RESEARCH ARTICLE

The importance of capturing late melt season sea ice conditions for modeling the western Arctic ocean boundary layer

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To better understand the response of the western Arctic upper ocean to late summer ice-ocean interactions, a range of surface, interior, and basal sea ice conditions were simulated in a 1-D turbulent boundary layer model. In-ice and under-ice autonomous observations from the 2014 Marginal Ice Zone Experiment provided a complete characterization of the late melt-season sea ice and were used to set initial conditions, update boundary conditions, and conduct model validation studies. Results show that underestimates of open water and melt pond fraction at the sea ice surface had the largest influence on ocean-to-ice turbulent heat fluxes reducing basal melt rates by as much as 32%. This substantial reduction in latent heat loss was attributed to underestimates of open water areas and the exclusion of melt ponds by low-resolution synthetic aperture radar imagery. However, the greatest overall effect on the ice-ocean boundary layer came from mischaracterizations of basal roughness, with smooth ice scenarios resulting in 7 m of summer halocline shoaling and preservation of the near-surface temperature maximum. Rough ice conditions showed a 23% deepening of the mixed layer and erosion of heat storage above 40 m. Adjustments of conductive heat fluxes had little effect on the near-interface heat budget due to small internal thermal gradients within the late summer sea ice. Results from the 1-D boundary layer simulations highlight the most influential components of sea ice structure during late summer conditions and provide the magnitude of errors expected when ice conditions are mischaracterized.

Keywords: Through-ice transmissivity; Sea ice characterizations; Ice-ocean boundary layer; Summer haloclines; Near-surface temperature maximums; Local turbulence closure

1. Introduction

Sea ice has an unusually strong influence on the high latitude geophysical system given its volume in the air-ice-ocean system. Although small in scale, sea ice acts as the primary coupler of atmospheric forcing (heat and momentum) and controls much of the oceanic response to these inputs. Given the importance of this geophysical material, realistically representing the sea ice surface condition, internal gradients, and basal topology are fundamental to predicting ice-ocean interactions. On the surface, the wide-ranging optical properties of varied ice conditions scale the intensity of incoming shortwave radiation (Perovich, 2005). Light et al. (2008) showed that the transmissivity of light through sea ice is significant, with 3–10 times more solar radiation passing through the ice than previously understood or modeled. The transmissivity, or fraction of solar radiation surviving passage through the sea ice, has been shown to be highly heterogeneous and to increase with the development of melt ponds in early summer as surface albedos decline (Frey et al., 2011). The transition from ridged perennial ice to smoother seasonal ice further enhances this transmissivity response as melt pond coverages have expanded from 30–40% on multi-year ice (Tschudi et al., 1997; Fetterer and Untersteiner, 1998; Perovich et al., 2002) to as much as 70% on first-year ice (Polashenski et al., 2012). In the sea ice interior, vertical temperature gradients affect the magnitude of conductive heat fluxes and the thermodynamic growth/melt rate at the sea ice base (Pringle et al., 2007). Additionally, sea ice temperatures control permeability, as the “law of fives” are satisfied, permitting buoyant meltwater to enter under-ice cavities and adjacent leads (Golden et al., 2007). Under the sea ice base, the lengths of roughness elements determine the efficiency of ice-ocean momentum transfer and the strength of turbulent eddies (McPhee et al., 2002; Shaw et al., 2008). Recent work by Cole et al. (2017) showed that air-ice-ocean momentum transfer is not static, but changes seasonally as a function of ice topography, ice concentration, and ocean stratification. These findings are particularly important given the expansion of the western Arctic seasonal ice zone and the transition from ridged perennial ice to smooth first-year ice (Comiso, 2012; Hwang et al., 2017).
The summer melt season has significant influences on the properties and structure of the Arctic ocean boundary layer. Changes in sea ice optical properties and open water fraction add substantial heat to the near surface ocean as downwelled solar radiative fluxes are accumulated (Perovich et al., 2007). In summer, this absorbed heat generally has one of two fates: 1) residence within the layer in which it was deposited; or 2) loss to latent heat transfer at the ice-ocean interface due to turbulent mixing. Recent studies with ice-tethered profiler data and one-dimensional boundary layer models in the western Arctic have shown that this partitioning is approximately 0.23/0.77, respectively, and that thermodynamic forcing is primarily from the vertical component with limited lateral advectisons (Toole et al., 2010; Gallaher et al., 2016). Upper ocean buoyancy, provided by summer season meltwater from the sea ice, is the ‘thermostat’ that governs this distribution of heat as turbulent eddies interact with near-surface density gradients (McPhee, 1994). In addition to controlling heat partitioning, intensification of melt season buoyancy gradients decreases boundary layer depth (Peralta-Ferriz and Woodgate, 2015). This effect is especially true in the western Arctic where ice-tethered profiler observations have shown a reduction in the mean summer mixed layer to an average depth of 16 m (Toole et al., 2010).

In this study, late melt-season sea ice conditions were investigated to evaluate the influence that the surface, interior, and base of the ice have on the underlying ocean boundary layer. To accomplish this goal, we used data from the Office of Naval Research Marginal Ice Zone Experiment in 2014 (MIZEX 2014) to capture late summer sea ice and upper oceans conditions, then systematically adjust sea ice surface, interior, and basal conditions using a one-dimensional (1-D) model to measure turbulent boundary layer responses (McPhee, 2008). The objectives of this approach were to rank the most impactful sea ice component during late summer and quantify resulting errors to ocean-to-ice heat flux and boundary layer depth when conditions are mischaracterized.

2. Data and Methods

Observations used to characterize on-ice, in-ice, under-ice, and upper-ocean conditions were acquired from the first six weeks of the MIZEX Cluster 5 (C5) time series collected in the Canada Basin from 16 August through 26 September 2014 (Figure 1). Surface sea ice conditions, to include estimates of open water and melt pond fractions, were derived from 1-m resolution panchromatic MEDEA satellite images collected during the first week of the study (Kwok and Untersteiner, 2011; Webster et al., 2015). During the second week of the study, satellite based RADARSAT-2 and TerraSAR-X synthetic aperture radar (SAR) images were made available by the Center for Southeastern Tropical Advanced Remote Sensing. SAR images are an excellent all-weather alternative to the MEDEA visible images; however, they are disadvantaged by lower resolution (RS-2 ~ 100 m, TS-X ~ 8 m) and low signal-to-noise ratio due to significant speckle noise (Hwang et al., 2017). To improve the speckle noise contamination and improve ice floe interrogation, an edge-preserving bi-lateral filter was applied to the SAR images (Tomasi and Manduchi, 1998). Errors associated with variations in overhead scan angles

![Figure 1: Marginal Ice Zone (MIZ) Cluster 5 study area in 2014. Background is topo-bathymetric map of the Canada Basin showing the start (green square) and end (yellow square) position of MIZ Cluster 5 between 16 August and 26 September. RADARSAT-2 images from 21 August (left) and 26 August (right) 2014 are overlaid to show the relative position of the study time series to seasonal ice zone conditions. DOI: https://doi.org/10.1525/elementa.391.f1]
are assumed negligible due to the StripMap technology of TerraSAR-X (Stangl et al., 2003) and the small RADARSAT-2 image subset (10 km × 10 km) used in this study.

Two ice mass balance (IMB) systems were used to measure sea ice temperature and thickness: 1) the Cold Regions Research and Engineering Laboratory (CRREL) 2014F IMB (Polashenski et al., 2011), and 2) the Scottish Association for Marine Science (SAMS) IMB 22 (Jackson et al., 2013). The CRREL 2014F IMB was located on the MIZ C5 ice floe and recorded sea ice temperatures throughout the time series; however, the acoustic rangers used to determine ice thickness did not operate until 16 September. To mitigate the time gap, ice thickness/base observations were extrapolated from the IMB 22 located approximately 10 km to the west of MIZ C5. For this study, the IMB 2014F temperature time series was employed to update model sea ice boundary conditions; whereas, IMB 22 observations were used for validating basal ice melt in the model. Due to the 10-km separation between IMBs, basal ice melt validation using IMB 22 assumes similar ice-ocean interactions are occurring over this area, which is well supported given the close tracking of IMB observations after 16 September. Estimates of sea ice base roughness were measured from the turbulent flux package and GPS receiver onboard the autonomous ocean flux buoy (AOFB) 29 (see Shaw et al., 2008, for full AOFB details). Using velocity perturbations observed by the AOFB acoustic current meter (Falmouth Scientific Inc., ACM 3D current meter, 1.5 mm s⁻¹ rms noise level), momentum fluxes were calculated approximately 2.5 m below the ice base at MIZ C5 using eddy correlation methods detailed in Gallaher et al. (2016). The magnitude of shear-generated momentum fluxes, also known as friction velocity (\( u_* \)), were calculated using

\[
\begin{align*}
\text{(1)} \quad u_* &= \left( <u'w'>^2 + <v'w'>^2 \right)^{0.25}
\end{align*}
\]

where \(<u'w'>\) and \(<v'w'>\) are the x and y components of the turbulent Reynolds stresses, averaged over 15-minute ensembles, and represent the vertical transport of horizontal momentum. The height-adjusted (2.5 m below ice) ice-ocean drag coefficient \((C_d)\) was then calculated by

\[
\begin{align*}
\text{(2)} \quad C_d &= \frac{u_*^2}{U^2},
\end{align*}
\]

where \( u_* \) is the friction velocity observed by the AOFB at ~2.5 m below the ice base, and \( U \) is the observed ice speed. Ice speeds were calculated from GPS differencing of the AOFB 29 position data. During the 41-day study period, the AOFB turbulent package measured an average \( C_d \) of 5.3 × 10⁻³ (Figure 2). The observed \( C_d \) of 5.3 × 10⁻³ is slightly lower than the established ice-ocean drag coefficient under drifting pack ice (5.5 × 10⁻³; McPhee, 1980) and slightly higher than the 6.5-m drag coefficient of 4.3 × 10⁻³ observed under MIZ C5 (Cole et al., 2017). Given these variables, sea ice roughness length \((z_o)\) can be obtained from (McPhee, 2008)

\[
\begin{align*}
\text{(3)} \quad z_o &= h e^{-\kappa u_*/\sqrt{u_*}},
\end{align*}
\]

where \( \kappa \) is the Von Karman’s constant (0.4) and \( h \) is the distance from the interface (2.5 m). Roughness length is a parameterization that represents a 30th of the root mean square size of the under-ice roughness elements and defines the hydrodynamic effects that these elements have on boundary layer flow (Nikaradse, 1933). The substitution of results from Equation (2) into Equation (3) results in a roughness length of 1.26 cm, which was the value used in all 1-D turbulence-model control simulations for ice-ocean drag coupling.

Sea ice temperatures and winds were observed from the Vaisala multi-parameter weather station located ~2 m

Figure 2: Observed ice-ocean drag coefficient. Instantaneous drag coefficient \((C_d, \text{Equation } 2)\) observed by the autonomous ocean flux buoy 29 (AOFB 29) turbulence package at MIZ Cluster 5. Black horizontal line indicates mean ± standard deviation (dashed red line) over the 41-day study period. DOI: https://doi.org/10.1525/elementa.391.f2
above the sea ice on AOFB 29, as was the Hukseflux SR03 pyranometer for measuring downwelled solar irradiance. In the ocean, in-situ salinity and temperature profiles were collected from 6–250 m every 3 hours by the ice-tethered profiler 80 (ITP 80) at 1-m resolution (Krishfield et al. 2008; Toole et al. 2011). All temperatures in figures have been converted to temperature above freezing, calculated by subtracting the freezing temperature of the seawater (function of salinity and pressure) from the in-situ temperature \( T = T_o[S,p] \) to estimate available ocean heat. A full sensor schematic of the MIZEX on-ice, in-ice, and under-ice cluster sensors can be found in Figure 2 of Gallaher et al. (2016).

### 3. Modeling framework, validation, and parameter experimentation results

#### 3.1 Turbulent boundary layer model

To examine the sensitivity of the western Arctic upper ocean to mischaracterizations of the air-ice-ocean interface, the local turbulence closure (LTC) model was run with a variety of sea ice conditions. The LTC model parameterizes the development and maintenance of shear-driven instabilities through modification of interfacial energies based on buoyancy conditions and limits imposed by similarity scaling of the non-rotating (surface layer) and rotating portions (Ekman layer) of the ocean boundary layer (McPhee, 2008). The LTC domain occupies 100 vertical levels across the top 60 m of the ocean (0.6-m resolution). Sea ice temperature and ocean salinity and temperature initial conditions were updated at time zero (year day 228 or 16 August) from the ITP and IMB observations, respectively.

All changes in air-ice-ocean momentum and thermodynamic forcing were updated through the LTC interface submodel during each 15-minute time step from the AOFB ice speeds, IMB 2014F thermistor string, and AOFB pyranometer. Kinematic sea ice conductive flux \( \dot{q} \) was calculated by

\[
\dot{q} = \frac{K_w[\Delta T_p - T_o]}{\rho_w c_w h},
\]

where \( T_p \) is the temperature 25% up from the ice base, \( T_o \) is the temperature at the ice base, \( K_w \) is the thermal conductivity of sea ice (~2 W m\(^{-1}\) K\(^{-1}\)), \( \rho_w \) is the seawater reference density (1025 kg m\(^{-3}\)), \( c_w \) is the specific heat capacity of near freezing seawater (3986 J kg\(^{-1}\) K\(^{-1}\)), and \( h \) is sea ice depth. Momentum transfer of sea ice velocities was scaled by the observed roughness length \( z_o = 1.26 \text{ cm} \), Equation 3) to provide the appropriate ice-ocean stresses to the modeled ocean boundary layer. However, this model does not take into account the turbulent kinetic energies generated at the base of the ocean boundary layer (sub-mesoscale eddies and unstable internal waves). The CRREL IMB 2014F time series data provided the appropriate thermal gradients necessary to update conductive heat fluxes to the near-interface heat balance. Solar radiative fluxes were distributed to the water column per an exponential attenuation function with an e-folding depth of 4 m (see McPhee, 2008, for full LTC details). However, before applying the e-folding attenuation function, an estimate of the through-ice radiation must be determined accurately. To achieve this accuracy, solar radiative fluxes through open water \( F_{ovm} \) and through ice \( F_{im} \) were estimated by (Pervich et al., 2007, Stanton et al., 2012; Gallaher et al., 2016).

\[
F_{ovm} = F_{rad} A_{mp} (1 - \alpha_{ovm}),
\]

\[
F_{im} = F_{rad} (1 - A_{ovm}) [A_{mp}(1 - \alpha_{ovm}) h_o + A_{mp} (1 - \alpha_{ovm})]^2,
\]

where \( F_{rad} \) is the observed incoming solar shortwave radiation at MIZ C5, \( I_{ovm} \) and \( I_{im} \) are the bare and ponded ice light attenuation coefficients (Light et al., 2008), \( \alpha_{ovm} \) and \( \alpha_{im} \) are the ocean, bare ice, and melt-ponded ice albedos, and \( A_{mp} \) are the (fractional) areal coverage of open water (OWF), bare ice, and melt ponds. Actual values used in these calculations are provided in Table 1. Fundamental to this approach is the precise calculation of open water, ice, and melt pond fractional areas, which were derived from histogram thresholding of 10 km\(^2\) sections of MEDEA 1-m resolution satellite imagery targeted over the study site (Figure 3). Open water, melt ponds, and sea ice exhibit relatively large differences in gray-scale pixel intensity resulting in tri-modal distributions. Thresholds can be applied to the inflection points of the peaks designating pixel cells into one of the three categories (Kim et al., 2013; Gallaher et al., 2016). Results from this open water, bare ice, and melt pond masking product find corresponding fractional areas of 0.079, 0.263, and 0.658, respectively. Applying Equations 5 and 6, the estimated shortwave radiative transmissivity with parameterization (albedo) error is 0.14 ± 0.0117 or, restated, on average 14% of the incoming solar radiation will pass through the combined open water, ponded ice, and bare ice surface to the underlying ocean.

### Table 1: Sea ice shortwave radiation attenuation coefficients used in Equations 5 and 6. DOI: https://doi.org/10.1525/elementa.391.t1

| Ice type          | Attenuation coefficients* |
|-------------------|---------------------------|
|                   | \( \alpha \) (VIS) | \( \alpha \) (NIR) | \( I_f \) (VIS) | \( I_f \) (NIR) | \( K \) (VIS) m\(^{-1}\) | \( K \) (NIR) m\(^{-1}\) |
| Melting MY ice    | 0.753 | 0.454 | 0.93 | 0.26 | 0.794 | 4.74 |
| Ponded MY ice     | 0.251 | 0.081 | 0.99 | 0.48 | 0.645 | 4.38 |

*Albedo (\( \alpha \)), surface scattering (\( I_f \)), and extinction (\( K \)) coefficients for multi-year (MY) ice and melt ponds for the visible (VIS) and near-infrared (NIR) wavelength bands; all values from Table 4 in Light et al. (2008).
3.2 Boundary layer model validation

To ensure that the model-generated turbulent fluxes properly redistribute upper ocean properties, the LTC model was validated against the following two metrics:

1) proper representation of the near-surface pycnocline depth defined by a 0.1 kg m$^{-3}$ change in density from surface values (Peralta-Ferriz and Woodgate, 2015); and
2) representative LTC model ocean-to-ice heat fluxes with IMB-observed latent heat losses (basal ice melt).

LTC model reproductions of the ocean boundary layer under MIZ C5 compared remarkably well with observed conditions. This agreement is demonstrated in the close rendering of the model mixed layer depth (Figure 4b) with the observed mixed layer depth (Figure 4a) ($R = 0.92$). Similarly, erosion of the near-surface temperature maximum at 30-m depth occurs on nearly the same day of the time series (year day 250, Figure 4c, d). Latent heat losses at the sea ice base also show excellent agreement between

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Figure 3: Surface sea ice characterization from remote sensing. (a) 10 km$^2$ high resolution (1-m) MEDEA visible image over Marginal Ice Zone Cluster 5 (red box) and (b) masked image of surface type classification for open water ($A_{owf}$), melt ponds ($A_{mp}$), and bare ice ($A_{ice}$). Inset values indicate fractional areal coverage. DOI: https://doi.org/10.1525/elementa.391.f3

Figure 4: Boundary layer model validation. Upper ocean (a) observed (ITP-80) and (b) modeled salinity for the 41-day study period with corresponding (c) observed and (d) modeled temperatures above freezing. Red line indicates the ocean mixed layer depth (MLD) defined by the first upper ocean level to be 0.1 kg m$^{-3}$ greater than the surface density (Peralta-Ferriz and Woodgate, 2015). LTC modeled MLD correlated well with the observed MLD throughout the time series ($R = 0.92$). DOI: https://doi.org/10.1525/elementa.391.f4
the LTC-predicted and observed basal ice melt (Figure 5, R = 0.86). Both observed rate of melt (~0.5 cm day\(^{-1}\)) and total ice loss (~22 cm) tracked well with LTC model simulations of basal ice melt. These model results substantiate that the LTC model had representative boundary conditions throughout the study period. Model validation also suggests that the ocean boundary layer was influenced primarily by turbulence generated by ice-to-ocean shear, with limited contributions from lateral (ice-edge form drag) and/or vertical advections.

### 3.3 Sea ice parameter simulation analysis

The variability of late melt-season sea ice conditions have a significant influence on air-ice-ocean interactions. Here, sea ice parameters were intentionally mischaracterized to feature the sea ice components that most impact the ice-affected portion of the upper ocean. The applied mischaracterizations are based on errors generally encountered by field scientists and modelers while attempting to represent the sea ice surface, interior, and base.

#### 3.3.1 Sea ice surface sensitivity

Although the optical properties of sea ice are well understood, the difficulty in defining the integrated solar radiative input into the near-surface ocean stems from the inherent heterogeneity of its surface. Only a handful of tools are capable of capturing this variability. The MEDEA visible imagery used in this study is one of those tools and effectively categorizes both the type and area of various sea ice surfaces; however, due to cloud contamination and small image footprints, only one of eight images collected during the MIZ experiment were usable. Many researchers have relied on synthetic aperture radar (SAR) products to characterize the sea ice with the understanding that the area, and therefore the increased solar radiative input, of ponded sea ice will be neglected. To assess the consequences of this mischaracterization of the sea ice surface, the LTC model was run with surface characterizations from the 100-m resolution RADARSAT-2, 8-m TerraSAR-X SAR imagery, and 1-m MEDEA visible imagery. SAR imagery results show significant underestimations of the areas associated with open water, ice, and melt ponds as compared to the MEDEA image control (Figure 6). Evaluation of LTC model output for these three cases shows no change in mixed layer depth or ocean heat storage between the three cases (not shown); however, basal melt rates were reduced by 29% in the 8-m TerraSAR-X imagery and by as much as 32% in the 100-m resolution RADARSAT-2 imagery (Figure 7). These results demonstrate the considerable regulation surface melt ponds have on through-ice solar radiative fluxes, hence affecting the near-surface heat balance due to reduced turbulent heat fluxes near the ice-ocean interface.

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**Figure 5: Model validation of the ice-ocean interface heat budget.** Color-contoured sea ice temperatures from the CRREL ice mass balance (IMB) 2014F along with ice surface (white line) and ice base observations (black dots). Temporary malfunction of the IMB 2014F acoustic ranging sensor between deployment and 17 September required fusion of the IMB 22 data (light magenta dots) with IMB 2014F data. IMB 22 was located 10 km to the west of MIZ C5 and deployed in sea ice 0.5 m deeper than IMB 2014F, hence the +0.5 m adjustment (dark magenta dots) to the IMB 22 data on the figure. Blue dashed line is the LTC model ice base which correlated well with IMB observations (R = 0.86). DOI: https://doi.org/10.1525/elementa.391.f5
3.3.2 Sea ice interior sensitivity

Similar to the sea ice surface, conductive heat fluxes through the sea ice have been studied thoroughly (Untersteiner, 1961; Maykut, 1978); however, the challenge is the availability of internal sea ice temperature gradients due to the demanding equipment requirements of deploying sensors in decaying sea ice. Fortunately, MIZ C5 had two IMB time series to characterize the late summer thermal transmissivity conditions.
The temperature gradients between the on-site (IMB 2014F) and off-site (IMB 22) ice mass balance thermistor strings shared similar magnitudes; therefore, IMB 2014F was used as the control for this sensitivity test (Figure 5). To extract potential impacts that internal thermal properties have on the ocean boundary layer, thermal gradients were tested under weaker \(0.5 \times \nabla_z T\) and stronger \(2 \times \nabla_z T\) than observed conditions. Results of the LTC model demonstrate that mischaracterization of vertical profiles of temperature have little effect on either boundary layer penetration or basal melt rates (Figure 8). This outcome with such a large adjustment in temperature gradients suggests that conductive heat fluxes are a second or third order effect on near-interface thermodynamics and, therefore, on upper ocean buoyancy fluxes.

### 3.3.3 Sea ice base sensitivity

Interactions between the sea ice base and the upper ocean are complex and difficult to resolve observationally. After four decades of research on ice-ocean interactions, estimates of basal ice roughness elements can vary by an order of magnitude depending on ice age, formation conditions, and deformation forcing (Lu et al., 2011). Secondary, local ocean turbulent mixing can be influenced by floe-edge form drag, near-by keels, and melt pond cavities. For these reasons, ice-ocean drag may be the most commonly mischaracterized parameter of sea ice. Therefore, to appreciate the full range of under-ice influences on the ocean boundary layer, we evaluated ice-ocean interactions under relatively rough ice to very smooth ice. To set constraints, only drag coefficients previously observed in the western Arctic are used. For rough ice, the model roughness length is set to 5.39 cm based on a drag coefficient of \(9.94 \times 10^{-3}\) (Equation 3) observed under deformed multi-year ice during the 1983 MIZEX-West campaign (Bruno, 1990). For smooth ice, the roughness length is set to 0.017 cm based on a drag coefficient of \(1.0 \times 10^{-3}\) at Cluster 2 during MIZEX 2014 under a conglomerate of first-year and multi-year sea ice in the Beaufort Sea (Cole et al., 2017).

Results of the under-ice sensitivity simulations show substantial changes to near-surface pycnocline depths. For smooth ice, mixed layer depths shoaled by 8 m, and the heat under 23 m was preserved throughout the 41-day time series (Figure 9a, d). Conversely, ice-ocean interactions under rough ice deepen the mixed layer considerably, adding 6 m of isohaline/isothermal conditions and mixing completely through the early summer near-surface temperature maximum just 10 days into the experiment (Figure 9c, f). The response of latent heat losses (proxy for ocean-to-ice heat fluxes) were less dramatic, with cumulative basal ice melts changing only \(\pm 14\%\) (~3 cm) (Figure 10). These results suggest that under-ice roughness conditions have a substantial influence on the intensity of ocean boundary layer turbulent eddies and the resulting redistribution of upper ocean buoyancy gradients.

### 4. Discussion

Results of the sensitivity study show a clear prioritization of which sea ice properties most affect the ice-ocean boundary layer in late summer. The order from largest affect to smallest was under-ice roughness, sea ice surface type/coverage, and internal sea ice temperature gradients. The order of the first two are somewhat surprising, but after further investigation this sequence was logical given the late summer forcing. In late summer, solar radiative input is well off the summer solstice and has diminished substantially, suggesting that this prioritization might be different during an early summer (surface first, roughness second) test case. During winter, internal sea ice temperature gradients likely dominate, as (negative)
sensible heat fluxes modify ice temperature from the top down. Therefore, and reiterating, these findings/rankings are only valid between the melt and freeze seasons when solar angles are low and sea ice temperatures are still relatively high.

This study highlights the importance of identifying representative ice-ocean drag coefficients during late summer, particularly as many modelers still use a single coefficient to parameterize system momentum fluxes during all conditions. However, recent work by Cole et al. (2017)
demonstrates how drag coefficients fluctuate by an order of magnitude in the same geographic region and seasonally at the same location. Factors affecting regional variability depended on fractional size of embedded multi-year and first-year sectionals. For seasonal variations, drag coefficient changed with ice concentration, boundary layer stratification, and mixed layer depth.

The high correlation between the study boundary layer model and upper ocean observations were encouraging but likely benefited from the “steady state” condition at MIZ C5 during the selected 41-day study period. Unlike the MIZ observation sites in Cole et al. (2017), MIZ C5 remained within the continuous ice cover and well outside of the marginal ice zone, maintaining a consistent sea ice concentration >90% according to RADARSAT-2 images (Figure 1). Additionally, the continuous ice cover helped to maintain the orientation of both smooth and ridged ice floes near MIZ C5, making regional drag less variable. Finally, lower sun angles, limited meltwater, and cooler ice temperatures destabilized the boundary layer, allowing turbulent eddies to deepen routinely past the AOFB turbulence package at 2.5 m, providing reliable roughness measurements from observed friction velocities. Ideally, boundary layer scientists would like to study turbulent drag under all conditions, requiring investment in autonomous 3-D acoustic current meters (turbulence) and salinity (stratification) sensors stationed less than 3 m from the ice base (Cole et al., 2017; Gallaher et al., 2017). Such observations would still not resolve the large spatial heterogeneity of sea ice morphology, however, a combination of these shallow flux instruments and large-scale mapping of under-ice topography (by glider technology, NASA IceBridge flights, NASA ICESat imagery) could provide important regional bulk characterization of ice bottom roughness, offering climate modelers considerable improvements to the near-surface energy balance.

Basal melt rate responses to changes in under-ice roughness were notably muted. For smoother ice, basal melt rates did not change substantially, because a shallower boundary layer still has access to the primary source of heat used to convert the ice base to meltwater – downwelled solar radiation absorbed just beneath the sea ice. For rougher ice, in substantial change is likely due to the limited reservoir of available heat contained in the near-surface temperature maximum. Just 7.2 MJ m⁻² of heat were stored in the 7 m of water ventilated by the stronger mixing associated with the rougher ice test case (7 m³ 3986 J K⁻¹ kg⁻¹ 1025 kg m⁻³ 0.25 K); or in latent heat terms, ~3 cm of basal ice melt. Basal ice melt is a proxy for ocean-to-ice turbulent heat fluxes; however, the bulk of the heat available for turbulent heat flux comes from the very near-surface ocean where solar radiation deposits most of its energy (Frey et al., 2011). Heat stored in the early summer near-surface temperature maximum is generally inaccessible due to the strong summer halocline supported by the abundance of late melt season buoyancy. Although solar zenith has descended since the solstice, shortwave radiation integrated over the 41-day time series is still the largest contributor of heat to ice–ocean interactions during late summer. Therefore, errors associated with mischaracterizations of the sea ice surface have significant penalties, evidenced by the 32% reduction in basal ice melt when commonly used SAR imagery mischaracterizes the sea ice surface, eliminating solar radiative pathways through ponded ice. However, an alternative to visible high-resolution imagery is desirable given the frequent loss of surface reflectance overcome by cloud contamination and the difficulty of targeting observation sites with 15 km² scan windows. The all-weather SAR product is less prone to these limitations; however, previous studies have shown that exclusion of melt ponds from shortwave radiative parameterization of the sea ice surface can result in errors as large as 23% (Gallaher et al., 2016), highlighting the need for robust algorithms to identify melt ponds.

5. Conclusions

This study explored the relative importance of sea ice parameters in modulating late season sea ice–ocean interactions. Thermal gradients in late summer sea ice are weak and therefore had virtually no influence on ice base melt or boundary layer depth. Of all of the model test cases, changes in ice basal roughness showed the most substantial influence on turbulent penetration, with boundary layer mixing depths shoaling by 26% underneath smooth ice and deepening by 21% under more deformed sea ice. Of the two, the smooth ice case likely carries the greater consequence, given that heat stored in the early summer near-surface temperature maximum is preserved due to less vigorous turbulent mixing. This scenario facilitates the ventilation of upper ocean stored heat to the ice during boundary layer destabilization in the fall and would potentially delay ice production despite strongly negative conductive heat fluxes (Timmermans, 2015). At the surface, three different sensitivity tests were conducted using visible and SAR imagery. As expected, the omission of melt ponds by the SAR imagery resulted in nearly a third less basal melt; however, the boundary layer responses were minimal due to the limited solar shortwave radiation available during late summer.

Perhaps the most important finding emerged from the comparisons of under-ice roughness treatments between boundary layers under marginal ice zones and boundary layers beneath non-seasonal sea ice. The LTC model validated observations remarkably well during the 41-day time series, but mostly because consistent roughness and neutral buoyancy prevailed at MIZ C5. Comparison of these results with the work of Cole et al. (2017) in the MIZ made clear that similar model validation within the ever-changing seasonal ice zone would not have been probable given the regional and seasonal variability of ice–ocean drag coefficients in the MIZ. This clarity suggests that modelers should select a “mode” for assigning drag parameterizations to spatial and temporal domains. For example, the universal drag coefficient (~5.5 × 10⁻³; McPhee, 1980) would have been suitable in the continuous sea ice at MIZ C5. Furthermore, the high correlation coefficients suggest that the influence of submesoscale features and internal gravity wave instabilities were not significant on the pack ice side of the marginal ice zone. Seasonally, this “mode”
works best during late summer and conceivably through mid-spring when the boundary layer is neutral or unstable. During the alternate mode (MIZ and/or melt season), more work is needed to gain an approach for drag characteristics under melting sea ice and seasonal ice zones. The results of this study show that forcing in the upper western Arctic remains largely local; however, this may change in future decades as the seasonal ice zone continues to expand.

Data Accessibility Statement
Electro-optical (MEDEA) imagery was made available by the U.S. Geological Survey Global Fiducial Library Data Access Portal (http://gfl.usgs.gov). IMB 2014F data were provided by the Cold Regions Research and Engineering Laboratory (http://imb/erci/dren.mil) and IMB 22 data were provided by Scottish Association of Marine Science. The ice-tethered profiler data were collected and made available by the Ice-Tethered Profiler Program (Toole et al., 2011; Krishfield et al., 2008) based at the Woods Hole Oceanographic Institution (https://www.whoi.edu/ipt). The MIZ experiment data were consolidated and made available by the University of Washington Applied Physics Labs (http://www.apl.washington.edu/project/project.php?id=miz).

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Competing interests
The author has no competing interests to declare.

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