \text{ km}$ regions remained above 90% at all mooring sites throughout the study period.

Subsequent MODIS imagery revealed anticyclonic ice motion and additional ice fracturing from coastal promontories between these bounding leads (Supplementary Movie 1). This behavior suggests the ice pack may have lost contact with the coast (Jewell et al. 2023) relieving internal ice stresses at SODA–A and SODA–B. The ice pack at SODA–C, in contrast, was isolated from this breakout by the northernmost lead, suggesting sustained internal ice stresses at the northern mooring. In situ and satellite ice velocities confirm differences in ice mobility across the mooring

array following lead formation. The wind factor (the ratio of ice speed to wind speed) calculated from ice speeds observed by mooring ADCPs from 22–23 February was elevated at SODA–A and SODA–B relative to SODA–C (Figure 2c), which experienced near-stationary ice on 23 February (Figure 2a). On 23 February, Polar Pathfinder ice speeds (Tschudi et al. 2019) were higher around the southern moorings with a distinct decrease between SODA–B and SODA–C (not shown).

c. 24–26 February: Storm passage

From 24–26 February 2019, a low-pressure system moved from the Bering Sea to the Chukchi and Beaufort Seas on the west side of the SODA mooring array. The storm originated in the Pacific Ocean and its track into the Beaufort Sea agrees with the unusual northward jet stream orientation that developed in this region in February–March 2019 (Overland 2020). On 25 February, the 975 mb low pressure center was located approximately 1000 km west of the center of the array (Figure 1b), then continued northward and dissipated over the following 24 hours. Wind speeds at the mooring sites peaked late on 25 February, reaching 14 m s^{-1} at SODA–A (Figure 3b), 16 m s^{-1} at SODA–B (Figure 4b), and 13 m s^{-1} at SODA–C (Figure 5b). Broadband ice speeds were fastest near the center of the mooring array (Figure 1b), reaching about 0.5 m s^{-1} at SODA–B and 0.3 m s^{-1} at SODA–A and SODA–C (Figure 2a).

Lead fraction decreased to below 7% near SODA–A and SODA–B on 24–25 February but increased to 14% on 26 February (Figure 2a). Though leads can be sites of rapid ice growth, they remain local weaknesses in the ice cover (Feltham 2008). These trends in ice fraction near the southern moorings likely reflect refreezing in leads followed by lead reopening along these preexisting weaknesses by the storm. Lead fraction near SODA–C remained below 5% during and after the storm (Figure 2b).

d. 26–28 February: Observed inertial response

Following the low pressure system’s passage, high pressure developed over the SODA mooring array (not shown). Over the course of these two weather systems the clockwise turning of the wind (Figures 3b, 4b, 5b) excited inertial oscillations in the sea ice and ocean mixed layer. Observed inertial oscillation strength varied across the mooring array, with stronger oscillations at the fractured southern sites. The maximum observed near-inertial mixed layer speed was 0.049 m s^{-1}

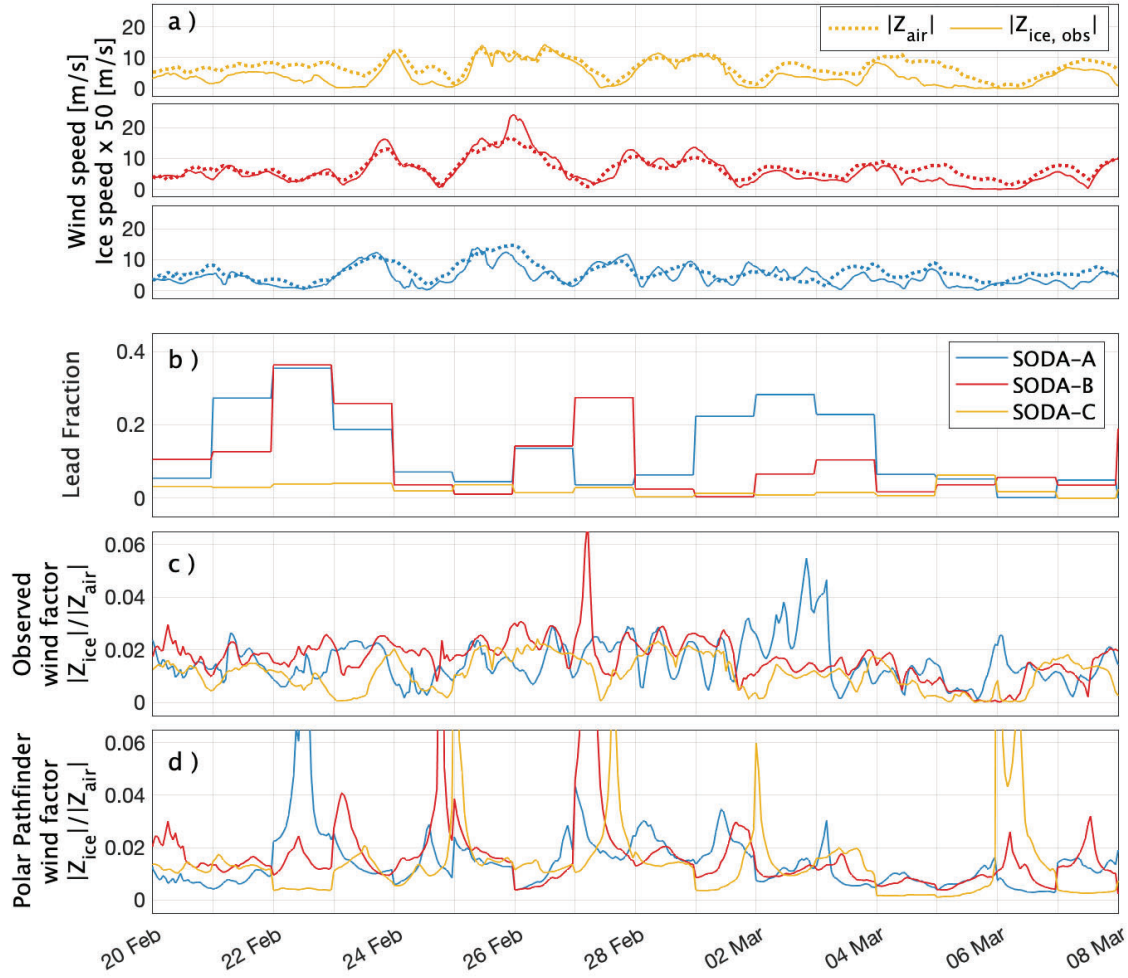


FIG. 2. Comparison across moorings of (a) wind speed (dotted lines) and observed ice speed scaled by a factor of 50 (solid lines), (b) lead fraction (Willmes et al. 2023) in a 200 km × 200 km box centered on the mooring site, (c) wind factor calculated from ERA5 wind speed and observed ice speed, (d) wind factor calculated from ERA5 wind speed and Polar Pathfinder ice speed (Tschudi et al. 2019). Blue lines are for SODA–A, red lines are for SODA–B, and yellow lines are for SODA–C.

on 28 February at SODA–A (Figure 3e), 0.042 m s^{-1} on 26 February at SODA–B (Figure 4e), and 0.018 m s^{-1} on 25 February at SODA–C (Figure 5e).

The regional pattern in modeled inertial oscillation strength differed from the regional pattern in observed inertial oscillation strength. Modeled inertial oscillations were fastest at SODA–C and slowest at SODA–A, indicating the storm’s inertial oscillation generation *potential* was strongest

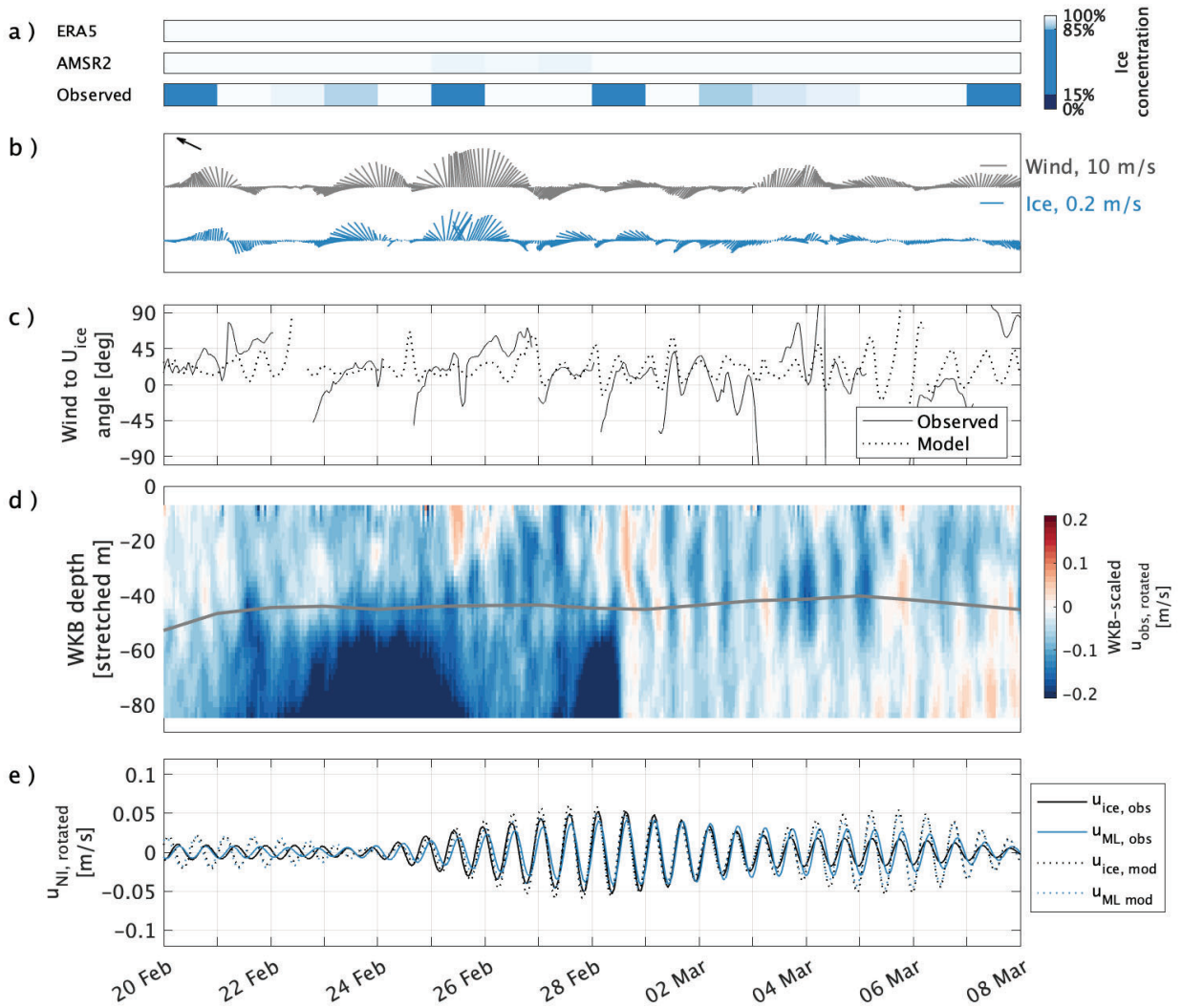


FIG. 3. Data for the SODA-A mooring site. (a) Sea ice concentration from ERA5, AMSR2, and observed by the mooring altimeter. (b) Wind velocity from ERA5 and ice velocity measured by mooring ADCP. (c) The angle between the wind velocity and the observed (solid line) and modeled (dotted line) ice velocity, with positive values indicating ice deflected to the right of the wind. Angle not plotted for ice speeds below 0.03 m s^{-1} . (d) WKB-scaled velocity (colors) and mixed layer depth (gray line). (e) Observed (solid lines) and modeled (dotted lines) near-inertial sea ice velocity (black) and mixed layer averaged velocity (blue). The direction of maximum wind speed defined a rotated coordinate system in which all velocities are plotted; north is indicated by the black arrow in the upper left corner of panel (b).

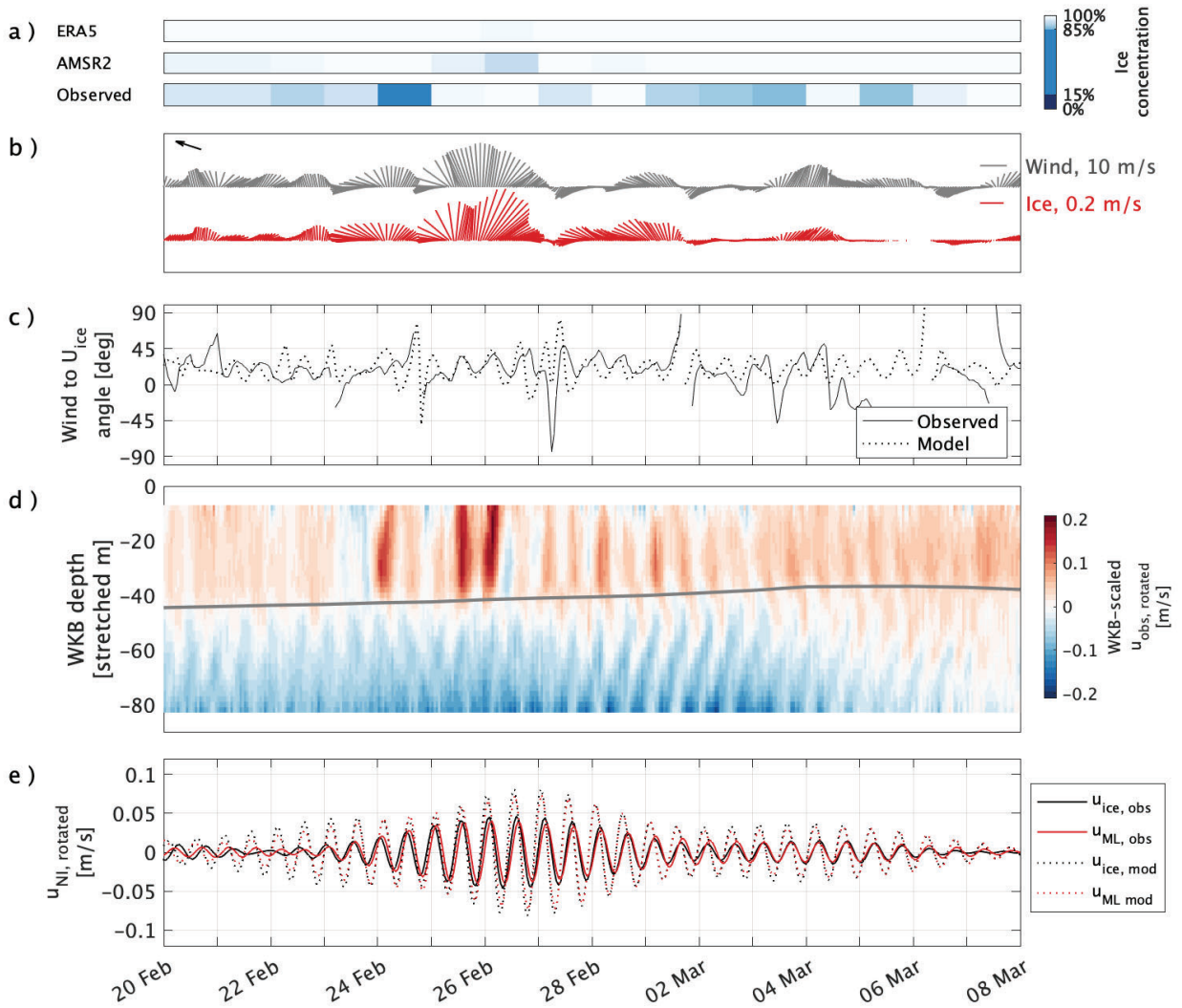


FIG. 4. Data for the SODA-B mooring site. (a) Sea ice concentration from ERA5, AMSR2, and observed by the mooring altimeter. (b) Wind velocity from ERA5 and ice velocity measured by mooring ADCP. (c) The angle between the wind velocity and the observed (solid line) and modeled (dotted line) ice velocity, with positive values indicating ice deflected to the right of the wind. Angle not plotted for ice speeds below 0.03 m s^{-1} . (d) WKB-scaled velocity (colors) and mixed layer depth (gray line). (e) Observed (solid lines) and modeled (dotted lines) near-inertial sea ice velocity (black) and mixed layer averaged velocity (red). The direction of maximum wind speed defined a rotated coordinate system in which all velocities are plotted; north is indicated by the black arrow in the upper left corner of panel (b).

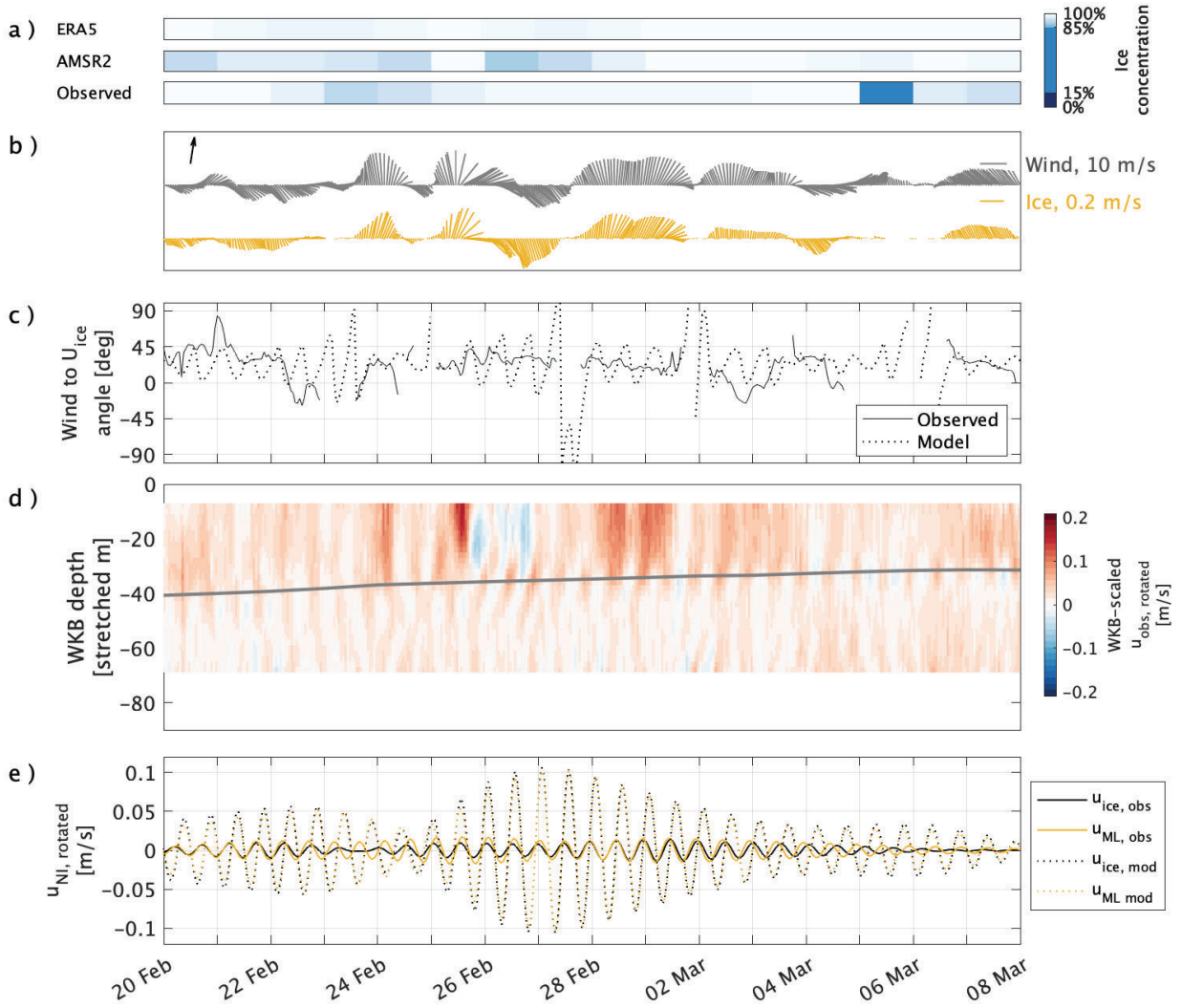


FIG. 5. Data for the SODA-C mooring site. (a) Sea ice concentration from ERA5, AMSR2, and observed by the mooring altimeter. (b) Wind velocity from ERA5 and ice velocity measured by mooring ADCP. (c) The angle between the wind velocity and the observed (solid line) and modeled (dotted line) ice velocity, with positive values indicating ice deflected to the right of the wind. Angle not plotted for ice speeds below 0.03 m s^{-1} . (d) WKB-scaled velocity (colors) and mixed layer depth (gray line). (e) Observed (solid lines) and modeled (dotted lines) near-inertial sea ice velocity (black) and mixed layer averaged velocity (yellow). The direction of maximum wind speed defined a rotated coordinate system in which all velocities are plotted; north is indicated by the black arrow in the upper left corner of panel (b).

at the northern mooring site. Regional differences in modeled inertial oscillation strength reflect regional variation in the resonance of the wind forcing and differences in mixed layer depth but do not account for impacts of ice dynamics, which are not included in the model. The mixed layer was shallowest at SODA–C (average mixed layer depth was 44 m at SODA–A, 41 m at SODA–B, 36 m at SODA–C; Figures 3d, 4d, 5d), contributing to stronger modeled velocities (Equation 4).

Comparing the predicted and observed near-inertial response at each mooring indicates varying degrees of inertial oscillation modification by unmodeled processes (i.e., ice dynamics). The maximum modeled near-inertial mixed layer speed at SODA–A was 0.050 m s^{-1} on 28 February (Figure 3e), comparable to the maximum near-inertial speed observed at that site. This similarity between observations and model results shows that ice damping was modest at SODA–A despite high ice concentrations. At SODA–C, conversely, the model predicted a much stronger inertial response than developed in the observations. Mixed layer oscillations at SODA–C were predicted to reach about 0.10 m s^{-1} on 27 February (Figure 5e), making the modeled SODA–C response to this storm the strongest potential mixed layer inertial oscillation event under pack ice ($A > 0.85$) at any of the moorings throughout the year-long record. The 82% reduction in observed near-inertial velocity at SODA–C relative to the model implies significant damping by ice dynamics.

Importantly, these ice dynamics at SODA–C impacted ice mobility differently at different frequencies. Despite the weaker near-inertial currents at SODA–C relative to SODA–A, observed broadband ice speeds and wind factor at SODA–C were comparable to those at SODA–A on 25 February (Figure 2b-c). Ice dynamics appeared to limit near-inertial velocities at SODA–C by restricting the angle by which ice velocities could deviate from wind velocities. The angle between the wind and the SODA–C *observed* ice velocity remained relatively unchanging over the course of an inertial period (Figure 5c solid line) while the *modeled* ice velocity varied by 45° or more relative to the wind as the modeled ice underwent strong inertial oscillations (Figure 5c dotted line). At SODA–A, in contrast, both the observed and modeled ice velocities experienced comparable deviations from the wind direction over the course of an inertial period (Figure 3c), suggesting the broken ice allowed the ice to change direction swiftly and oscillate independently of the wind direction.

The SODA–B observed near-inertial response was weaker and lower frequency than the model’s predicted response (Figure 4e). The maximum modeled near-inertial mixed layer speed was 0.074

m s⁻¹ on 26 February, or 43% greater than the maximum observed speed, indicating an intermediate amount of damping by ice dynamics at this central mooring. The frequency of the observed near-inertial mixed layer response was 0.99-1.01*f* while the modeled response was 1.02-1.04*f*. This slight frequency mismatch may be due to the reanalysis failing to exactly reproduce the winds at SODA-B. Indeed, winds from another reanalysis product (NCEP) have been shown to have lower energy than observations in this region at frequencies near and above the inertial frequency (Fine et al. 2021).

4. Discussion

This paper’s results support Brenner et al. (2023a)’s suggestion that internal ice stresses filter out high-frequency ocean velocities while transmitting momentum at sub-inertial frequencies. McPhee (1978) came to a similar conclusion while examining data from AIDJEX: “Even though a whole region of ice moves with little apparent resistance, internal forces at smaller scales in the compact winter pack rapidly quell the inertial component.” He attributed the inhibition of high-frequency oscillations to a relatively unbroken ice pack, stating “Apparently as freeze-up progresses, the ice becomes rigid on a large enough scale to inhibit the short-period motion almost entirely” (McPhee 1978). Our observations support McPhee (1978)’s longstanding hypothesis that floe scales determine if inertial motions are damped. However, our results challenge the notion that this rigidity is a strictly seasonal phenomenon, as large-scale wintertime lead formation episodes have been documented in decades-old satellite imagery (e.g., Jewell et al. 2023).

While this paper has focused on the magnitude of the near-inertial response, ice may also impact other aspects of this response. Hodographs of ice and ocean near-inertial velocity showed sometimes-elliptical orbits, with differences in near-inertial current structure across mooring sites as well as differences between the ice and ocean near-inertial velocity at each mooring (Figure 6). The theoretical ratio of the major to minor axes lengths is ω/f for free waves, so elliptical orbits can result from super-inertial frequencies. However, the elliptical axis ratios in Figure 6 are too large to be explained by the observed frequencies. The observed frequencies for the ice and mixed layer near-inertial velocities were 0.97-0.99*f* and 0.99-1.01*f*, respectively, at SODA-A, and 0.99-1.01*f* for both the ice and mixed layer at SODA-B. At SODA-C, the mixed layer frequency was

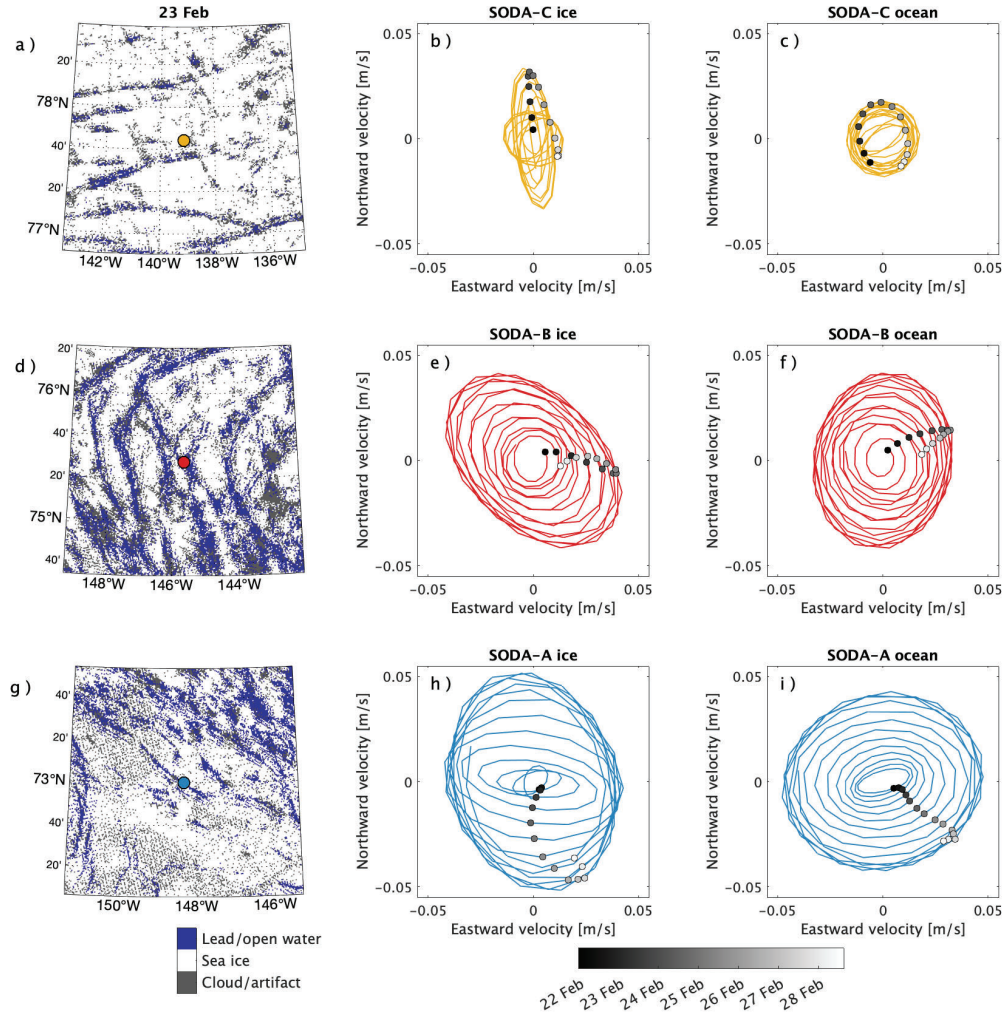


FIG. 6. Sea ice leads and near-inertial hodographs at each mooring site. (a,d,g) Identified leads (Willmes and Heinemann 2016; Reiser et al. 2020) on 23 February in 200 km $\times$ 200 km regions surrounding each mooring, with mooring locations shown by colored dots. Hodographs at each mooring of observed near-inertial (b,e,h) sea ice velocity and (c,f,i) ocean mixed layer velocity. Grayscale points are plotted every inertial period.

1.06-1.07 f . No fitted frequency exceeded the $R^2 = 0.95$ goodness of fit threshold for ice velocity at SODA-C; the best fit frequency was 1.05-1.06 f ($R^2 > 0.85$).

As the observed frequencies were close to the inertial frequency, the elliptical orbits suggest these inertial oscillations were not free waves. It is possible that the ice fracture orientation mechanically restricted the inertial current orientation; at SODA-B, in particular, the ice inertial oscillation ellipses align with the dominant axis of lead formation (Figure 6d-f). The major axes of the mixed

layer inertial oscillation ellipses at SODA–A and SODA–B were rotated to the right of the sea ice ellipses, implying rotational shear in the near-inertial band. We suggest that the ways in which ice characteristics, including lead orientation, imprint on near-inertial currents is an important topic for future research.

Observations capable of resolving near-inertial motions, such as the ADCP data presented in this paper, are rare in the Arctic. Our results demonstrate the importance of ice dynamics to the development of near-inertial motions, so we next ask how these ice impacts could be parameterized. Ice concentrations above 0.85 and wind factor values below 2% have been used as “rule-of-thumb” thresholds to identify times when low-frequency ice motion is damped by internal ice stresses (e.g., Brenner et al. 2021), but we find that these proxies are inadequate for predicting inertial motion damping in our data.

Ice concentration from AMSR2 was high at all mooring sites throughout the storm (>0.97 at SODA–A, >0.92 at SODA–B, >0.89 at SODA–C; Figures 3a, 4a, 5a), which might suggest region-wide damping by internal stresses. However, observed differences in near-inertial damping between the fractured and unfractured sites demonstrate some amount of independence of inertial motions and ice concentration (i.e., inertial motion was stronger at SODA–A despite its higher ice concentration). This independence likely arose because leads can even be found in areas where ice concentrations derived from passive microwave data are near 100% (Willmes and Heinemann 2016). Remotely sensed sea ice concentrations depend on instrument resolution, ice thickness, and, for thermal infrared measurements, accurate cloud masking (Ludwig et al. 2020). Ice concentrations dropped below 85% at SODA–A and SODA–B (Figures 3a, 4a) when calculated from ADCP altimeter data as the ratio of ice-covered distance to total (ice and open water) distance (Brenner et al. 2021), supporting the idea that the southern moorings experienced localized lead openings that were not captured by the passive microwave product.

We next tested if inertial oscillation damping could be inferred from wind factor values computed using a readily available satellite-derived daily ice drift product, Polar Pathfinder (Tschudi et al. 2019). The wind factor can be thought of as measuring ice mobility, or how readily the ice responds to forcing. When calculated using hourly ice velocity from the ADCP, the wind factor at SODA–C was around 2% on 26 and 28 February but decreased to near zero on 27 February in conjunction with very low ice speed (Figure 2b-c). This behavior is consistent with internal ice stresses

decreasing ice motion. However, Polar Pathfinder did not capture this observed decline in wind factor. Instead, Polar Pathfinder wind factor near the northern mooring increased on 27 February (Figure 2d) due to unrealistically large Polar Pathfinder ice speed (around 0.1 m s^{-1}). Overall, we find no consistent differences in Polar Pathfinder wind factor across mooring locations between 25 February and 1 March (Figure 2d). In summary, our results do not show a relationship between inertial oscillation strength and either passive microwave ice concentration or low frequency ice mobility during this event.

Though our focus on a single storm limits our confidence in the universality of these results, we suggest that a metric like floe size distribution (FSD), which describes the horizontal scales of an assortment of floes (Rothrock and Thorndike 1984), might better predict near-inertial damping. While distinct floes are more easily identified in marginal ice zones, in winter the lengths of line segments on floes, or chord lengths, can be used to characterize FSD (Rothrock and Thorndike 1984). FSD has been shown to influence many aspects of the atmosphere-ice-ocean system including ice melt rate and surface wave propagation into the ice pack (Steele 1992; Collins III et al. 2015). Algorithms have been developed to extract FSD statistics from satellite imagery (Denton and Timmermans 2022; Stern et al. 2018) and satellite altimeter data (Horvat et al. 2019). While existing FSD data sets do not cover the time and location of our study and applying these algorithms is outside the scope of the current work, we suggest that FSD records could be compared to existing high frequency Arctic mixed layer velocity records (such as the Ice-Tethered Profiler with Velocity data set; Williams et al. 2010) to refine the relationship between FSD and near-inertial motions.

Understanding the conditions that favor lead formation is an area of active research (e.g., Jewell and Hutchings 2023; Jewell et al. 2023; Lewis and Hutchings 2019; Rheinländer et al. 2022). Ice strength increases with ice thickness so, in general, stronger mechanical forcing (i.e., faster winds) will more easily break up thinner ice (Weiss 2013; Graham et al. 2019). However, most leads form under tension rather than compression or shear (Lewis and Hutchings 2019) and ice has a much lower tensile strength than compressive or shear strength. This tensile failure mode may explain why observed lead formation is independent of ice age, thickness, and compressive strength (Lewis and Hutchings 2019), and why there has been no observed trend in lead formation in the Beaufort Sea over the last two decades (Lewis and Hutchings 2019; Willmes et al. 2023). Despite ongoing

ice thinning and multiyear ice loss, future wintertime lead activity may depend primarily on future atmospheric patterns and their characteristic alignments of wind and coastal geometry (Lewis and Hutchings 2019). Thus, basin geometry, by way of its influence on lead formation, may be an unappreciated contributor to regional differences in Beaufort Sea inertial motions when midwinter pack ice conditions are present.

5. Summary and Conclusion

This paper presented high temporal resolution observations of the sea ice and ocean near-inertial velocity response to a storm in February 2019 in the Beaufort Sea. These observations were interpreted alongside results from a coupled ice-ocean slab model that isolated the effects of ice dynamics by controlling for regional variation in wind forcing. Contrasting the observed and modeled responses across a region of varying sea ice lead activity showed that ice dynamics strongly damped the near-inertial response at the northern mooring (77.7°N) but minimally impacted in the near-inertial response at the southern mooring (73°N). Our primary findings can be summarized as follows:

- A fractured ice pack with a recent history of lead formation allowed inertial oscillations to develop in the sea ice and ocean mixed layer even in a dense, mid-winter ice pack.
- Floe size distribution may be a useful metric for predicting near-inertial motions in consolidated ice. Passive microwave ice concentration and low frequency ice mobility did not predict near-inertial motions during this event.
- Unlike the near-inertial response, the broadband ice velocity did not vary in accord with the history of lead activity. Ice dynamics selectively filtered out high frequency motions.

Our results demonstrate the first-order role that ice dynamics, particularly lead formation, play in modifying near-inertial energy input to the ocean mixed layer. Leads, which can extend hundreds of kilometers from coastal promontories into the ice pack, therefore link ocean dynamics to distant coastal boundaries. Future trends in winter inertial oscillation strength may depend more on trends in lead activity and floe size than on commonly modeled parameters like ice thickness. Consequently, efforts to predict wintertime inertial motions in the Arctic will benefit from ongoing efforts to accurately represent floe size distribution in models (Zhang et al. 2016, 2015; Roach

491 et al. 2018) and to develop ice rheologies that realistically model cracks and leads in the ice pack
492 (Ólason et al. 2022).

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Data availability statement. The mooring measurements and model results described in this paper will be made available on the University of Washington ResearchWorks Archive before publication (similar to the repository at <https://digital.lib.washington.edu/researchworks/handle/1773/49495>). ADCP ocean velocity measurements and mixed layer depth for the full deployments of the SODA moorings are available at <https://digital.lib.washington.edu/researchworks/handle/1773/49495>. ERA5 reanalysis (Hersbach et al. 2023) is available at <https://doi.org/10.24381/cds.adbb2d47>. Daily NSIDC Polar Pathfinder sea ice drift (Tschudi et al. 2019) is available at <https://nsidc.org/data/nsidc-0116/versions/4>. AMSR2 sea ice concentration (Spreen et al. 2008) is available at <https://seaice.uni-bremen.de/sea-ice-concentration/amsre-amsr2/>. Sea ice lead classification (Reiser et al. 2020; Willmes and Heinemann 2016) is available at <https://meteo.uni-trier.de/v2/arcleads.php>. MODIS imagery (MODIS Characterization Support Team 2017) is available at <https://ladsweb.modaps.eosdis.nasa.gov/missions-and-measurements/products/MOD021KM> and <https://ladsweb.modaps.eosdis.nasa.gov/missions-and-measurements/products/MYD021KM>. Tidal velocities from the Arc5km2018 model (Erofeeva and Egbert 2020) are available at <https://arcticdata.io/catalog/view/doi:10.18739/A21R6N14K>.

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