Inside Katabatic Winds Over the Terra Nova Bay Polynya: 2. Dynamic and Thermodynamic Analyses

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Abstract  Surface fluxes and atmospheric boundary layer budgets of enthalpy and momentum are quantified using bulk and integral methods based on measurements obtained during a research vessel transect across the Terra Nova Bay polynya during an intense late autumnal katabatic wind event. The surface sensible, latent, and net radiation heat fluxes had maximum upward values of 2,500 ± 600, 400 ± 100, and 100 ± 7 Wm$^{-2}$, respectively, which occurred over open regions that had been cleared of sea ice by the wind stress. In these open areas, sea spray enhanced the sensible and latent heat fluxes (LHF$s$) by an estimated 106% and 18%, respectively. As sea ice formed on the surface and became thicker in the downwind direction, the sensible and LHF$s$ decreased to 50% of their maximum values over pancake ice and to 5% at the downwind end of the transect, where snow-covered young ice flows were present. Snow growth removed over 50% of the water vapor that came from the surface. The surface wind stress ranged from mean values of 2.9 Nm$^{-2}$ in the most upwind regions to 0.9 Nm$^{-2}$ at the end of the transect, with smaller scale variations of almost a factor of three due to different surface types (roughnesses) and gustiness. The downwind slowing and turning of the wind vector can be almost entirely explained by frictional and inertial forces, indicating the horizontal pressure gradient was weak. This case could serve as a case study for future modeling studies of coastal polynyas.

Plain Language Summary  In certain regions of coastal Antarctica cold air flows down valleys like a huge river, sometimes reaching speeds as great as in a hurricane. When this river of air flows out over the ocean, it pushes the ice back away from the coast, creating areas of open water called “coastal polynyas.” In the winter, the air is much colder than the ocean surface, which means that the strong winds carry away tremendous amounts of heat from the surface of the polynyas, which causes large amounts of sea ice and dense salty ocean water to form. The goal of this research was to directly measure the amount of heat that was lost over a polynya in the Terra Nova Bay while it was being transited by the research vessel R/V Nathaniel B. Palmer. We found that due to the strong winds, cold air, and contribution of sea spray from breaking and wind-sheared waves, the measured heat loss from the surface of the polynya was one of the largest measured anywhere on Earth. Scientists will use this information to understand how the processes that occur in coastal polynyas can affect deep ocean circulation and global climate.

1. Introduction

Antarctic coastal polynyas (ACPs) occur where air is funneled into valleys and ejected horizontally over adjacent ocean regions, pushing any newly formed sea ice away from the coast. During cold seasons, the exposure of the relatively warm ocean surfaces to the sometimes hurricane-force winds causes great amounts of enthalpy (sensible and latent heat energy) to be lost from the surface of the polynyas, resulting in sea ice formation (e.g., Gordon & Comiso, 1988; Z. Zhang et al., 2015), and brine rejection (Mathiot et al., 2012) which can lead to the creation of deep ocean water masses that circulate throughout the world ocean (Orsi & Wiederwohl, 2009; Rusciano et al., 2013). Although the surface areas of ACPs are relatively small (e.g., Bromwich & Kurtz, 1984), their potential impacts on regional and global climate are significant due to the large amounts of sea ice and dense water that are created within them (Fusco et al., 2009; Gordon, 2001) and the injection of heat and moisture into the atmosphere (Renfrew, 2002).

Due to their small size, ACPs are not resolvable by the general circulation models (GCMs) that are used to understand and predict future climate change. Therefore, to quantify the potential impacts of ACPs on global climate and regional ice production, their enthalpy loss and related effects must be simulated using sub-grid scale parameterizations. Typically, these parameterizations of surface fluxes are based on bulk
methods, which use measurements or model simulations of the mean atmospheric state variables (wind vector, temperature, humidity, pressure) and surface parameters (temperature, humidity, roughness) to estimate the surface fluxes of enthalpy and momentum. The bulk method requires the specification of the transfer coefficients, which relate the state and surface variables to the fluxes. The only published measurements of heat fluxes and transfer coefficients over an ACP are from Knuth and Cassano (2014) who measured values of 12–485 W m⁻² for sensible heat flux (SHF) and 56–152 W m⁻² for latent heat. Flux parameterizations designed for short fetch lead and polynya situations (e.g., Andreas & Cash, 1999) and over heterogeneous ice surfaces (Schröder et al., 2003) are applicable for some ACP situations. However, no previous polar studies have attempted to analyze the impact of sea spray on surface fluxes and transfer coefficient values. This study will approach that problem by using studies based on tropical cyclones (TCs) to provide guidance on expected effects of sea spray in high-wind, open-water ACP situations.

The Polynya and Ice Production and seasonal Evolution in the Ross Sea (PIPERS) field program (Ackley et al., 2020) provided an opportunity to develop and verify quantifications of transfer coefficients and atmospheric boundary layer physics for use over ACPs. On May 8, 2017, as part of the PIPERS program, the R/V Nathaniel B. Palmer steamed on a downwind transect across the Terra Nova Bay polynya (TNBP) during a strong katabatic wind event, while the author and colleagues performed surface layer (ship platform) and upper-air (rawinsonde) measurements. The physical setting, measurements, surface characteristics and details regarding the vertical structure and horizontal evolution of the state variables wind vector U, the potential temperature θ, and the specific humidity q, for this case are presented in Guest (2021), hereafter referred to as “Part 1.” The figures and results in Part 1 will be referred to frequently throughout this study.

During the 8 May PIPERS transect of the TNBP (in this study referred to as “this case”) the author and associates were able to perform four successful rawinsonde profile measurements of the state variables at locations 17, 54, 84, and 99 km downwind from the Nansen Ice Shelf edge (Part 1, Figures 1 and 2). There were also continuous measurements of the state variables at the Manuela automatic weather station (AWS), which was located 1 km upwind of the Nansen Ice Shelf edge (Lazzara et al., 2012). Continuous observations of surface conditions and measurements of the state variables in the atmospheric surface layer were performed as the ship steamed between locations P1 and P4. In this study, the downwind distance from the ice shelf edge is referred to as “x.” The value of x was up to 3% larger than the actual physical distance from the ice shelf edge due to the curvature of the air parcel trajectories.

This case was characterized by a strong low-level jet, with maximum wind speeds of 30 m s⁻¹ occurring at approximately two-thirds of the atmosphere boundary layer (ABL) height and decreasing from there up to ~700 m elevation, at which point the wind vector became light and variable (Part 1, Figure 4a). Well-defined temperature inversions existed above mixed layers with bases at 294, 325, 226, and 196 m elevation for P1–P4. These are defined as the ABL depths and indicated by the variable z. Although the jet (U profile) and boundary layer structure (θ and q profiles) remained qualitatively similar for all locations, there was a significant decrease in wind speed U, turning in wind direction WD, and increases in θ and q in the downwind direction in response to the dynamic and thermodynamic forcing for this case.

This case represents a unique opportunity to understand air-sea interactions during extreme wind events. By using the R/V Palmer as the primary measurement platform, we were able to sample surface and near-surface properties that would not have been possible using in situ aircraft (unflyable conditions), dropsondes (no surface information), or satellites (cloud and aerosol obstruction). Most knowledge regarding air-sea interactions in high wind conditions has come from studies of TCs, and a small number of polar lows. These phenomena occur over the open ocean, where waves, spray, and wind blast would endanger the survival of any research vessel, let alone allow performing any of the measurements used for this case (Part 1, Table 1). However for this case, the relatively short fetches of open ocean and the suppression of waves and spray by frazil and solid sea ice allowed surface layer and upper-air measurements to be performed that would have been impossible in virtually any other marine situation with hurricane-force winds. Therefore, the results of this study, in particular with respect to sea spray effects, in addition to being relevant for katabatic winds and ACPs, may also be applicable to other high wind marine (polar lows, TCs) situations, where direct in situ surface and surface layer measurements are not possible.
This study is an analysis of the thermodynamic and dynamic forcing that occurred within the katabatic wind jet over the TNBP. Although this study focuses solely on this one case, it was similar to the two other polynya transect cases performed during PIPERS (results not shown), the September 19, 2012 case from Wenta and Cassano (2020), and the cases described by Parish and Bromwich (1989), and was likely to have had many of the same characteristics as the lower atmosphere in other ACP situations. Following this introduction, Section 2 describes the theory and methods used to perform the thermodynamic analyses. Section 3 presents the results of these methods as applied to this case. Sections 4 and 5 provide a similar analysis of the dynamics for this case and Section 6 concludes the study with a summary and discussion.

2. Thermodynamics Theory and Methods

2.1. Thermodynamics Introduction and Simplifications

There are two main goals in analyzing the jet thermodynamics: (a) quantifying the sensible heat and moisture surface fluxes (together the “enthalpy flux”), and (b) understanding the heat and moisture budgets within the jet ABL. This section examines the theory used to understand the jet thermodynamics; the next section presents the results from this case.

The design of the cruise plan and the synoptic situation for this case allowed some simplifications that facilitated the thermodynamic analysis. The Palmer steamed directly downwind as quickly as possible (stopping only for rawinsonde launches) during the 12-hr time period, 0000 to 1200 May 8, 2017 UTC, when atmospheric and surface conditions did not change significantly (Part 1, Figures 1 and 2). The ABLs for this case were well mixed, with generally constant (except for turbulent fluctuations) wind direction WD, potential temperature $\theta$, and specific humidity $q$, below capping temperature inversions with distinct bases at elevation $z_i$ (Part 1, Figures 4c and 4d). In this situation, $U$, $\theta$, and $q$ could be represented by single mean ABL values $U_{ABL}$, $\theta_{ABL}$, and $q_{ABL}$, rather than as functions of elevation. The mean values were calculated using the procedures described in Part 1. The inversion base height $z_i$ was assumed to be the top of the ABL. By having the ship track air parcel trajectories/streamlines and assuming steady-state conditions, the real-world four-dimensional domain was transformed into a much simpler one-dimensional domain, where the only independent variable was time $t$, which is equivalent to downwind distance from the ice shelf $x$, when multiplied by the average ABL wind speed $U_{ABL}$. While not a perfect representation of reality, the one-dimensional assumption was adequate to allow reasonably accurate quantifications of the surface fluxes and ABL enthalpy budget.

Because steady-state conditions were assumed for this case, the ABL mean quantity time derivatives shown in this study were determined based on the measured spatial derivatives

$$\frac{d}{dt} = U_{ABL} \frac{\partial}{\partial x}$$

where $x$ represents the distance along the slightly curving parcel trajectories in the downwind parcel trajectory direction.

An appropriate thermodynamic state variable for this case is the ice static energy, $H_I$, defined as

$$H_I = C_p T_{air} + L_s q + gz$$

where $C_p$ is the specific heat capacity of air at constant pressure, $T_{air}$ is the air temperature, $L_s$ is the specific heat of sublimation, $g$ is the gravitational constant, and $z$ is elevation. The ice static energy is analogous to the moist static energy. However, because $H_I$ is conserved during vapor/ice phase changes, it is more appropriate for this cold case. For this study, the potential temperature $\theta$ is defined as

$$\theta = T_{air} + \frac{gz}{C_p}.$$ 

This sets $\theta$ equal to $T_{air}$ at the surface, which clearly emphasizes the magnitudes of the important air-surface temperature differences. Setting the variables in Equation 1 equal to the mean ABL values and using the $\theta$ definition defines an ABL mean specific enthalpy, $H_{ABL}$

$$H_{ABL} = C_p \theta_{ABL} + L_s q_{ABL}$$

which is conserved during vapor/ice phase changes.
2.2. ABL Total Enthalpy Budget

Using the one-dimensional assumption, the First Law of Thermodynamics for air parcels in the ABL can be expressed as a budget equation

\[ \rho_{ABL} \frac{dH_{ABL}}{dt} = \text{SHF} + LHF_A - LW \frac{\Delta p}{\rho_{ABL}} + E_T - E_q + \text{other} \]

where \( \rho_{ABL} \) is the mean ABL air density and \( d / dt \) is the time derivative following air parcels in the ABL. The right-hand-side (RHS) of Equation 4 contains the forcing terms and includes the surface turbulent SHF, the surface turbulent latent heat flux (LHF) \( \Delta p / \rho_{ABL} \) (defined in terms of sublimation), the radiational divergence in the ABL \( LW \frac{\Delta p}{\rho_{ABL}} \), and the entrainment of sensible and latent heat into the ABL from above \( E_T \) and \( E_q \). “Other” includes the effects of precipitation from above the ABL, frictional heating, and other effects that were not significant compared to the uncertainties of the other terms and the one-dimensional assumption.

The term “latent heat” can be confusing because its definition depends on which water phase changes are involved. For this study, LHF \( \Delta p / \rho_{ABL} \) is defined as the surface flux of \( L \) \( q \), the subscript “A” indicating that this is the more relevant enthalpy flux component for the atmosphere for this cold air case. The symbol “LHF” without the subscript, refers to the surface flux of \( L \) \( q \) where \( L_q \) is the latent heat of evaporation. The latter is the traditional definition of LHF and represents the immediate enthalpy loss from the ocean. LHF \( \Delta p / \rho_{ABL} \) is 13.5% greater than LHF, representing the extra heat obtained from sublimation versus evaporation. The following subsections will describe the interaction between the sensible and latent components of \( H_{ABL} \) (Equation 4) followed by descriptions of the methods used to estimate the values of each of the terms in Equation 5.

2.3. ABL Moisture Budget and Precipitation

During this case, the specific humidity \( q \) was always equal to the saturation value with respect to ice \( q_{sat} \) in the surface layer at elevation of 15 \( \pm \) 8 m (Part 1). This has major implications for understanding the katabatic jet thermodynamics, in particular, the moisture budget. Liquid water at the surface or as suspended droplets had a much higher surface vapor pressure than the ice saturation value. Therefore, where liquid water was present, the surface flux of humidity was acting to drive the atmospheric humidity higher than ice saturation. Some process such as ice crystal growth and fallout must have been removing moisture from the ABL during this case. The following paragraphs will examine the ABL moisture budget to understand implications of the \( q \) equal to \( q_{sat} \) phenomenon.

The change in \( q_{ABL} \) of a parcel in the ABL is

\[ \frac{dq_{ABL}}{dt} = \frac{Q_{vapor}}{z_i \rho_{ABL}} \]  

where \( Q_{vapor} \) (kg m\(^{-2}\) s\(^{-1}\)) represents all ABL sources/sinks of water vapor. By conservation of water, precipitation \( P_r \) must be the difference between the other sources/sinks of water vapor (surface fluxes and entrainment) and the change in the total amount of water vapor in the ABL, \( Q_{vapor} \)

\[ P_r = \frac{LHF_A}{L_s} - \frac{E_q}{L_s} - Q_{vapor} \]  

where the surface moisture flux, \( LHF_A / L_s \) was a source, and moisture entrainment, \( E_q / L_s \) was represented as a (negative) sink because the air above \( z_i \) was usually drier. For most of the following discussion, the entrainment, \( E_q \) is assumed to be negligible; however, the term is kept in the equations to understand how nonzero entrainment could potentially affect the interpretation of the results. Multiplying by \( L_s \) and substituting for \( Q_{vapor} \) using Equation 5 results in

\[ H_p = LHF_A - E_q - \rho_{ABL} L_s z_i \frac{dq_{ABL}}{dt} \]  

where \( H_p \) is the enthalpy that is converted from latent heat \( L \) \( q \), to sensible heat \( C_p T_{air} \) due to precipitation.

The mean value of the last term on the RHS of Equation 8, containing the \( q_{ABL} \) time derivative, could be estimated for this case for each of the intervals between rawinsonde launch locations using the upper-air \( q \) measurements. However, the rawinsonde data alone cannot distinguish how much of the change in \( q_{ABL} \) was due to precipitation \( P_r \) versus surface moisture flux, \( LHF_A / L_s \). Therefore an independent estimate of \( LHF_A \) such as from the continuous ship-platform measurements, must be used to determine the value of
H_p. In fact, as will be shown following, for this case, H_p was able to be estimated based on ship measurements alone, with no direct in situ upper-air information needed.

It was observed that changes in q_{ABL} were matched by changes in surface layer specific humidity q_{SL} (Part 1):
\[
\frac{\partial q_{ABL}}{\partial x} = \frac{\partial q_{SL}}{\partial x}, \quad \therefore \quad \frac{dq_{ABL}}{dt} = \frac{dq_{SL}}{dt}.
\]  
(9)

Because q_{SL} was always close in value to q_{sati},
\[
\frac{dq_{SL}}{dt} = \frac{dq_{sati}}{dt} = \frac{\partial q_{sati}}{\partial t}
\]  
(10)

where \(\partial q_{sati}/\partial T\) is the change in ice saturation q as a function of temperature, a known mathematical relationship (Buck, 1981). For this case, the potential temperature in the surface layer \(\theta_{SL}\) was always very close to \(\theta_{ABL}\) (Part 1, Figures 3c, 3d, 4c and 4d)
\[
\theta_{SL} = \theta_{ABL}
\]  
(11)

The warming of the ABL was caused by all the sensible heat sources, \(F_{sens}\)
\[
\frac{d\theta_{ABL}}{dt} = \frac{F_{sens}}{\rho_{ABL} C_{p} \theta_{i}}
\]  
(12)

Applying Equations 8–12 results in
\[
H_p = LHF_A - E_q - \frac{L_{v}}{C_{p}} F_{sens} \frac{\partial q_{sati}}{\partial t}.
\]  
(13)

The last term on the RHS represents the enthalpy of the change in the ice saturation humidity in the ABL due to \(F_{sens}\), that is, the extra amount of q storage available due to warming. The precipitation enthalpy \(H_p\) represents the transfer from latent heat to sensible heat due to vapor to ice phase changes. For simplicity, this process will hereafter be referred to as “snow.” However, any process that creates ice crystals from water vapor and is then removed, such as deposition on frozen sea spray or existing hydrometeors, is also included in \(H_p\) because, thermodynamically, these processes are indistinguishable from snow growth and fallout.

The ABL sensible heat source term \(F_{sens}\) consists of the surface SHF, radiation convergence, LW_{ABL}, entrainment, \(E_{\theta}\), and sensible heat gained due to snow, \(H_p\). Therefore
\[
H_p = \frac{LHF_A - E_q - \left( SHF + E_{\theta} + LW_{ABL} + H_p \right) \frac{L_v}{C_p} \frac{\partial q_{sati}}{\partial t}}{C_{p}}.
\]  
(14)

Note that while \(H_p\) can be solved explicitly, in this form of the equation, it appears on both sides to highlight the negative feedback effect due to the increase in temperature resulting from the snow.

This subsection showed that although horizontal changes in q_{SL} or q_{ABL} cannot distinguish between the effect of surface moisture flux versus snow, the precipitation enthalpy, \(H_p\) (and associated snowfall rate), can be specified entirely from estimates of the SHF and LHF_{A} (with corrections for radiation and entrainment). Therefore, because the continuous ship data were used to estimate the fluxes using the bulk method, described below, continuous snowfall estimates were also possible; no knowledge of ABL depth or aerosol microphysics was required to make these estimates.

2.4. Steam Fog

The lowest \(~30\) m of the atmosphere in the more open ocean areas (primarily at \(x < 40\) km) contained steam fog, which tended to be concentrated in small (1–6 m diameter) vortex swirls. The steam fog reduced visibility and immediately froze upon hitting any surface, thus contributing to vessel icing. Steam fog could potentially have significant effects on radiation at the surface and within the ABL, especially in otherwise clear conditions. However, for this case, sea spray, hydrometeors (snow), and upper level clouds were also present; therefore it was difficult to detect, let alone quantify, any steam fog effect on the radiation.

All the steam fog that eventually evaporated in the ABL had no net effect on the terms in enthalpy budget terms in Equation 5. It was not known how much of the steam fog froze. However, any frozen steam fog that falls out will have virtually the same effect as snow (direct vapor to solid conversion) and would have been included in the precipitation calculation described above.
The surface layer temperature and humidity profiles were altered slightly due to steam fog formation and evaporation, which would change the surface layer stability. However, in this high wind environment, the resulting effects on the surface fluxes were negligible and will not be included in the bulk surface flux calculations described later.

2.5. Radiation

For this case, four sources of information were used to quantify radiation effects: (a) radiation flux measurements from the ship helicopter deck, (b) surface IR temperature measurements from the ship boom, (c) upper-air temperature and humidity profiles from the rawinsondes, and (d) human observations of clouds, weather and visibility (as described in Supporting Information S1). To facilitate the interpretation of the radiational fluxes and relate them to air and surface temperatures, a brightness temperature $T_{\text{bright}}$ is defined as

$$T_{\text{bright}} = \left( \frac{LW}{\sigma} \right)^{\frac{1}{4}}$$

where LW is any longwave radiation flux component and $\sigma$ is the Stefan-Boltzmann constant. Using a one-layer model, the net longwave radiation convergence $LW_{\text{ABL}}$ for the ABL was estimated by

$$LW_{\text{ABL}} = \epsilon_{\text{ABL}} \sigma \left( T_{\text{sky}}^4 + \epsilon_{\text{df}} T_{\text{df}}^4 + (1 - \epsilon_{\text{df}}) T_{\text{df}}^4 \right) - 2T_{\text{ABL}}$$

where $\epsilon_{\text{ABL}}$ was the bulk ABL emissivity, $\epsilon_{\text{df}}$ was the surface emissivity (assumed to be 0.98 for all surfaces), $T_{\text{sky}}$ was the brightness temperature of the radiation coming into the top of the ABL, $T_{\text{df}}$ was the surface temperature, $T_{\text{ABL}}$ was the mean ABL temperature, and $T_{\text{df}}$ was the brightness temperature of the downward longwave radiation measured from the ship. The latter was used to calculate surface reflected radiation, for example, Vihma et al. (2009). The upper sky brightness temperature $T_{\text{sky}}$ was estimated based on the rawinsonde temperature and humidity profiles, in particular, the temperature at the bottom of the nearest cloud base to $z_i$. For $x$ locations between rawinsonde measurements, the value of $T_{\text{sky}}$ was linearly interpolated. The average boundary layer temperature $T_{\text{ABL}}$ was directly measured at the rawinsonde locations (P1–P4). However, it could also be estimated for the in-between locations by using the 21 m potential temperature $\theta_{21}$, as a proxy for the potential temperature throughout the ABL. The assumed adiabatic lapse rate (Equation 3) could then be used to estimate the mean ABL temperature profile and the mean temperature $T_{\text{ABL}}$. The ABL depth $z_i$ needed for this calculation was linearly interpolated (in $x$ space) between the nearest rawinsonde profiles.

The bulk emissivity of the ABL $\epsilon_{\text{ABL}}$ in the one layer model was

$$\epsilon_{\text{ABL}} = \frac{T_{\text{df}}^4 - T_{\text{ABL}}^4}{T_{\text{ABL}}^4 - T_{\text{sky}}^4}$$

At locations $x < 35$ km, the $\epsilon_{\text{ABL}}$ derived from Equation 17 was noisy because the value of $T_{\text{ABL}}$ was near $T_{\text{sky}}$ due to a low warm cloud in the inversion layer. Also in these $x$ locations, the mean $\epsilon_{\text{ABL}}$ value determined using Equation 17 was equal to an unrealistic value of 1.1 because large amounts sea spray, hydrometeors, and steam fog in the lower part of the ABL (where $T_{\text{ABL}}$ was warmer than $T_{\text{ABL}}$) were the sources of most of the radiation detected at the ship level. To address these limitations in the one-layer radiation model, $\epsilon_{\text{ABL}}$ derived using Equation 17 was divided by 1.1.

There was enough shortwave radiation to allow some visibility of objects beyond the ship lights. However, the measured downward values were never higher than 2 Wm\(^{-2}\); therefore solar radiation effects will be ignored for this case.

2.6. Surface Heat and Moisture Fluxes Estimation Methods

The surface heat and moisture fluxes (collectively “enthalpy flux”) were the dominant factors controlling the katabatic jet thermodynamic and ABL enthalpy characteristics for this case, and certainly other similar cold season situations as well. Due to the (potentially global) impacts of surface fluxes over ACPs, the most important goal of this research was to quantify the heat and moisture fluxes using in situ, rather than...
modeled or inferred data. There are five potential methods that can be used to estimate surface enthalpy fluxes based on in situ measurements: (a) eddy correlation, (b) inertial-dissipation, (c) profile, (d) bulk, and (e) integral. All five methods were employed at times during the PIPERS cruise. However, during this case, the icing conditions were too severe to perform the turbulent measurements required for methods (a) and (b). The profile method (c) was applied based on the difference between the 9 and 21 m temperature and humidity measurements on the Palmer. The results were generally consistent with the bulk flux estimation methods described below, and demonstrated that there were no systemic biases in the measurements at these levels. However, the results were very noisy due to ship movements and flow distortion effects; therefore they will not be considered any further in this analysis. The bulk (d) and integral (e) methods proved to be successful for this case and will be described next in the two following subsections. The results of these methods for this case are discussed in Section 3.

2.7. Bulk Method

2.7.1. Introduction

The bulk method estimates surface fluxes based on measurements of $U$, $\theta$, and $q$ in the surface layer and the surface temperature $T_{sfc}$. An advantage of the bulk method is that it is not particularly sensitive to ship-induced flow distortion or turbulence-generation issues. Another advantage for this case is that due to the large air-surface temperature differences, the relative effects of any temperature measurement errors on the enthalpy flux estimates were minimal. A disadvantage is that the bulk method requires pre-specification of roughness parameters that were not directly quantified for this case and must be estimated based on previous research.

A factor that needed to be considered for this case was the large amount of sea spray that was generated by the high winds over the open water areas. Considerable recent research (e.g., Garg et al., 2018; Ortiz-Suslow et al., 2016; T. Zhang & Song, 2018) has shown that sea spray contributes significantly to the surface enthalpy fluxes in high-wind TCs, and it is reasonable to expect that it would have similar effects over open ocean areas during this case. We follow the approach of Andreas (1992) and separate the surface enthalpy fluxes into two components, (a) interfacial and (b) sea spray. Andreas assumes that the interfacial flux component is only controlled by turbulent processes in the surface layer and can be estimated using Monin-Obukhov similarity theory (MOST), while the spray flux component depends on droplet microphysics. The next subsection will describe how the interfacial component of the enthalpy flux was estimated for this case, followed by a sea spray component subsection.

2.7.2. Bulk Method Interfacial Fluxes

The surface interfacial enthalpy fluxes $SHF_{interfacial}$ and $LHF_{interfacial}$, and the momentum flux $\tau$ for this case were calculated using the following formulas

$$SHF_{interfacial} = \rho c_p C_{HZ} U_{33} \left( T_{sfc} - \theta_{9,21} \right)$$

$$LHF_{interfacial} = \rho L_e C_{EZ} U_{33} \left( q_{sfc} - q_{9,21} \right)$$

$$\tau = \rho C_{DZ} U_{33}$$

where $\rho$ was air density, $q_{sfc}$ was the surface humidity, $U_{33}$ was the wind speed measured at 33 m, elevation, $\theta_{9,21}$ and $q_{9,21}$ were the potential temperature and specific humidity, each measured both at 9 and 21 m, and $C_{HZ}$, $C_{EZ}$, and $C_{DZ}$ were the surface transfer coefficients for heat, moisture, and momentum (sometimes termed Stanton number, Dalton number, and drag coefficient). The subscript "z" indicates that the value of the transfer coefficients depended on the specific elevations of the $U$, $\theta$, and $q$ measurements.

The transfer coefficients were determined using MOST as follows

$$C_{HZ} = \frac{k^2}{\ln \left( \frac{z_{9,21}}{z_0} \right) - \psi_h \left( \frac{z_{9,21}}{L} \right) \ln \left( \frac{z_{13}}{z_0} \right) - \psi_m \left( \frac{z_{13}}{L} \right)}$$

(19a)
where $k$ is von Kármán’s constant (equal to 0.4), $z_{ot}$, $z_{oq}$, and $z_o$ are the roughness length scales for heat, moisture, and momentum, respectively, $z$ (with subscripts in meters) are the measurement heights. The empirical stability functions $\psi_h$ and $\psi_h$ are functions of $z/L$ where $L$ is the Obukhov length.

$$L = \frac{\bar{\vartheta} u_*^2}{kg \theta_v}.$$  \hfill (20)

$\bar{\vartheta}$ is the mean virtual potential temperature and the scaling parameters $u_*$, $\theta_v$, $q_v$, and $\theta_{v*}$ defined by

$$u_* = \left( \frac{\tau}{\rho} \right)^{1/2}$$ \hfill (21a)

$$\theta_v = \frac{\text{SHF}}{u_* \rho C_p}$$ \hfill (21b)

$$q_v = \frac{\text{LHF}}{u_* \rho C_p}$$ \hfill (21c)

$$\theta_{v*} = \theta_v + \frac{0.61 \bar{\vartheta}}{(1 + 0.61 \bar{q})} q_v$$ \hfill (21d)

where $\bar{\vartheta}$, $\bar{q}$, and $\rho$ are mean values and SHF and LHF include the sea spray component, as described below.

Surface layer conditions during this case were always unstable; the Paulson (1970) $\psi_m$ and $\psi_h$ stability functions (not shown here) were used in Equation 19.

### 2.7.3. Bulk Method Roughness Length Scales

The key to obtaining accurate flux estimates using the bulk method is to correctly specify the values of the roughness lengths, $z_{ot}$, $z_{oq}$, and $z_o$. The effects of the roughness length scales on the surface fluxes are more easily understood and interpreted if they are expressed in terms of the commonly used 10 m neutral transfer coefficients $C_{\text{HN10}}$, $C_{\text{EN10}}$, and $C_{\text{DN10}}$. These are calculated by setting all the $z$ values to 10 m and all the $\psi$ functions to zero in Equation 19.

The following subsection describes the method used to assign the roughness lengths for this case. Although there was a large variability in the value of the momentum roughness length scale $z_o$, across the domain, it will be shown that this had relatively minor impacts on the calculations of enthalpy fluxes (e.g., the values of $C_{\text{HN10}}$ and $C_{\text{DN10}}$) compared to the momentum fluxes ($C_{\text{DN10}}$). This is because heat and water vapor fluxes (unlike momentum) were not directly affected by the pressure transport term (form drag) associated with floe edges, ridges, and other sharp topographic features (e.g., Lu et al., 2011). Despite being relatively unimportant to the enthalpy fluxes, $z_o$ was used in the bulk flux equations; therefore the method used to quantify it will be presented here, with the results being more importantly applicable for the momentum analysis later in this study.

There have been several published studies of how $z_o$ behaves in hurricane-force open ocean conditions. There also have been published quantifications of $z_o$ as a function of sea ice characteristics in lower wind conditions. However, we are unaware of any published in situ measurements of $z_o$ from conditions with near-hurricane force winds, large waves and the presence of sea ice, as was experienced in the TNBP during the PIPERS cruise. Despite this shortcoming in previous research guidance, during PIPERS there was a considerable amount of information available regarding surface conditions (both over the more open ocean areas and the sea-ice covered areas) based on human observations, still and video photography, and satellite imagery, as described in Part 1. These observations were used to match the particular surface conditions observed during this case with the closest analogy to past situations where $z_o$ had been quantified.
Although human and machine imaging observations were crucial for understanding the surface conditions, it was not possible to use them to classify the surface conditions for each of the hundreds of data points used for the analysis of this case. Therefore, an automated method was developed to identify ice type and \( z \) values. As a result of comparing measurements of surface temperature \( T_{\text{sfc}} \) with ice conditions during this case and also during other portions of the PIPERS cruise (sometimes by looking out the window while monitoring the real time \( T_{\text{sfc}} \) readout), the author developed a technique to identify sea ice types by \( T_{\text{sfc}} \) signatures. In particular, and not surprisingly, \( T_{\text{sfc}} \) was correlated with ice thickness, \( d \). We defined an effective ice thickness \( d_{\text{ice}} \) as

\[
d_{\text{ice}} = k_1 \frac{T_{\text{sfc}} - T_{\text{sea}}}{\text{HF}}
\]

where \( k_1 \) was the nominal ice thermal conductivity (set to 2.0 Js\(^{-1}\) m\(^{-1}\) K\(^{-1}\)), \( T_{\text{sea}} \) was the ocean temperature at the bottom of the ice (equal to \(-1.7^\circ\text{C}\)), and HF was the total net surface heat flux

\[
\text{HF} = \text{SHF} + \text{LHF} - \epsilon_{\text{sfc}} \text{LW}_{\text{sfc}} \downarrow + \epsilon_{\text{sfc}} \sigma T_{\text{sfc}}^4
\]

where the last two terms represent the downwelling and upwelling longwave radiation fluxes, both directly measured near the surface for this case. Equation 22 assumes steady-state thermal forcing conditions, that is, a linear temperature profile in the ice. This condition was met well in this case (due to relatively thin ice) and the effective ice thickness \( d_{\text{ice}} \) was close in value to the actual ice thickness \( d \) when the surface was dry and not covered with snow. Surface wetting warmed the surfaces; therefore \( d_{\text{ice}} \) was less than \( d \) in these situations, which generally occurred when the ice was less than 4 cm thick and small pancakes or smaller ice types were present. Snow is an effective thermal insulator, so \( d_{\text{ice}} \) was greater than the actual thickness \( d \) when snow was present, which, for this case, generally occurred on floes that were thicker than 35 cm.

Although there was not perfect agreement between the effective ice thickness \( d_{\text{ice}} \) and the actual thickness \( d \), there was a strong relation between \( d_{\text{ice}} \) and the surface sea ice types. For the bulk method, information on sea ice type is more important than knowing the actual physical ice thickness of the ice. This is because the sea ice type can be more easily related to \( z \) values than ice thickness alone. Also, because \( T_{\text{sfc}} \) was measured directly, the details on ice thickness and conductive heat transport through the ice were not needed to estimate surface fluxes using the bulk method for this case.

The following surface types were observed, as described in Part 1: open ocean, grease ice, shuga, small pancakes, fused pancakes, dragon skin, rough young ice floes, and snow-covered ice floes. Each surface type was associated with a specific effective thickness \( d_{\text{ice}} \) to which values of \( C_{\text{DN10}} \) and other parameters associated with the surface wetness and spray flux calculations (discussed later) were assigned. These represented the “archetypal” values for a particular surface type. For \( d_{\text{ice}} \) values between the archetypal values, \( C_{\text{DN10}} \) and the other parameters were linearly interpolated.

Open ocean areas were significantly rougher than soupy grease ice and lumpy shuga surfaces. However, distinguishing these surface types from each other using the automated method was a challenge because all three surface types had \( T_{\text{sfc}} \) values that were within the measurement error (±1.4°C, Part 1) of the liquid ocean surface temperature, \(-1.7^\circ\text{C}\). This problem was addressed by tuning the archetypal \( d_{\text{ice}} \) values so that the overall surface areas of open ocean, grease and shuga predicted by \( d_{\text{ice}} \), best matched the human observations from the Palmer bridge and other information (such as IR photography), when available. In this way, the accuracy of the larger scale (greater than 1 km) enthalpy and momentum flux calculations, using the automated surface type classification scheme, was optimized, although some of the smaller scale variability of these surface types were unrealistically represented by pseudo-random measurement noise. The next section addresses the roughness assignments (expressed as a neutral 10 m drag coefficient \( C_{\text{DN10}} \)) for the archetypal surface types.

### 2.7.4. Surface Aerodynamic Roughnesses

#### 2.7.4.1. Open Water

This case was characterized by high winds, short fetches and relatively shallow water, all of which would be expected to increase \( C_{\text{DN10}} \) compared to average open ocean conditions. However, several recent studies of TCs and in laboratories indicate that, above \( \sim 25 \) ms\(^{-1}\), the value of \( C_{\text{DN10}} \) levels off and starts to decrease at higher wind speeds (e.g., Donelan, 2018). At some point, the surface becomes “saturated” with respect to
roughness as spume is swept off of wave crests into the troughs and the entire surface is covered with a mass of foam, spray and torn waves. Published estimates for the saturation value of $C_{DN}$ have composite means that are $\sim 2.0 \times 10^{-3}$, which occurs at wind speeds of 30 ms$^{-1}$ (see Figure 2 in Bryant & Akbar, 2016). The open water surface during the PIPERS katabatic winds appeared to be saturated with respect to roughness (see description in Part 1); therefore this value was assigned to the polynya open water areas.

2.7.4.2. Grease and Shuga Ice

Grease ice formed in bands (see Part 1 description) when there was enough frazil (suspended ice crystals) to suppress the capillary and other small surface waves. As more frazil was formed, the surface became a viscous soupy mass of water and ice. In the thickest parts, the ice congealed to form clumps up to 4 cm in diameter, referred to as shuga. By suppressing the small waves, and preventing larger waves from growing, these surface types were aerodynamically smoother than the open ocean areas. Previous measurements of the aerodynamic roughness of grease ice by Guest and Davidson (1991), hereafter GD, resulted in low (aerodynamically smooth) $C_{DN10}$ values of $0.7 \times 10^{-3}$. However, GD was based on measurements taken during lower wind speed conditions than this case. For this high wind case, larger gravity waves were propagating through the grease/shuga bands. In addition to directly creating roughness, the waves also caused some of the shuga clumps to project above the surface, providing some additional roughness. The grease and shuga were in close contact and could not be distinguished using the $T_{sfc}$ measurements (both were near freezing); therefore they were grouped together and assigned an archetypal $C_{DN10}$ value of $1.3 \times 10^{-3}$, which was intermediary between the GD grease ice $C_{DN}$ values and the open ocean value.

2.7.4.3. Small Pancakes

Farther downwind enough ice had congealed to form small pancake ice floes. Similar to grease and shuga, pancake ice suppressed the shorter length ocean waves. However, the floe edges provided additional surface roughness, so archetypal $C_{DN10}$ was increased to the GD value of $1.6 \times 10^{-3}$ for medium pancakes.

2.7.4.4. Fused Pancakes

When the pancakes fused, rafted and became thicker from mechanical processes, the edges often protruded out of the water several centimeters from wave action. Additional roughness elements were created by floe rafting and collection of frozen debris, especially around the edges. Visually the surface appeared rough and the archetypal $C_{DN10}$ was set to $2.1 \times 10^{-3}$, which was 10% greater than the GD value for fused pancakes to account for the more dynamic conditions in this case compared to the GD data set.

2.7.4.5. Dragon Skin and Young Floe Ice

Dragon skin (highly compacted and tilted pancakes) and young ice floes (primarily formed thermodynamically) were quite different in appearance from each other. However, they did have similar mean effective ice thicknesses $d_{ice}$ and mean surface temperatures $T_{sfc}$ and were therefore difficult to distinguish using the automated ice classification method. The dragon skin surface was covered with small (order 20 cm) roughness features caused by the tipped pancake edges. The young floes had relatively flat centers where the ice had formed thermodynamically. However, at the edges, and between flows, there were relatively large form drag features created by the floe freeboard edges and considerable rafted and floating ice debris created by the strong internal ice shear and compression forces. Although the roughness characteristics were different, the overall effect on wind drag was estimated to be similar, and dragon skin and young floe ice were grouped together as one ice type and assigned a value of $2.8 \times 10^{-3}$, which was the GD value for rough young ice.

2.7.4.6. Snow-Covered Young Floes

In the farthest downwind locations of the transect for this case, snow had accumulated to depths of 2–5 cm on top of the young ice floes. The rough young ice surface was smoothed as the snow drifted into the lower areas and filled over some of the roughness elements. For these areas, the archetypal $C_{DN10}$ value was lowered to $2.6 \times 10^{-3}$, which was the GD value for medium-smooth young ice.

Considering all of the ice types, the $C_{DN10}$ values used in this study were similar to the values estimated by Schröder et al. (2003). The exception was grease ice, for which Schröder and coworkers reported a surprisingly high $C_{DN10}$ value of $2.5 \times 10^{-3}$. 
2.7.5. Surface Heat and Moisture Roughness Scale

For this case, $z_{oq}$ and $z_{m0q}$ were determined using the Andreas (1987) model, which employs empirical functions of the roughness Reynolds number, $R_R$, to specify $z_o$ and $z_{m0q}$ where

$$R_R = \frac{z_o u_*}{v}$$

(24)

and $v$ is the kinematic viscosity of the air. Andreas et al. (2005) verified this model for Antarctic conditions. Although the Andreas empirical functions were based on measurements over snow and ice surfaces, the results were not significantly different (compared to other uncertainties) from the often-used functions developed for open ocean conditions by Liu et al. (1979). Therefore, the Andreas (1987) model for $z_o$ and $z_{m0q}$ was applied for all surfaces for this case, including the open ocean. The model was not particularly sensitive to changes in $z_o$ and the predicted values of $C_{IN10}$ and $C_{IN11}$ (they were slightly different from each other) ranged from 1.1 to 1.35 $\times$ 10$^{-3}$, even though the surface roughness values varied considerably over the domain of this case.

2.7.6. Surface Wetness

Surface humidity cannot be directly measured. However, previous research has shown that over snow and ice surfaces in cold seasons the surface specific humidity $q_{sfc}$ is always equal to the ice saturation value $q_{sati}$ (e.g., Andreas et al., 2010). Over open ocean, $q_{sfc}$ is equal to 98% of the fresh water saturation value $q_{satw}$ (e.g., Fairall et al., 2003), and for this case, grease ice, and smaller shuga clumps were assumed to have the same $q_{sfc}$ as the open ocean. As the shuga transitioned to small pancakes, some of the ice was exposed to the air long enough so that the surfaces froze and became dry. The smallest pancakes had partially wet surfaces from splashing, spray, and dunking. However, as these became fused and grew into larger pancakes, these processes were greatly reduced, so that anything thicker than the archetypal fused pancakes (or $d_{ic}$ greater than 0.05 m) had virtually dry surfaces at all times. The surface specific humidity $q_{sfc}$ was determined by

$$q_{sfc} = 0.98 \alpha_{wet} q_{satw} + (1 - \alpha_{wet}) q_{sati}$$

(25)

The wetness factor $\alpha_{wet}$ was the portion of the surface that was wet. For open water and grease ice surfaces, $\alpha_{wet}$ was set to 1; for large fused pancakes and thicker ice types, it was set to 0. For ice types between grease/shuga and small pancakes, $\alpha_{wet}$ ($d_{ic}$ between 0.004 and 0.008 m) was interpolated from 0.0 to 0.5 (in $d_{ic}$ space) and similarly from 0.5 to 1.0 in the small pancake to large pancake $d_{ic}$ range, 0.008–0.05 m.

This completes the description of how the interfacial fluxes were calculated for this case using the bulk method. For surfaces that had no significant amounts of open water, such as the fused pancakes or thicker ice types, this was all that was required to estimate the surface enthalpy fluxes. However, in areas with open ocean exposed, sea spray was generated, and its effects must be accounted for in the bulk method, as described in the next subsection.

2.7.7. Sea Spray

2.7.7.1. Sea Spray Previous Results

Recent studies of TCs show that including the enhancing effects of sea spray on the enthalpy fluxes improves numerical simulations of TC-observed features (Bao et al., 2000, 2011; He et al., 2018; Ma et al., 2017). Numerical process studies (Garg et al., 2018; T. Zhang & Song, 2018) and laboratory studies (Ortiz-Suslow et al., 2016) also suggest that at wind speeds similar to the PIPERS case, the surface SHF is enhanced by sea spray by a factor of two or more. The importance of sea spray is not universally accepted. ABL measurements during the CBLAST project (Drennan et al., 2007), direct numerical simulations of sea spray droplets (Peng & Richter, 2019) and a TC numerical model (Haus et al., 2010) downplay the importance of sea spray during high wind situations. With that caveat, this study will assume that sea spray was important for the PIPERS case and will check that assumption by comparing the different flux method results.

2.7.7.2. Sea Spray Model—Baseline Case

Because the sea spray flux of enthalpy has different physics from the interfacial flux, MOST surface layer theory does not apply (Andreas, 1992). To account for the effects of sea spray during this case, we used the Andreas et al. (2015), hereafter “AMV” for the three authors, ocean surface flux model that includes
parameterizations for sea spray microphysics, and is applicable to low (but nonfreezing) temperatures. The AMV model was applied for only one baseline case using the average values of the state variables wind speed $U_{33}$, potential temperature $\theta_v$, surface temperature $T_{sfc}$, and humidity $q_v$ when the Palmer was in areas of open water and $x$ less than 24 km. The state variables did not change greatly across that area, justifying using the result of just one representative set for the baseline sea spray estimates. We define baseline sea spray enhancement factors $\alpha_{\text{sens}}$ and $\alpha_{\text{lat}}$ as

$$\alpha_{\text{sens}} = \frac{\text{SHF}_{\text{sprayA}}}{\text{SHF}_{\text{interfacialA}}} = 1.10,$$  \hspace{1cm} (26a)

$$\alpha_{\text{lat}} = \frac{\text{LHF}_{\text{sprayA}}}{\text{LHF}_{\text{interfacialA}}} = 0.25$$  \hspace{1cm} (26b)

where $\text{SHF}_{\text{sprayA}}$, $\text{SHF}_{\text{interfacialA}}$, $\text{LHF}_{\text{sprayA}}$, and $\text{LHF}_{\text{interfacialA}}$ are the spray and interfacial components of the SHF and LHF as predicted by the baseline AMV case. The sensible heat was more greatly enhanced because of the droplet surface microscale, sensible heat transfer is more efficient than evaporation (e.g., Andreas, 1998). Note that we used the AMV model only to estimate the ratio of the interfacial versus sea spray components for this one baseline case, not to quantify the actual total fluxes for each location in the PIPERS case.

### 2.7.7.3. Sea Ice Effects on Sea Spray Production

Sea ice damps breaking waves and is a physical obstruction between the ocean surface and the atmosphere. Lacking a published physical model of these effects on spray enthalpy fluxes, we estimate them based on the visual observations described above and in Supporting Information S1. We specify a spray production ratio $\alpha_{\text{prod}}$, which is multiplied by the AMV enhancement factors $\alpha_{\text{sens}}$ and $\alpha_{\text{lat}}$ to estimate the relative sea spray contributions to the enthalpy fluxes. In the open ocean areas, the presence of subsurface frazil and nearby grease/shuga bands reduced the sea spray production by an estimated 20% or $\alpha_{\text{prod}}$ was set to 0.8. Within the bands, the grease ice and shuga had a clearly observable damping effect on the ocean surface, particularly capillary waves, but did not totally prevent wave breaking or spume generation; $\alpha_{\text{prod}}$ was set to 0.4. Pancakes further damped the waves and also prevented direct exposure to the liquid surface. But even in these areas, some sea spray was generated between the pancakes on top of waves and when the pancakes tumbled down steep or breaking wave faces; for pancake areas, $\alpha_{\text{prod}}$ was set to 0.2. Once the pancakes began to fuse and became rafted, the sea spray was virtually eliminated and assumed to contribute nothing to the surface fluxes. To summarize using the $d_{\text{ice}}$ equivalents to the surface types

$$\alpha_{\text{prod}} = \begin{cases} 0.8 & d_{\text{ice}} \leq 0.0035 \text{ m} \\ 0.8 - 0.4 & 0.0035 \text{ m} < d_{\text{ice}} \leq 0.005 \text{ m} \\ 0.4 - 0.2 & 0.005 \text{ m} < d_{\text{ice}} \leq 0.008 \text{ m} \\ 0.2 - 0.0 & 0.008 \text{ m} < d_{\text{ice}} \leq 0.05 \text{ m} \\ 0 & 0.05 \text{ m} < d_{\text{ice}} \end{cases}$$  \hspace{1cm} (27)

where the dashes indicate linear interpolations.

### 2.7.7.4. Freezing Droplet Effects on SHF

The AMV model does not include the effects of freezing and frozen spray on the particle microphysics. Frozen spray of the amount $q_{\text{freeze}}$ converted the enthalpy associated with the freezing $L_f q_{\text{freeze}}$ to sensible heat $C_w \Delta T$ in the atmosphere, where $L_f$ is the specific latent heat of fusion and $\Delta T$ is the change in air temperature due to the freezing. We define the frozen sea spray ratio $\beta$, as

$$\beta = \frac{q_{\text{freeze}}}{q_{\text{spray}}}$$  \hspace{1cm} (28)

where $q_{\text{spray}}$ is the total mass of the sea spray, including the frozen part. The sensible heat transfer from a sea spray droplet to the atmosphere in the AMV model; $H_{\text{dsA}}$, comes from the change in the particle temperature $\Delta T_w$

$$H_{\text{dsA}} = C_w \Delta T_w$$  \hspace{1cm} (29)

where $C_w$ is the heat capacity of seawater and $\Delta T_w$ is the change in droplet temperature. If fraction $\beta$ of that droplet water freezes, an additional $\beta L_f$ of sensible heat is added to the atmosphere for a total spray sensible heat transfer, $H_{\text{ds}}$.  

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\[ H_{\text{sh}} = C_w \Delta T_w + \beta L_f \]  

(30)

The freezing particles stay warmer and therefore transfer more sensible heat to the atmosphere compared to the AMV model prediction. The potential enhancement of the total spray SHF \( \text{spray} \), including potential droplet freezing, compared to the AMV spray prediction is expressed by the ratio \( \alpha_{\text{Sfreeze}} \)

\[ \alpha_{\text{Sfreeze}} = \frac{H_{\text{sh}}}{H_{\text{shA}}} . \]  

(31)

If \( T_w \) decreases to the air temperature \( T_{\text{air}} \) in the AMV model and this case then

\[ \Delta T_w = T_{\text{sea}} - T_{\text{air}} . \]  

(32)

Therefore,

\[ \alpha_{\text{Sfreeze}} = 1 + \beta \left( \frac{L_f}{C_w} \left( T_{\text{sea}} - T_{\text{air}} \right) \right) \]  

(33)

The freezing of droplets is a powerful effect, increasing the sea spray SHF \( \text{spray} \) by a factor of five for the portion that freezes and cools to air temperature. If the droplets freeze but do not cool to the air temperature, then the potential enhancement compared to AMV would be even greater than predicted by Equation 33.

### 2.7.7.5. Freezing Droplet Effects on LHF

When sea spray droplets freeze, the spray LHF from the ocean \( LHF_{\text{spray}} \) is reduced because the ice surface vapor pressures are lower than the liquid droplet surface vapor pressures. Freezing will occur preferentially on the smaller droplets and outsides of larger droplets, and therefore the attenuation effect of frozen spray on the \( LHF_{\text{spray}} \) will be greater than the mass ratio \( \beta \). This moisture blocking effect was parameterized by defining a “surface area attenuation factor” \( \gamma \), so that the decrease in \( LHF_{\text{spray}} \) is \( \gamma \beta \) where \( \gamma \) is a number greater than one and less than \( \beta^{-1} \) that represents the greater impact of freezing on surface area ratios versus mass ratios. The \( \gamma \) factor also includes time scale effects (droplet surface areas freeze faster than droplet mass) and the small moisture exchange from ice to vapor after the particles have frozen and before they fall out. The attenuation effect of freezing on \( LHF_{\text{spray}} \) is represented by

\[ \alpha_{\text{Lfreeze}} = \left( 1 - \gamma \beta \right). \]  

(34)

The effect represented by \( \alpha_{\text{Lfreeze}} \) is not as potentially powerful as \( \alpha_{\text{Sfreeze}} \) because at most (if all the spray instantly froze) \( \alpha_{\text{Lfreeze}} \) would be near zero, which represents a factor of one reduction in \( LHF_{\text{spray}} \) which is significantly less than factor of five increase in \( \text{SHF}_{\text{spray}} \) which would occur under the same circumstances.

### 2.7.7.6. Quantifying the Overall Effects of Sea Spray

During this case, it was observed that \( \sim 5\% \) of the sea spray that hit the bridge windows was already frozen before impact (Part 1), providing a ballpark estimate for \( \beta \) of 0.05. The determination of \( \gamma \) for this case was speculative; it was set to a nominal value of 2.0, with the understanding that \( \gamma \) represents a large amount of unknown physics and could be in error by a large factor. Setting the change in droplet temperature (Equation 32) to the approximate air/sea temperature difference value of 21°C results in values for \( \alpha_{\text{Sfreeze}} \) and \( \alpha_{\text{Lfreeze}} \) of 1.2 and 0.9, respectively. The large uncertainties in the values of \( \beta \) and \( \gamma \) will be considered when assessing the total accuracy of the flux estimates for this case.

The total sea spray production, sea ice attenuation, and droplet freezing effects of sea spray on the surface enthalpy fluxes, as represented by \( \alpha_{\text{sensA}}, \alpha_{\text{latA}}, \alpha_{\text{prod}}, \alpha_{\text{Sfreeze}}, \) and \( \alpha_{\text{Lfreeze}} \), were assumed to be independent of each other. The overall interfacial sensible and latent heat ratios, \( \alpha_{\text{sens}} \) and \( \alpha_{\text{latent}} \), could then be specified

\[ \alpha_{\text{sens}} = \alpha_{\text{sensA}} \alpha_{\text{Sfreeze}} \alpha_{\text{prod}} = (1.1)(1.2) \alpha_{\text{prod}} = 1.32 \alpha_{\text{prod}} \]  

(35a)

\[ \alpha_{\text{latent}} = \alpha_{\text{latA}} \alpha_{\text{Lfreeze}} \alpha_{\text{prod}} = (0.25)(0.9) \alpha_{\text{prod}} = 0.225 \alpha_{\text{prod}} \]  

(35b)

where \( \alpha_{\text{prod}} \) was a function of \( d_{\text{m}} \), representing surface type (Equation 27). The estimated sea spray components \( \text{SHF}_{\text{spray}} \) and \( \text{LHF}_{\text{spray}} \) were determined by
The total surface enthalpy fluxes (SHF and LHF) using the bulk method were

\[
\text{SHF}_{\text{spray}} = \alpha_{\text{sens}} \text{SHF}_{\text{interfacial}} \tag{36a}
\]

\[
\text{LHF}_{\text{spray}} = \alpha_{\text{latent}} \text{LHF}_{\text{interfacial}} \tag{36b}
\]

The effect of \(H_{\text{fspray}}\) was included in the bulk SHF calculation and therefore did not need to be included as a separate heat flux term for the ocean.

When snow or frozen sea spray fell back to the surface it either melted or, more likely in this case, contributed directly to sea ice formation. With regard to frozen sea spray, the additional ocean enthalpy loss when the spray returns to the surface \(H_{\text{fspray}}\) was

\[
H_{\text{fspray}} = (\alpha_{\text{freeze}} - 1) \text{SHF}_{\text{interfacial}} \tag{38}
\]

It was shown earlier how the heat gained by the atmosphere due to snow growth \(H_p\) could be estimated based on the bulk flux estimates (Equation 8). However, because this warming occurred mostly in the upper ABL, it was not included in the bulk SHF calculation. When the snow falls to the surface, the ocean loses heat (or directly gains sea ice) from falling snow \(H_{\text{snow}}\) determined by

\[
H_{\text{snow}} = P_t \frac{L_s}{L_s} = H_p \frac{L_s}{L_s} \tag{39}
\]

The snow enthalpy \(H_p\) was defined in terms of sublimation, which is equal to the fusion plus evaporation enthalpies. The evaporation part of \(H_p\) was accounted for in the bulk LHF flux calculation. However, the fusion component was not; therefore the total heat loss from the ocean should also include \(H_{\text{snow}}\). Two forms of surface LHF were classified in this study: LHF and LHF\(_A\), each related to the evaporation and sublimation enthalpies respectively. To calculate ocean heat loss from latent heat, either LHF plus \(H_{\text{snow}}\) or just LHF\(_A\) could have been used for this case, because all extra moisture from the surface went into making snow, which, when it returned to the surface, made up the difference between LHF and LHF\(_A\).

### 2.7.7.8. Bulk Method Summary and Numerical Procedure

The key parameters used in the bulk surface enthalpy and momentum flux calculations for this case are listed in Table 1. The effective ice thickness \(d_{\text{ice}}\) was the independent (determining) variable for all the other (nondimensional) parameters, which were linearly interpolated between the archetype values shown in Table 1.
Because the roughness lengths $z_x$, $z_r$, $z_q$, the stability functions $\psi_{\text{fr}}$, $\psi_{\text{fl}}$, and all the $d_{\text{ref}}$-dependent parameters listed in Table 1 were functions of the fluxes SHF, LHF, and $\tau$, the solutions were intrinsic, and were obtained after assigning initial guesses by iteration of Equations 18–25, 27, and 35–37, with numerical convergence occurring within five cycles for all data points. This completes the description of the bulk method that was used to estimate the surface fluxes of sensible heat SHF, latent heat LHF and momentum $\tau$.

2.8. Integral Method

2.8.1. Rawinsonde Profiles

The integral method (sometimes referred to as the integral-profile method) estimates the sources and sinks of a particular quantity by measuring the change in the value of that quantity following the air parcels within an atmospheric layer. The simplified ABL enthalpy budget (Equation 5) formed the basis of the integral method for this case. The components of the ABL enthalpy $H_{\text{ABL}}$ were separated into the sensible $C_p \theta_{\text{ABL}}$ and latent $q_{\text{ABL}}$ components and after assuming steady-state conditions (Equation 1), the budget equations can be expressed in terms of finite differences between the rawinsonde locations.

$$F_{\text{sens}} = \rho_{\text{ABL}} C_p z_{i12} U_{\text{ABL}} \left( \frac{\theta_{\text{ABL}2} - \theta_{\text{ABL}1}}{x_2 - x_1} \right)$$

$$F_{\text{latent}} = \rho_{\text{ABL}} L_e z_{i12} U_{\text{ABL}} \left( \frac{q_{\text{ABL}2} - q_{\text{ABL}1}}{x_2 - x_1} \right)$$

which show that the sink/source terms for sensible and latent heat, $F_{\text{sens}}$ and $F_{\text{latent}}$, were determined from the horizontal gradients of $\theta_{\text{ABL}}$ and $q_{\text{ABL}}$ as measured by successive rawinsonde profiles. As elsewhere in this study, the “ABL” subscript refers to vertical averages within the atmospheric boundary layer, “1” and “2” refer to the upwind and downwind values and $z_{i12}$ is the $z$ average from locations $x_1$ and $x_2$.

The surface SHF and LHF were then derived as follows:

$$\text{SHF} = F_{\text{sens}} - \text{LW}_{\text{up}} \psi_{\text{ABL}} + H_p - E_{\rho}$$

$$\text{LHF} = F_{\text{latent}} - H_p + E_q$$

where the radiation divergence $\text{LW}_{\text{up}}$ and snow enthalpy $H_p$ were estimated from the ship measurements and bulk model as described earlier. The entrainment terms $E_{\rho}$ and $E_q$ were set to zero for the integral flux estimates. However, entrainment was considered in the discussion of the results and therefore the terms were left in Equation 40. The radiation $\text{LW}_{\text{up}}$ and snow $H_p$ terms were averaged in the Lagrangian framework by using the following technique. First, each distance interval between rawinsonde profiles was divided into 20 subintervals. The estimates of $\text{LW}_{\text{up}}$ and $H_p$ from every 1-min-average data point were then assigned to the appropriate subinterval (depending on the ship’s $x$ location) and averaged over each subinterval, to produce spatial $x$ means instead of ship-time-over means. A final step was to weight (multiply) each of the subinterval spatial means by $1/U_{x_{\text{up}}}$ so that the averaging dimension was air-parcel-time-over for each $x$ subinterval. The same procedure was applied to the bulk surface fluxes estimates and other ABL enthalpy terms; therefore the resulting means of the bulk method estimates for each of rawinsonde profiles intervals were relevant in the same Lagrangian reference frame that was used for the integral method.

2.8.2. Manuela AWS

The permanent Manuela AWS located at 74 m elevation, 1 km from the ice shelf edge (the position indicated by “P0” for this study) and almost directly upwind of the location the most upwind rawinsonde profile location P1 (see Figures 1 and 2 in Part 1), provided another opportunity to estimate SHF and LHF using the integral method across the most open upwind region of the TNBP where the greatest surface fluxes and ABL modifications occurred (Lazzara et al., 2012). Although upper-air profiles were not available at the Manuela location, it was assumed that $\theta$ and $q$ were well mixed from the measurement level up to a $z_i$ level of 150 m, a typical ABL depth value for coastal Antarctic locations in high winds during cold seasons (Vignon et al., 2019). The satellite images indicated that the surface conditions upwind of $x$ equals 17 (open water and grease/shuga bands) were similar to the 17–24 km $x$ region and therefore the same mean surface
parameters were used to estimate the surface fluxes and radiation and snow effects between Manuela, at the ice shelf edge, and the most upwind ship location at P1.

2.8.3. Ship Measurements

The integral method assumes that the $\theta_{\text{ABL}}$, $q_{\text{ABL}}$, and $U_{\text{ABL}}$ values represent mean conditions at the $x$ locations (as well as vertical ABL means). However, as shown in Part 1, because the rawinsonde measurements were virtually instantaneous, they were subject to sampling errors due to variations from turbulence and midscale features such as ABL rolls. It was also demonstrated that the Palmer mast measurements of $\theta_{21}$, $q_{21}$, and $U_{21}$ were good proxies for the average ABL conditions. By averaging the 1-sec-sampled $\theta_{21}$, $q_{21}$, and $U_{21}$ data for periods of several minutes, the sampling errors associated with natural turbulent variability, ABL circulations and/or waves, were minimized. This method will be termed the “surface integral” method to indicate it was based on surface layer measurements (except for $z_s$). Although the SHF and LHF could theoretically be estimated for any $x$ interval using the surface integral method, it was most useful to apply it to the same locations as the P1–P4 rawinsonde measurements because (a) these values could be directly compared with the rawinsonde integral estimates, (b) the intervals between rawinsonde locations were large enough to detect the signal of the average changes from the turbulent and midscale noise, and (c) the ship was relatively stationary during the rawinsonde launch times, which provided an opportunity to sample conditions at one $x$ location for 10–20 min. The same procedure was performed using the 9 m level Palmer boom $\theta_s$ and $q_s$ values; the results were almost identical and will not be shown. To distinguish the different surface flux estimates, the average enthalpy flux values for each interval will be referred to as SHF$_b$, SHF$_f$, LHF$_b$, LHF$_f$, and LHF$_s$ where the subscripts “b,” “f,” and “s” denote the “bulk,” “integral,” and “surface integral” estimation methods, respectively.

2.9. Enthalpy Flux Error Analysis

The accuracy of the surface flux estimates was estimated by considering all potential sources of error, including sensor accuracy (e.g., icing, aeration effects, lab accuracy), sampling errors (e.g., natural turbulent variability, representativeness of observed surface conditions), ship contamination effects (e.g., wind blockage, dynamic pressure fluctuations, heat sources), parameter specifications (e.g., transfer coefficients, $\alpha$ values), and theory validity (one-dimensional assumption, spray effects, no entrainment). The accuracies of the flux calculation input values were estimated based on the author’s experience, published accuracy claims, laboratory calibrations, sensor comparisons, and other methods, as described and quantified in Part 1 and Supporting Information S1. Uncertainties were expressed either as a percentage of the quantity value (relative) or a specific value range (absolute) or both, depending on the nature of the uncertainty. Every input variable used for the three flux method estimates was assigned an uncertainty value based on the relative plus absolute error estimates. The flux calculations were then redone separately for each input variable using the values representing the endpoints of the confidence interval for that input variable to determine the total effect on the flux uncertainty. Generally, the error sources were considered to be independent and therefore the total combined uncertainty estimates were root-mean-squares of each of the error sources; some bias or systematic errors were combined linearly. The uncertainty values (both relative and absolute) used in this study represent a plus or minus uncertainty; therefore the entire (80%) confidence interval range of the flux estimates were twice the values presented. The next few paragraphs provide a summary of the error analysis results.

The largest uncertainty for the bulk method was the sea spray effect quantification, while for the integral methods, it was the one-dimensional assumption. Sampling errors contributed lesser, but sometimes significant, relative uncertainty in lower flux regions. Other sources of uncertainty mentioned in the previous paragraph were not significant, defined as contributing less than 5% uncertainty. Estimating the error of the sea spray contribution to the bulk fluxes was challenging due to uncertainties regarding the application of the AMV model to this case where (a) sea spray generation was suppressed due to sea ice formation, (b) the SHF was enhanced (LHF suppressed) due to the freezing of sea spray droplets in the atmosphere, and (c) some researchers have questioned the validity of AMV, and whether sea spray actually enhances TC fluxes (and, by extension, coastal polynyas). Regarding (a), the values of the sea suppression factor $\alpha_{\text{prod}}$ (Table 1) were based on ship bridge observations and once pancakes began to fill in the surface, it was apparent that sea spray production had become negligible and setting $\alpha_{\text{prod}}$ to zero was accurate. Less certain were the sea ice suppression effects in the open water regions; the value of 0.8 in Table 1 could possibly have been anywhere in the 0.5 to 1 range.
1.0 range, representing a SHF (LHF) a uncertainty of 15% (2.5%) where sea spray was most intense. Regarding (b) (freezing droplet effects) if, in reality, freezing droplets had no effect on the sea spray fluxes, then the SHF \(_b\) (LHF \(_b\)) spray components would have been over (under) estimated by 20% (10%), based on the applied \(\alpha_{\text{Sfreeze}}\) and \(\alpha_{\text{Sfreeze}}\) values (Table 1). On the other hand, the upside “risk” of incorrectly specifying the droplet freezing parameters \(\beta\) and \(\gamma\) was much greater; for example if 25% of the droplet mass froze, instead of the estimated 5%, this would double the sea spray effect or increase SHF \(_b\) by 1,000 Wm\(^{-2}\) in the open areas. Regarding (c) (AMV validity) for this study, the assumption was made that the sea spray did increase the surface fluxes and AMV was a valid model for the sea spray effects (as adjusted for the polar situation) and no specific uncertainty was assigned to the AMV validity. The overall enthalpy flux (interfacial and spray) estimates were affected by sampling errors in the \(T_{sfc}\) measurement (\(\pm 1.4^\circ\text{C}\)), which translates into SHF \(_b\) (LHF \(_b\)) errors of 70 Wm\(^{-2}\) (15 Wm\(^{-2}\)), which were important (in the relative sense) only in the regions downwind of \(x\) equals 50 km, where the overall magnitudes of the surface fluxes (and therefore associated percentage errors) were much smaller than over the open ocean regions. Considering all these potential uncertainties, the estimated error for SHF \(_b\) was 20%, plus 100 Wm\(^{-2}\) and for LHF \(_b\) was 25% plus 20 Wm\(^{-2}\). These error estimates themselves have some uncertainty due to the possibility the AMV model was not valid in this situation. For this reason the confidence interval of the above assigned uncertainties was defined as 80% rather than the usual 95% standard.

The integral method (whether based entirely on rawinsondes or the surface proxy method) uses a one-dimensional approximation as described earlier. In reality, (a) there was some temporal variability in atmospheric and surface conditions, (b) the ship did not exactly follow the downwind trajectory of the ABL air parcels and (c) \(z_i\) may not have been the ABL (mixed layer) depth, and (d) there may have been some entrainment of heat or moisture from above the ABL or from the sides of the jet region. The totals of all these factors were estimated to contribute 25% error to the estimated SHF \(_i\), SHF \(_s\), LHF \(_i\), and LHF \(_s\) values. However, similar to the sea spray effect, there was considerable uncertainty in this error estimate and the resulting range in possible values represents only an 80% confidence interval. The natural atmospheric variability in temperature and humidity due to turbulent and midscale fluctuations created errors in SHF \(_i\) and LHF \(_i\) of \(~\sim 250\) and 50 Wm\(^{-2}\) at all \(x\) locations. The surface integral estimates were much less susceptible to these types of errors, due to averaging. However, sampling errors in \(z_i\) and biases associated with certain locations (e.g., relative to an atmospheric roll) created estimated uncertainties of 35 and 7 Wm\(^{-2}\) at all locations for SHF \(_i\) and LHF \(_i\). These sampling errors were insignificant compared to the other errors in the high flux upwind regions. However, in the most downwind interval, the small change in the mean \(\bar{\theta}_{\text{ABL}}\) and \(\bar{q}_{\text{ABL}}\) values (signal) compared to the natural variability (noise) could have caused errors large enough to change the sign of the (small) heat flux estimates. Errors due to incorrectly specifying the snowfall rate were not significant to SHF \(_{i,s}\) or to the total enthalpy fluxes. However, they were significant for the moisture budget and could affect the LHF \(_{i,s}\) by as much as 40%. Overall the estimated uncertainty ranges (80% confidence interval), for all the integral estimates was 25% of the total value plus 250, 40, 50, and 8 Wm\(^{-2}\) for SHF \(_i\), SHF \(_s\), LHF \(_i\), and LHF \(_s\), respectively.

### 2.10. Thermodynamic Theory Summary

This section showed how the enthalpy (sensible and latent heat) budgets of the ABL and surface fluxes were estimated from the PIPERS data set for this case. Because the surface layer air was saturated with respect to ice, the rate of ice crystal growth and fallout, termed “snow,” was estimated based on the bulk surface flux enthalpy estimates. A one-layer model for longwave radiation was used to estimate ABL radiation effects. Three methods for estimating the surface enthalpy fluxes were used: (a) a bulk method based on ship measurements which included the effects of sea spray, (b) an integral method based on rawinsonde measurements, and (c) an integral method based on ship (surface layer) measurements. The effects of snow and radiation divergence within the ABL were included in the integral methods. The bulk and (rawinsonde) integral methods were almost totally independent from each other, having been based on different measurements and totally different concepts. The only connection was that the snow enthalpy effect used in the integral method was based on the bulk method. The surface integral method uses some of the same measurements as the other two methods and uses the same concept, and snow and radiation corrections, as the rawinsonde integral; therefore it was not independent from the other methods. However, the surface integral method was potentially the most accurate, making use of the \(z_i\) information from the rawinsonde data while allowing time averaging of the state variables to reduce sampling errors.
3. Thermodynamic Results

3.1. Introduction

This section presents the results of the thermodynamic analyses for this case. The three different methods for estimating SHF and LHF (including the sea spray components SHF_{spray} and LHF_{spray}) will be compared. The effects of snow enthalpy $H_s$, net surface radiation $LW_{sfc}↑$ and ABL radiation divergence, $LW_{ABL}↑$, on the surface fluxes and ABL sensible heat budget as a function of $x$ will be shown.

3.1.1. Surface SHF

The three SHF estimates SHF$_b$, SHF$_i$, and SHF$_s$ as a function of downwind parcel trajectory distance $x$ (Figure 1) shows the strong influence of surface temperature $T_{sfc}$ on the SHF values, with very similar variation patterns (compare with Part 1, Figure 3a). The bulk method values SHF$_b$ (black line) were available for all data points, while the (rawinsonde) integral and surface integral method values, SHF$_i$ and SHF$_s$, were based on the intervals P0–P1, P1–P2 and so on, and are plotted as blue circles and green triangles at the interval midpoints. The parcel-time-over-mean SHF$_i$ values (red squares), are also shown at the midpoints. There were no ship or rawinsonde measurements in the region $x$ less than 17 km. However, as described earlier, using the Manuela data and assuming a similar surface as existed at P1, a single bulk estimate for the P0–P1 interval, along with the integral estimates, using Manuela as the upwind point, are shown on the left (upwind) side of Figure 1. In the region $x$ between 27 and 28 km, the Palmer reversed direction and headed upwind and then back downwind three times while several rawinsonde launches were (unsuccessfully) attempted. To avoid plot clutter in this data overlap region, the bulk flux values are plotted as individual points, instead of connected lines, for some of the points in this region in Figure 1 and the following figures.

All three methods compared well (within 100 Wm$^{-2}$ for all intervals), except P2–P3, which will be discussed later. The author made no attempt to tune any of the parameters in the three flux estimation methods toward each other; they were independently formulated. Because $\theta_{sfc}$ was a good proxy for $\theta_{ABL}$, it was not surprising that the two integral methods SHF$_i$ and SHF$_s$, match well for all the intervals. However, such a close match between SHF$_b$ with SHF$_i$ and/or SHF$_s$, was apparently fortuitous, given the estimated errors for each method.

Extremely high SHF values, $\sim$2,500 Wm$^{-2}$ for all three methods, occurred in regions of $x$ less than 30 km where the surfaces were open ocean, grease ice or shuga ice and $T_{sfc}$ was close to the freezing temperature of the sea water, $-1.7^\circ$C. In the region between $x$ equals 24 and 60 km, the ice thickened from thermodynamic growth (freezing), rafting, and tilting under compression, which insulated the ocean and resulted in a relatively steady decrease in SHF in the downwind ($\text{increasing } x$) direction as the ice became thicker. At $x$ greater than 60 km, large young ice floes with thicknesses of 30 cm and sometimes covered with snow were present, which kept SHF below 100 Wm$^{-2}$ except in a few locations where leads had opened in the young pack ice, creating upward spikes in SHF$_b$ (Figure 1). Most of these leads were traversed by the Palmer in less than the 1-min averaging period, which smoothed down the maximum fluxes. A notably large open area was crossed from $x$ equals 67–71 km; here SHF$_b$ increased to 1,400 Wm$^{-2}$ compared to 100 Wm$^{-2}$ in the adjacent young ice floe regions.

The P2–P3 interval was different from the others in that the three SHF estimation methods deviated from each other significantly. The two integral methods estimated values, SHF$_i$ and SHF$_s$ (764 and 640 Wm$^{-2}$), were further apart in value than the other locations, however, within the uncertainties. In contrast, the bulk method estimate SHF$_b$ was considerably lower, 292 Wm$^{-2}$, for reasons explained here. The region between $x$ equals 41 and 60 km was when the ship was in the accumulation zone, a transition between the more open fluid zone ice surface (frazil, grease, shuga, pancakes) and the floe zone of compacted rough young
ice floes, as discussed in Part 1 and Supporting Information S1. This transition region was at an oblique angle (~20°) to the ship track. Although the ship’s crew had been instructed to steam directly downwind, later analysis revealed that the track was, on average, 4.5° backed (to the left) compared to the downwind direction for the P2–P3 interval (the other intervals were considerably closer). This meant that the air parcels reaching the ship at P3 had been exposed to warmer surfaces in the fluid zone for a longer fetch than had been experienced along the ship cruise track, resulting in greater SHF\(_s\) and SHF\(_b\) values compared to the average SHF\(_b\) estimate. The large SHF\(_s\) and SHF\(_b\) estimates relative to SHF\(_i\) could have been realistic for the warmer upwind surface conditions experienced by the air parcels. However, because the air parcels reaching P3 may have already been warmer at the upwind point of the interval (near P2) the integral SHF estimates for P2–P3 were treated with suspicion and SHF\(_b\) will be the more trusted flux value for this interval.

### 3.1.2. Sea Spray Component (SHF)

Despite the uncertainties associated with each SHF estimation method, the bulk and integral methods were closely matched, or in the P2–P3 interval case, the integral estimates were higher than the bulk method. This was strong evidence that sea spray played a major role in enhancing the SHF. Had the sea spray effect SHF\(_{spray}\) been ignored, and SHF\(_b\) based entirely on conventional interfacial surface layer theory, the resulting SHF\(_b\) values in the open areas would have been only approximately one-half of the SHF\(_s\) and SHF\(_b\) estimates, which was outside of their estimated uncertainty ranges.

A comparison of the interfacial and sea spray components of SHF (Figure 2) shows that in the regions where sea water was most exposed, at \(x\) less than 30 km and in the lead centered at \(x\) equals 69 km, the sea spray component SHF\(_{spray}\) was slightly larger than the interfacial component SHF\(_{interfacial}\). By increasing the SHF by ~1,000 Wm\(^{-2}\) in those locations, sea spray had major impacts on ABL modifications, sea ice formation, deep water formation and all the other effects associated with the extreme loss of heat from the TNBP. At \(x\) equals 28 pancake ice started to form, and as the pancake areal coverage increased, the sea spray fluxes decreased dramatically, becoming negligible downwind of \(x\) equals 45 except within the occasional lead openings.

For the SHF\(_b\) calculations, it was assumed that the freezing of the sea spray droplets enhanced SHF\(_{spray}\) by 20%, as represented by the \(\alpha_{freeze}\) equal to 1.2 (Table 1). This assumption increased the SHF\(_{spray}\) by a maximum value of 215 Wm\(^{-2}\) in the most open regions and the freezing enhancement produced the same variation pattern as the SHF\(_b\) sea spray component shown in Figure 2. The \(\alpha_{freeze}\) factor was a function of poorly understood microphysical processes represented by the spray ice mass fraction \(\beta\). If all the sea spray had frozen before falling back into the ocean, the SHF\(_b\) would have been five times greater or more than what was used for this case, which would have been unrealistic; therefore large values of \(\alpha_{freeze}\) can be ruled out. The closeness of the final estimates of SHF\(_{spray}\) to SHF\(_b\) (and SHF\(_s\)), gave confidence that the uncertainty of \(\beta\) did not grossly affect the accuracy of the SHF\(_b\) estimates.

### 3.1.3. Surface LHF

Figure 3 shows the three LHF estimates LHF\(_b\), LHF\(_i\), and LHF\(_s\) as a function of \(x\), using the same template as Figure 1. The bulk estimates LHF\(_b\) will be addressed first. The variation pattern of LHF\(_b\) closely resembled the SHF\(_b\) estimates (Figure 1). This close relationship was because (a) the surface humidity \(q_{sfc}\) was a function of \(T_{sfc}\) (Equation 25), (b) the surface layer humidity \(q_{sfc}\) was essentially a function of \(\varepsilon_s\) (Equation 10), and (c) both flux estimates use \(U_{33}\) (Equation 18). Although the patterns of SHF\(_b\)

![Graph showing bulk, interfacial, and total sensible heat fluxes as a function of distance from the ice shelf.](image)

**Figure 2.** Bulk estimates of the interfacial component of sensible heat flux (SHF\(_{interfacial}\)) (red line), the sea spray component SHF\(_{spray}\) (blue line), and the total SHF\(_b\) (black line) as a function of \(x\). In some locations, SHF\(_b\) and SHF\(_{interfacial}\) are equal and the SHF\(_i\) plotted line covers the SHF\(_{interfacial}\) line.

**Figure 3.** Same as Figure 1 for the three latent heat flux (LHF) estimates: LHF\(_b\) (black line and red squares), LHF\(_i\) (blue circles), and LHF\(_s\) (green triangles).
(Figure 1) and LHF\textsubscript{i} (Figure 3) were similar, the correlation was not exact. The Bowen ratio \( B \) (SHF divided by LHF) varies from 3.9 to 7.0 (not shown). The \( B \) variation was due to (a) the nonlinearity of \( \frac{c_{q_{sat}}}{\partial T} \) (for ice or water surfaces) which causes relative enhancement of LHF at higher surface temperatures, (b) the enhancement of LHF over wet surfaces (\( q_{sat} \) is greater than \( q_{sat} \) for the same temperature), and (c) the greater relative enhancement of SHF due to sea spray (discussed below). The net result of these effects was to have a relatively low \( B \) (enhanced LHF) over the warmer surfaces, except over the most open areas (mostly \( x \) less than 27 km) where sea spray was significant and SHF received a bigger resulting boost. Pancake floes had relatively moist, warm surfaces. However the pancakes suppressed the sea spray and therefore the enthalpy fluxes over these surfaces had the lowest \( B \) values.

For the P0–P1 interval, the three estimated values of 356, 394, and 479 Wm\(^{-2}\) for LHF\textsubscript{sfc}, LHF\textsubscript{b}, and LHF\textsubscript{i}, respectively, would, in most situations, be considered enormous values. However, these LHF magnitudes were dwarfed by the SHF in this case, as indicated by the high \( B \) values. Similar to SHF, the value of LHF from all estimates was greatly decreased as the sea ice formed and thickened, becoming only \( \sim 20 \) Wm\(^{-2}\) over the young ice floes at \( x \) greater than 55 km. In this region, the primary moisture sources were the open leads, which caused the spikes of up to 300 Wm\(^{-2}\) in LHF.

The largest estimated sea spray enhancement of latent heat LHF\textsubscript{spray} was 50–60 Wm\(^{-2}\) occurring over the open areas and virtually zero for \( x \) greater than 45 km (figure not shown). The LHF\textsubscript{spray} proportion of total LHF was much less than for SHF (Equation 26) because the AMV model predicts that moisture is transferred from droplets less efficiently than heat. Due to the large uncertainty in the specification of \( \beta \) and \( \gamma \), the estimated LHF\textsubscript{spray} values could be in error by several factors. However, even if the actual values were at the high end of their uncertainty range, LHF\textsubscript{spray} would not have been a major factor in the overall enthalpy fluxes and budgets for this case.

For all but the P0–P1 interval, LHF\textsubscript{sfc}, LHF\textsubscript{b}, and LHF\textsubscript{i} were nearly identical in value, a result of the close (proxy) relationship between \( q_{sfc} \) and \( q_{ABL} \) (Equation 9). The high value of LHF\textsubscript{i} compared to LHF\textsubscript{sfc} for the P0–P1 interval was due to a positive spike in \( q \) below 50 m elevation at P1, which was apparent in the rawinsonde profile (see Part 1, Figure 4d, blue line). This spike was mostly averaged out in the rawinsonde-derived calculation of \( q_{ABL} \) (which integrated \( q \) up to \( z \) at 294 m for LHF\textsubscript{i}). However, this feature had a strong enhancement effect on LHF\textsubscript{sfc} (which was based on the \( q_{i} \) value). The spike was not a random turbulent (sampling error) feature because the value of \( q_{i} \) remained near the higher spike \( q \) value (0.34 vs. 0.29 g kg\(^{-1}\)) for several minutes after the P1 rawinsonde launch (Part 1, Figure 3d). However, the origin and physics of this spike were not known; therefore LHF\textsubscript{i} was considered suspect (too high) for this interval.

The estimated LHF\textsubscript{sfc} for the P0–P1 was less than the integral estimates. The amount of open ocean versus grease/shuga bands may have been underestimated for this region. Where there was just open water present (downwind of P1), LHF\textsubscript{sfc} was within \( \sim 10 \) Wm\(^{-2}\) of the LHF\textsubscript{P0–P1} estimate (Figure 3). Unlike P0–P1, for the other intervals, both upwind and downwind upper rawinsonde profiles were available, along with detailed surface and surface layer information, and therefore the estimated errors were considerably less. Indeed, similar to SHF, the three different LHF estimation methods matched very well for the P1–P2 and P3–P4 intervals, being near 165 and 20 Wm\(^{-2}\) respectively for these two intervals (Figure 3).

For the P2–P3 interval, the integral fluxes, LHF\textsubscript{sfc}, LHF\textsubscript{b}, and LHF\textsubscript{i} were greater than the average LHF\textsubscript{sfc}. The differences can be explained by the same reasoning as discussed above for SHF, that is, the surface conditions experienced by air parcels reaching the ship at P3 were different from what the ship observed in the P2–P3 interval due to the slight deviation of the ship course from the directly downwind direction. This meant that the air parcels reaching the ship had been over more humid surfaces than were observed along the ship cruise track. Apparently these surface humidity differences were not as great as the \( T_{sfc} \) differences because the relative differences of the LHF bulk versus integral estimates were less than SHF differences for this interval.

### 3.1.4. Surface Radiation Flux

This subsection examines the surface radiation flux, based on upward and downward-looking measurements at 9 m elevation, as described in Part 1. The direct measurement of downwelling longwave radiation LW\textsubscript{down} \textsuperscript{↓} (Figure 4, blue line) and upwelling radiation LW\textsubscript{up} \textsuperscript{↑}, based on \( T_{sfc} \) (red line) provided an accurate (\( \pm 7 \) Wm\(^{-2}\)) measure of the net surface radiation flux, LW\textsubscript{net} \textsuperscript{↓} (purple line), with positive net values indicating
The net radiation $LW_{\text{sfc}}$ (using Equation 15 and assuming $\epsilon$ equals 0.98) closely tracked (within 1°C) the 21 m air temperature $T_a$ (Part 1, Figure 3a), being highly correlated even at the smallest detectable (turbulent) time/space scales (comparison not directly shown). This indicated that the origins of $LW_{\text{sfc}}$ were within a few meters of the sensors, a result of the dominating effects of atmospheric particulates (steam fog, sea spray and hydrometeors) in controlling $LW_{\text{sfc}}$. Here, the net surface radiation flux $LW_{\text{sfc}}$ was relatively large ($\sim 75 \text{ Wm}^{-2}$) due to the mostly open sea surface with high $T_a$ values.

From 25 to 40 km, $LW_{\text{sfc}}$ began to decrease, even as $T_a$ was increasing gradually (Part 1, Figure 3a), indicating that the $LW_{\text{sfc}}$ sources were coming from higher in the atmosphere, above the sea spray, where the temperature was colder. This horizontal region had the greatest small-scale variability in $LW_{\text{sfc}}$, which can be clearly observed in Figure 4. In this region, pancakes were forming in bands, which intermittently suppressed the sea spray and allowed the surface to "see" the colder regions above the sea spray for brief periods.

Downwind of $\sim 42$ km, $LW_{\text{sfc}}$ had less fine scale variability. There was no longer any sea spray generation (according to the bridge human observations and spray model results, Figure 2) to create small-scale variability in $LW_{\text{sfc}}$. In these regions, different cloud conditions were the major cause of the $LW_{\text{sfc}}$ variability, with radiation sources switching from within the ABL (when cloudy) to above the ABL during clear conditions. The greatest net radiation $LW_{\text{sfc}}$, $\sim 100 \text{ Wm}^{-2}$, occurred within the large lead centered at $x$ equals 68 km, where a warm surface (large $LW_{\text{sfc}}$) was combined with a cold ABL (small $LW_{\text{sky}}$). Near the downwind end of the transect, at $x$ greater than 95 km, the surface radiational cooling $LW_{\text{sfc}}$ was always greater than 50 Wm$^{-2}$ due to clear skies, and it had become a dominant term in the surface heat budget, surpassing even the SHF over the snow-covered young ice floes.

### 3.1.5. Total Surface Flux and Comparison of Components

The surface enthalpy flux section will conclude with a comparison of total surface heat flux (using the bulk estimates) and its components SHF$_b$, LHF$_b$, and $LW_{\text{sfc}}$ (Figure 5). For almost the entire transect, SHF$_b$ the dominant term, accounted for more than 75% of the total heat flux. The patterns of SHF$_b$ and LHF$_b$ were very similar (Figure 5) because both were controlled by the same variations in $T_a$, $\theta_v$, and $U$. The net surface radiation $LW_{\text{sfc}}$ was an order of magnitude less than the turbulent fluxes and its contribution to the total surface flux was negligible, being lost in the noise of the turbulent fluctuations. There was an exception to this flux component dominance ranking near the downwind end of the transect, at $x$ greater than 97 km. Here, $LW_{\text{sfc}}$ became larger than SHF$_b$ and LHF$_b$ because the snow-covered young ice floes were effectively insulating the ocean and the lack of clouds increased the $LW_{\text{sfc}}$ surface cooling. The 2,500 Wm$^{-2}$ estimated magnitude of the total heat flux in the open regions of the TNBP was impressive, enough to create the equivalent of 64 cm of ice per day, assuming there was no oceanic heat flux.
\( H_{abl} \) were the precipitation enthalpy \( H_p \), radiation divergence \( LW_{abl}^{-} \) and (potentially) entrainment of heat \( E_q \) or moisture \( E_v \) from above the ABL. The estimations of the integral fluxes required determination of these other terms. In addition to their direct effects on the overall ABL enthalpy budget, the snow and radiation divergence have other important indirect effects such as impacts on visibility and radiative transfer, cloud formation, and surface snow accumulation.

### 3.2.2. Snow

Snow growth (defined here as any water vapor deposition into ice crystals, or frozen steam fog) within the ABL was represented by the precipitation enthalpy \( H_p \) as determined by Equation 14 and plotted in Figure 6. Although there was some uncertainty in these snow estimates, there was no question that snow had a major impact on the ABL moisture budget. Neglecting \( H_p \) would have resulted in LHF and LHF values that were lower than the uncertainty range estimates for the mean LHF, for both the intervals 0–21 and P1–P2. The largest magnitude \( H_p \) values (maximum 186 W m\(^{-2}\)) were over the open areas at \( x \) less than 27 km. However, the \( H_p \) enhancement was less than the LHF and SHF enhancement in this region (Figure 3). This was because sea spray (perhaps counterintuitively) suppressed snow growth by enhancing SHF more than LHF and therefore increasing the \( q \) vapor storage capacity (most right term in Equation 13) more than the moisture source term (LHF, in Equation 13). The greatest relative \( H_p \) (compared to LHF) occurred at \( x \) between 27 and 40 km where more than 50% of the moisture coming from the surface (represented by LHF) was removed by snow. This region was dominated by small pancakes, which had the highest Bowen ratios.

Although a major term for the ABL moisture budget, snow \( (H_p) \) contributed less than 10% of the total ABL enthalpy input, because of the dominance of SHF. When the snow fell into the ocean and melted or directly formed sea ice, the resulting enthalpy loss from the ocean \( F_{snow} \) was only 13% \( (L_f/L_v) \) of \( H_p \), which was an insignificant contribution (maximum 24 W m\(^{-2}\)) to the overall loss of heat from the ocean for this case. Therefore, although the direct effects of snow on the sensible heat and total enthalpy budgets of the atmosphere and ocean were within the uncertainty noise of the measurements, the impacts on the moisture budget (e.g., clouds) and surface conditions (snow accumulation) likely had significant longer-term thermodynamic effects in the Ross Sea region.

### 3.2.3. Radiation Divergence

The longwave radiation balance of the ABL (Figure 7) was represented by the overall ABL longwave radiation divergence \( LW_{abl}^{-} \), which was a function of radiation from the surface \( LW_{sfc}^{-} \), radiation from above the ABL \( LW_{abl}^{-} \), the ABL blackbody radiation \( LW_{abl}^{-} \), and the ABL emissivity \( \varepsilon_{abl} \) estimated using the one layer model \( \varepsilon \).

The largest values of \( LW_{abl}^{-} \) occurred in the most upwind regions, where the warm open ocean surface (the effect indicated by \( LW_{sfc}^{-} \)), presence of clouds just above the ABL (indicated by relative large \( LW_{sfc}^{-} \)) and the high \( \varepsilon_{abl} \) due to sea spray and hydrometeors all contributed to radiational ABL warming. In the region \( x \) between 24 and 47 km, \( LW_{abl}^{-} \) decreased in response to the cooler surface temperatures (decreasing \( LW_{sfc}^{-} \)), less upper level cloud cover (decreasing \( LW_{sfc}^{-} \)) and fewer hydrometeors and spray (decreasing \( \varepsilon_{abl} \)). These trends continued so that at \( x \) greater than 47 km, \( LW_{abl}^{-} \) was negative (ABL cooling), except over the open leads where the warm surfaces provided an enhanced radiation source. At \( x \) locations centered at 68, 85, and 99 km, the \( \varepsilon_{abl} \) values were significantly lower due to a lack of clouds, hydrometeors or sea spray in the ABL. The low \( \varepsilon_{abl} \) values caused the ABL to become less responsive to radiational
forcing and drove the LW\_\text{abl}↑ magnitudes to near zero in those locations. These results indicated that the radiation divergence was not a major factor in the katabatic jet enthalpy budget for this case. However, it did affect the ABL temperatures and was included in the integral heat flux estimates. In polar situations with slower wind speeds, thick ice and clouds within the ABL, LW\_\text{abl}↑ could be a significant and even dominant component of the ABL enthalpy balance.

3.2.4. Entrainment

The ABL enthalpy budget (Equation 5) includes terms for entrainment of sensible heat \( E_\theta \) and latent heat \( E_q \) from above the ABL, which could not be directly measured. However, they could potentially be estimated as residual terms in Equation 5, which were the differences between the bulk and integral surface flux estimates. The fact that for most intervals, the integral flux values \( \text{SHF\_ls} \) and \( \text{LHF\_ls} \) were close to the bulk values \( \text{SHF}\_b \) and \( \text{LHF}\_b \) was evidence that entrainment was not important to the enthalpy budget. However, the residual (integral vs. bulk differences) also includes quite large uncertainties, and it was possible an error masked an entrainment effect. Symptoms of entrainment would have been (a) an overestimation of \( \text{SHF\_ls} \) (compared to \( \text{SHF}\_b \)) due to warmer air above the ABL, (b) an underestimation of \( \text{LHF\_ls} \) (drier air above), and (c) increasing \( z_i \). Except for an increase in \( z_i \) between P1 and P2, none of these symptoms were observed. The initial increase in \( z_i \) could have been due to convergence of the wind vector, or a sampling error due to waves or rolls, neither of which requires entrainment. Therefore we conclude that there was no evidence that entrainment significantly affected the ABL enthalpy budget. However, due to various uncertainties, this could not be proven. This concludes the thermodynamic analyses for this case.

4. Dynamics Theory and Methods

4.1. Introduction

This section is an analysis of the katabatic jet dynamics, with focus on the downwind evolution of the jet wind vector. Similar to the thermodynamics analysis (Sections 2 and 3), there are two main goals: (a) to quantify the surface momentum flux and (b) to understand the momentum budget within the jet. This subsection will describe the theory and methods used to analyze this case; the next subsection presents the results of the analysis, based on the measurements described in Part 1.

4.2. ABL Momentum Budget

By having the ship steam directly downwind, and assuming steady-state conditions, similar assumptions and simplifications as were made for the enthalpy fluxes and budgets were applied to momentum. The average ABL wind vector \( \mathbf{U}\_\text{ABL} \) (bold indicates vector quantity, wind speed, and direction) was considered to be a single parameter for each horizontal location and the change in momentum of a parcel within the jet ABL was determined from the changes in \( \mathbf{U}\_\text{ABL} \). A Lagrangian coordinate system aligned in the downwind direction was defined so that the \( x \) component of wind vector was equal to the average ABL wind speed \( U\_\text{ABL} \), and the \( y \) (crosswind) component of wind speed \( V\_\text{ABL} \) was equal to zero. The horizontal momentum budget equations for these ABL averages can be expressed as

\[
\frac{dU}{dt} = \mathbf{U} \cdot \frac{\partial \mathbf{U}}{\partial x} = -\frac{\tau}{\rho} - \frac{1}{\rho} \frac{\partial P}{\partial x} + E\_\text{mx} \tag{42a}
\]

\[
\frac{dV}{dt} = \mathbf{U} \cdot \frac{\partial \mathbf{U}}{\partial y} = \frac{1}{\rho} \frac{\partial P}{\partial y} + E\_\text{my} \tag{42b}
\]

where \( P \) is atmospheric pressure, \( f \) is the Coriolis parameter, \( \tau \) is the surface wind stress, and \( E\_\text{mx} \) and \( E\_\text{my} \) represent entrainment from above the ABL. In Equation 42 and for the remainder of this section, the subscript “ABL” is dropped and any variable without a subscript represents an ABL vertical mean value. The left-hand equals sign in Equation 42a represents steady-state conditions. The downwind (Equation 42a) terms are (a) the change in momentum, (b) Reynolds stress divergence, (c) horizontal pressure gradient, and (d) vertical or horizontal advection, hereafter referenced as the inertia, friction, pressure gradient, and entrainment terms, respectively. The crosswind terms represent (a) sideways acceleration, (b) the Coriolis “force,” (c) the crosswind horizontal pressure gradient, and (d) entrainment of crosswind momentum.
The four intervals between P0 to P4 were used to quantify the gradient terms. The atmospheric density $\rho$, and surface wind stress $\tau$, were based on horizontal means (using the parcel-time-over method) of every available measurement, while the other variables were determined at the interval endpoints, as will be described next.

To quantify the $U$ values in the inertia terms (LHS Equations 42a and 42b) and Coriolis term, one method would have been to use the mean vertical wind vector within the ABL $U$ from the direct rawinsonde measurements (corrected for ship blocking effects) at each location P1, P2, P3, and P4. This was the procedure used to estimate $U$ at P1, where the ship mast wind sensor was nonfunctional. However, the rawinsonde measurements were subject to sampling errors when collected data were assumed to represent mean values. Similar to $\theta$ and $q$, the ship mast wind vector $U_{33}$ was a good proxy for $U_{ABL}$. Therefore, $U_{33}$ and WD$_{33}$, averaged for 10–20 min, were used, instead of the rawinsonde-derived values, to estimate the $x$ and $y$ momentum components in Equation 42 for locations P2, P3, and P4. The only information derived from the rawinsonde measurements at these locations were the $z_i$ values. This was analogous to the procedure used to estimate the surface integral enthalpy fluxes SHF$_s$ and LHF$_s$ to reduce sampling errors for $U$. The wind vector $U$ at P0 (Manuela AWS) was estimated by extrapolating the 20-min-average measured wind vector at 3–33 m elevation using surface layer theory, for the time period just prior (parcel travel time) to the rawinsonde launch at P1.

4.3. Quantifying Terms

The differences in $U$ values between adjacent stations P0–P4 were used to determine the average wind speed gradient for each station interval while the average of the two station values were used as the “stand-alone” $U$ in the steady-state inertia and Coriolis terms in Equation 42.

In theory, measurements of the inertia and pressure gradient terms in Equation 42 (and assuming entrainment was zero) could be used to estimate the surface drag term and the value of $\tau$, $C_{DZ}$, and $z_o$, as was done for the integral enthalpy flux estimates. However, unlike the enthalpy, momentum was potentially affected by a pressure gradient term, and was also more sensitive to uncertainties in the steady-state and one-dimensional simplifications. For these reasons, it was not possible to estimate surface momentum fluxes using an integral method for this case. Instead, the surface drag term in Equation 42a was estimated based on the bulk method described in Section 2. By using the detailed specifications of $C_{DN10}$ and direct $U$ and $z_i$ measurements described previously, it was estimated that the drag term could be determined with an estimated uncertainty of 15%.

The downwind pressure gradient term in Equation 42a for each interval was initially estimated near the surface using the rawinsonde air pressure values during the pre-launch adjustment period. The crosswind pressure gradient term could not be estimated from the $in situ$ measurements; initially it was assumed to be equal to zero.

4.4. Ice Breeze

The strong SHF caused the ABL to become warmer in the downwind direction, which tended to decrease the surface pressure, creating a thermal wind. This was the “ice breeze” effect, which was analogous to a sea breeze, but flowing toward the ocean. The ice breeze effect for intervals P1–P4 was determined based on the differences in ABL average density $\rho$, difference across each interval, integrated from the surface to $z_i$. The density was determined using the ideal gas law based on the ship proxies for $\theta$ to reduce sampling error. There was no upper air information at P0 to determined $z_i$. However, because the change in ABL density was primarily due to SHF, an alternative method was used to estimate the ice breeze effect for this interval. By assuming hydrostatic equilibrium, the change the change in surface pressure $P_{sfc}$ over time is

$$
\frac{dP_{sfc}}{dt} = \text{SHF} \frac{g}{C_{D} \theta_K}
$$

where $\theta_K$ is the ABL potential temperature (K). Applying the bulk formula for SHF (Equation 18a) and converting to a downwind space derivative by dividing by $U$, results in an expression for the ice breeze pressure gradient

...
\[ \frac{1}{\rho} \frac{\partial P}{\partial x} = \frac{C_{H} (\theta_{sfc} - \theta) g}{2 \theta_{k}} \]  

(44)

where \( C_{H} \) represents an ABL heat transfer coefficient, determined using surface layer theory, extrapolated to 33 m which was assumed to represent the ABL values. The factor of 2 in the denominator appears because surface heating has no effect on the pressure gradient at \( z_{i} \), therefore the average ice breeze pressure gradient throughout the ABL is one-half the surface value. Note that because the same \( U \) that was assumed to advect the air parcels (Equation 1) was also used in the bulk estimate of SHF (\( U_{33} \) in Equation 18a), wind speed \( U \) does not explicitly appear in Equation 44 (although it does have a minor effect on the value of \( C_{H} \)). Also note that \( z_{i} \) does not appear in Equation 43 because the surface pressure change does not depend on where (vertically in the atmosphere) the heat from SHF is transported. Because only surface temperature measurements are required in Equation 44, it is well-suited for estimating ice breeze effects based on satellite or aircraft remote sensing data, when the details on the ABL state variables are not available.

4.5. Entrainment

The final RHS terms in Equations 42a and 42b, \( E_{mx} \) and \( E_{my} \), represents entrainment of \( x \) and \( y \) momentum from above the ABL, the former assumed to be negative because any entrained momentum would be less than within the jet. It was not possible to directly quantify these terms. The earlier thermodynamic analysis showed that entrainment was not required to explain the ABL enthalpy budgets; it will be assumed to be zero for momentum also, with the understanding there is some uncertainty in that assumption.

4.6. Analytical Solution

If the pressure gradient and entrainment terms in Equations 42a and 42b are set to zero, then \( U \) and wind direction \( WD \) can be solved analytically as a function of downwind distance \( x \) (using Equation 18c to substitute for \( \tau \) in Equation 42a),

\[ U = U_{0} e^{-x d_{i} / Z_{i}} \]  

(45)

\[ WD = WD_{0} - \arctan \left( \frac{\dot{x}_{i}}{U} \right) \].  

(46)

where the “0” subscript denotes the upwind values. Equation 46 is valid for relatively small \( x \), such as the distance between rawinsonde stations.

5. Dynamics Results

5.1. Surface Wind Stress

The estimated wind stress (downwind surface momentum flux) \( \tau \) as a function of trajectory distance from the ice shelf edge \( x \) shown in Figure 8 was based on measurements of \( U, \theta, q, \) and \( T_{sfc} \) using the bulk method described previously in Section 3. Before discussing the results shown in Figure 8, some caveats and explanations are needed. As discussed previously, \( U \) upwind of 54 km was based on a linear interpolation in \( x \) space between the measured values at P0, P1, and P2. Surface roughness (parameterized by specifying \( C_{DN10} \)) was specified according to surface (sea ice) type categorized by using the value of \( d_{i} \), as described in Section 3. The Palmer did not get closer to the ice shelf edge than 17 km; for this region, \( C_{DN10} \) was set to the average value for \( x \) between 17 and 24 km where surface conditions were mostly open ocean with some grease/shuga bands.

Because of the various methods used to estimate the state variable and surface parameters required for the bulk method, the variability of \( \tau \) due to spatial surface variability was not captured in Figure 8 for \( x \) less...
than 17, nor were midscale and mesoscale effects on $U_{33}$ measured for $x$ less than 54 km. These $x$ values mark the two logistical (as opposed to natural) boundaries between regions with (a) an almost straight-line increase in $\tau$ where the values represented realistic mean values for those general $x$ locations, but with none of the natural variability, (b) where the $\tau$ values change rapidly in response to changing surface roughness conditions and represent averages along the ship track, and (c) where the additional variability due to wind gusts provides a realistic “snapshot” of wind stress variability that was not shown in the other locations on Figure 8.

The $\tau$ pattern in Figure 8 appears similar to $U_{33}$ (Part 1, Figure 3e) with a general decrease in $\tau$ from the maximum values of 2.9 Nm$^{-2}$ in the most upwind location to 0.7 Nm$^{-2}$ over the smooth snow-covered ice floes in the most downwind locations. This large overall decrease in $\tau$ was caused by the decrease in the mean value $U_{33}$ over the domain of this case, representing the single largest source of overall variability of $\tau$. The major differences between the $U_{33}$ and $\tau$ patterns were caused by the surface roughness, with $C_{DN10}$ varying from 1.3 x 10$^{-3}$ over grease/shuga to 2.8 x 10$^{-3}$ for rough young ice, resulting in a corresponding change in $\tau$ at the same (sometimes very small) scales as the $T_{sfc}$ variations (Part 1, Figure 3a). On mesoscale space scales (several kilometers), the $\tau$ values were relatively depressed (compared to $U$) in the region $x$ between 30 and 45 km. This was because this region had large amounts of grease/shuga and pancakes that were smoother (lower $C_{DN10}$) than either the open ocean or young ice surfaces. There was a region of enhanced $\tau$ located 50–57 km downwind corresponding to the center of the accumulation zone (Part 1) where the sea ice was highly compressed and distorted, creating the roughest observed surfaces. Additional variability in $\tau$ (for locations greater than $x$ equals 54 km) shown in Figure 8 was due to the wind gustiness, estimated to create 20% variability in $U_{33}$ which translates to an ~44% variability in $\tau$. This value was not as great as the effect of surface roughness variability. However, gustiness was still a major factor contributing to the instantaneous wind vector variability.

The surface temperature $T_{sfc}$ affected the estimated $\tau$ values because of its effect on surface layer stability, represented by $\psi_{m}$ in Equation 19. However, despite the large surface heat flux values, the surface layer stability effect (a function of $L$) increased wind stress by no more than ~5% because the buoyancy effects on turbulent momentum transfer (which were proportional to $U$) were far surpassed by the mechanical production effects (which were proportional to $U^3$) for this high wind speed case. The use of $T_{sfc}$ for determining $d_{sc}$ (Equation 22) had a much greater impact on $\tau$ (due to $C_{DN10}$ changes) than the effects of $T_{sfc}$ on surface layer stability (Equation 19).

### 5.2. ABL Momentum Budget

#### 5.2.1. Downwind Component

This subsection quantifies and analyzes the terms of the simplified katabatic jet ABL downwind momentum budget (Equation 42a) for this case. The methods for estimating each of the terms in Equation 42a were described in Section 4.1. The ship measurements of air pressure P indicated a decrease of almost 5 hPa from P1 to P4 (not shown), a distance of 72 km. The operational Antarctic Mesoscale Prediction System (AMPS) forecast fields for this period predicted a pressure gradient of approximately of 3 hPa from P0 to P4, in the downwind direction, a distance of 99 km. Using either one of these values (especially the ship measurements) for the pressure gradient term in Equation 42a would have caused a forcing imbalance in the positive $x$ direction, that is, the air parcels in the ABL would not have slowed down as much as was observed. The AMPS also predicted (not shown) warm air advection from the southwest during the time period of the ship transect in the region between 700 and 1,500 m elevation, a feature verified by the rawinsondes (Part 1, Figure 4c). This caused P to decrease everywhere in the Terra Nova Bay (TNB) region, complicating the attempts to estimate the pressure gradient term. Although the upper-level or external (synoptic) pressure gradient was uncertain, the horizontal gradients of $\theta$ in the ABL meant that an ice breeze effect must have been present; therefore this effect was included as a pressure gradient forcing term. For this analysis, a residual term was calculated by subtracting the estimated friction and ice breeze terms from the inertia term in Equation 42a. The residual term represents the synoptic (nonice breeze) pressure gradient plus any errors in the measurements or assumptions used in the estimation of the other momentum terms, including the zero entrainment assumption.
The magnitude of the estimated inertia, friction, ice breeze, and residual terms for each of the four intervals between P0 and P4 are plotted in Figure 9. The sign convention in Figure 9 is negative compared to Equation 42a, that is, the slowing wind speed and opposing friction term are plotted as positive magnitudes, while the ice breeze term (which acts to speed up the wind) is indicated as negative in Figure 9. A positive residual term indicates that the air parcels slowed more than expected based on the known terms. The most remarkable result of this case was the close correlation between the inertia and friction terms (Figure 9). This result indicates that the effect of horizontal pressure gradient on ABL air parcel accelerations was weak, and Equation 45 was approximately applicable for this case. The large decrease in the friction and (resulting decrease in the) inertia terms between the first two intervals was due to both (a) an increase in \( z_i \) and (b) a decrease in surface friction \( \tau \), as \( U \) decreased. Between P2 and P3, \( U \) and \( \tau \) continued to decrease. However, \( z_i \) also decreased, resulting in a friction term that did not significantly change. Between P3 and P4 another \( z_i \) decrease more than counteracted the wind speed (stress) decrease, resulting in a slight increase in the friction term over this last interval. All these changes were well-matched by the inertia term, indicating their close relationship and lack of external pressure gradient forcing.

The ice breeze effect shown in Figure 9 was significantly smaller in magnitude than the inertia and friction terms, being at most 12% of the magnitude of the friction term, directed in the opposite direction. Including the ice breeze caused the residual term to be larger for the P0–P1 interval. However, including the ice breeze effect decreased the residual for the more downwind momentum balance estimates.

The residual term was also relatively small, indicating that either the external pressure-gradient forcing and errors were small, or that they had somehow counteracted each other. An explanation for the positive residual value in the P0–P1 interval could be that there was some entrainment into the ABL, which slowed the air parcels and caused the observed increase in \( z_i \). Farther downstream, where the ABL depth had adjusted to the surface conditions, such entrainment would be expected to decrease or even for detrainment to occur, which would explain the observed residual term pattern in Figure 9. However, given the uncertainties, this explanation is highly speculative, and similar to the enthalpy budget analysis, an entrainment effect was not required to explain the results for this case.

### 5.2.2. Crosswind Component—Wind Direction WD

The magnitudes of the Coriolis terms in Equation 42b were close in value to the inertia and frictional drag forces, being within 10% of their values for the last three intervals. However, this term was directed orthogonally to the \( U \) vector, and therefore only affected WD, not \( U \) and therefore is not shown in Figure 9. The close match in magnitudes between the Coriolis term and the inertia and friction terms was a coincidence that was due to the particular combination of \( f, z_i, U, \) and \( C_f \) that occurred for this case, and was not a significant or universal result. The Coriolis force caused the air parcels to turn to the left (decreasing WD) going downwind (Figure 10). The ABL average wind direction from the rawinsondes WD, and ship mast level WD, compared well with the predicted values using Equation 46, where the initial value WD, used in Equation 44 was from the most upwind rawinsonde measurement at P1. The good match between the WD changes predicted by Equation 46 and the measurements, indicated that the crosswind component of momentum was primarily controlled by earth’s rotation and any crosswind pressure gradients were

![Figure 9](image-url)

**Figure 9.** Atmospheric boundary layer downwind (x component) simplified momentum equation terms (Equation 42a) as a function of \( x \). The terms are inertia (blue circles), friction (green diamonds), ice breeze (red triangles), and a residual (purple squares) that represents the unaccounted for pressure gradients and uncertainties in the values of the other terms.

![Figure 10](image-url)

**Figure 10.** Wind direction (WD) as a function of \( x \) as measured from the ship 33-elevation mast WD (black dots), the atmospheric boundary layer (ABL) vertical average from the rawinsondes WD (colored circles), and as predicted assuming the air parcels follow the curvature of an inertial radius (red line) using Equation 46. The upwind ice shelf was approximately perpendicular to 285°.
weak. The ship WD$_{33}$ measurements indicated that there was large variability in the 1-min average values. However, at longer time/space scales, the WD$_{33}$ values were $\sim 5^\circ$ greater than the ABL average values, an expected result as frictional effects closer to the surface inhibited some of the rotational turning in the upper ABL, an effect also seen in the rawinsonde wind direction measurements (Part 1, Figure 4b).

The WD at P3 (green circle in Figure 10) was less than the prediction and most of the concurrent WD$_{33}$ measurements. At the time of the P3 rawinsonde launch, there was a microscale gust with an anomalously southerly (crosswind) component. This was an example of why using the wind vector $U_{33}$ based on a several-minute mean was more likely to better represent the average ABL wind at the locations of the rawinsondes than the actual rawinsonde measurements of $U$.

5.3. Scale Analysis

Some features of the jet dynamics can be elucidated by examining the values of some commonly used scaling parameters. A Froude number, Fr, for this case can be defined by

$$Fr = \frac{U}{\sqrt{g' z_{\text{mid}}}}$$

(47)

where the reduced gravity $g'$ is

$$g' = \frac{(\theta_{\text{mid}} - \theta_{\text{ABL}})g}{\theta_h}$$

(48)

and $z_{\text{mid}}$ is the elevation midway between $z_i$ and $z_{\text{top}}$, representing the depth of the longest and most important gravity waves, $\theta_{\text{mid}}$ is the potential temperature at $z_{\text{mid}}$. The reduced gravity $g'$ represents the background or external stabilizing influence. This Fr is the ratio of $U$ and the average phase and group velocity of the dominant gravity waves associated with the katabatic wind $c_{\text{kat}}$.

$$c_{\text{kat}} = \frac{\sqrt{g' z_{\text{mid}}}}{\theta_h}$$

(49)

For this case Fr decreased from 5.2 to 2.7 during the intervals between P0 and P4, indicating that, although the jet momentum was considerably reduced, the flow remained supercritical or "shooting" (Ball, 1956; Renfrew, 2004) for the entire transect, with no katabatic (also known as "hydraulic") jump (e.g., Vignon et al., 2020), as would occur if Fr became unity anywhere. Gravity waves traveled at speeds $c_{\text{kat}}$ slower than the wind speed within the jet and thus were not able to efficiently dissipate momentum and adjust the mass upstream in this situation, allowing the katabatic wind air parcels to travel several tens of kilometers over the ocean. Parish and Bromwich (1989) also observed supercritical flow over the TNB. In that case, aircraft measurements tracked the katabatic winds out to at least 250 km from the shelf break. During the initial phases of this PIPERS case and other katabatic wind events, Fr would have had to reach unity, starting near the ice edge and moving seaward as the winds increased. The reverse would have to happen as the wind event slowed. However, no obvious katabatic jumps, as indicated by sharp changes in wind vector, were observed at any time during the PIPERS cruise.

The Rossby and Ekman numbers $R_o$ and $E_k$ (defined as the magnitudes of the inertia term or friction terms, respectively, divided by the Coriolis term in Equation 42) were between 0.9 and 1.9 for all intervals (not shown). These values indicated that, although the Coriolis term was important, the flows were highly ageostrophic, due to the large air parcel decelerations and weak horizontal ($x$ and $y$) pressure gradients.

Two length scales were relevant for the PIPERS katabatic wind event dynamics: the frictional adjustment length $L_M$ (representing the e-folding scale of the frictional wind speed decrease, Equation 45), and the inertial radius $L_{IR}$ (representing the radius of inertial oscillations, Equation 46)

$$L_M = z_i / C_D$$

(50)

$$L_{IR} = U / f.$$ (51)

The value of $L_M$ for each of the four P0–P4 intervals was 144, 193, 141, and 106 km and for $L_{IR}$ was 229, 193, 162, and 139 km, respectively. These scales closely matched the observed parcel decelerations and curva-
tudes, a result of the weak pressure gradients as discussed earlier. Seaward of P4 (x greater than 99 km) there were no measurements from the Palmer for this case. However, satellite images and AMPS model analyses indicated that the air parcels were starting to be affected by pressure gradients associated with marine synoptic phenomena, and the above length scales would no longer have been relevant.

The Rossby radius of deformation $L_R$, defined as $L_R = L_t / F_s$, is a measure of the horizontal length over which mass adjusts toward geostrophic balance in response to an imbalance, and defines the horizontal dimensions of many mesoscale phenomena in the atmosphere and ocean, including low-level jets in the Arctic (Guest et al., 2018). For this case, $L_R$ was 45 ± 5 km at all locations, which does not correspond to any actual downwind observed length scale. The supercritical flow in this situation prevented mass adjustment; therefore $L_R$ was irrelevant in the downwind direction. In the crosswind direction, gravity waves could propagate, and the size of the crosswind transition zone between the jet and the surrounding geostrophically balanced atmosphere in the absence of topographic influences would be expected to scale with $L_R$. However there were no crosswind in situ measurements to verify this for this case.

### 6. Summary, Conclusions, and Discussion

Many of the results regarding the physics of the katabatic winds over the TNBP presented in this study, such as the supercritical flow, weak synoptic forcing, and strong surface heat fluxes were anticipated as early as Bromwich and Kurtz (1984) and have been emphasized in many subsequent studies. However, there has been very little direct in situ verification of the various physical processes taking place over the TNBP or any ACP. This study addressed that shortcoming and also presented new results regarding the $n$ of sea spray and air humidity that may be applicable to other ACP cases.

The total surface heat flux was estimated to be 2,500 ± 500 Wm$^{-2}$ (80% confidence interval) over the open areas of the TNBP, which is comparable (or even greater) than values reported for the strongest TCs or polar lows. Unlike TCs, SHF dominated over LHF, contributing 80% of the total surface heat flux. The air over the TNBP was ∼30% denser than air within a TC due to the cold temperatures, with a resultant relative enhancement of all the surface fluxes. The ocean mixed layer within the TNBP was at the freezing value or colder and therefore virtually all of the heat loss from the ocean surface resulted in ice formation. As sea ice formed thicker floes and the surfaces became cooler, the SHF and LHF were reduced. Over snow-covered young ice, the turbulent fluxes were reduced enough that longwave radiation dominated the surface heat budget. The three methods for estimating the surface heat fluxes were significantly closer in value than was expected (in most locations) based on the estimated uncertainties (greater than 25% for each method), which was strong evidence that sea spray played a significant role in the heat fluxes. Had the sea spray effect not been included in the bulk estimate of SHF$_p$, its estimated value would have been significantly less than the two integral methods SHF$_a$ and SHF$_e$.

Further evidence suggesting sea spray enhancement of SHF in over ACPs comes from the September 19, 2012 case by Went a and Cassano (2020). Although they did not publish any SHF estimates, the values displayed in their Figure 7 can be used to estimate SHF using the integral method, Equation 40a, and ignoring radiation, entrainment and snow effects. The resulting SHF$_a$ estimates were 3,240, 2,360, and 60 Wm$^{-2}$ over surfaces of mostly open water, transition and sea ice covered regions respectively. Given the surface temperatures, such high fluxes for the first two intervals would not have been predicted using a bulk method unless a significant sea spray effect was included. Interestingly, if a surface temperature of −1.7 was assumed for the first (most upwind and open) interval of the 19 September Wenta & Cassano case, the resulting ABL sensible heat coefficient $C_{th}$ (calculated by dividing SHF$_a$ by $\rho C_p (T_{sfc} - \theta_{ABL})$) would be $2.2 \times 10^{-3}$ which was almost identical to the $C_{th}$ value used for the P0–P1 interval for this case. This represents a more than doubling of SHF due to sea spray compared to classic interfacial theory predictions.

The author views the integral estimates of enhanced SHF values (compared to values calculated using classical interfacial flux parameterizations) from this case and Went a and Cassano (2020) as strong evidence that sea spray plays an important role by enhancing marine heat fluxes over ACPs, and, by extension (however, with more uncertainty), other marine high wind situations, such as polar lows or TCs. As demonstrated with the Went a and Cassano (2020) case described earlier and based on the year-round measurements at Manuela AWS and other coastal Antarctic locations, we know that higher wind speeds and colder tem-
perature katabatic wind events than this case sometimes occur over ACPs, with resulting surface heat loss even greater than the large values estimated for this case. This demonstrates the need for GCMs to include special parameterizations for ACPs. Because of their small size, the warm surface temperature signature of the ACPs may be smoothed by the surrounding cold region areas that are still within the same GCM grid cell. In many situations, such smoothing does not cause large errors in mean surface fluxes because in high winds the heat flux responds linearly in response to changes in the air-surface temperature difference. However, in the case of open water ACPs, the more than doubling of SHF due to sea spray would not be properly represented by just using the average surface temperature for grid points that contain ACPs and assuming only interfacial surface fluxes are present.

Although LHF was only marginally significant in terms of the overall surface heat flux, the TNBP represented a major moisture source for the coastal Ross Sea region increasing the ABL specific humidity \( q \) by a factor of 6. The moisture flux potentially had far-reaching indirect effects as a storm energy source and by causing phenomena such as snow, steam fog, suspended ice crystals (“diamond dust”), frost flowers, and clouds, which affect radiation transfer within the atmosphere and at the surface.

The surface flux SHF was mostly responsible for the increase in temperature in the downwind direction, with small radiation effects (less than 5%) and no detectable entrainment. Snow growth in the upper part of the ABL contributed as much as 180 W m\(^{-2}\) to the sensible heat budget, which, although small (9%) compared to SHF, was an important term in the ABL humidity budget, removing up to 50% of the moisture provided from the surface moisture flux. The occurrence of snow was inevitable because of the nonlinearity of the saturation vapor pressure with respect to temperature and may be a universal feature of all similar high wind situations over open water regions with air temperatures below \(-25^\circ\text{C}\), unless there is a large amount of dry advection or entrainment. The related observation that the air within the surface layer was always near the ice saturation value for this case facilitated the moisture budget analyses used in this study and was a major factor controlling the LHF.

The surface wind stress \( \tau \) field was highly variable on many space (and time) scales due to variations in wind speed and surface roughness. This \( \tau \) variability likely had impacts (although in ways that are poorly understood) on the ocean surface layer circulation and features such as windrows, ice bands, leads, ice deformation and polynya outline shape. In the fluid surface zone (\( x < 41 \) km, where the surface was open or covered with grease, shuga or uncompact pancakes), free drift (ice movement proportional to mean \( u_0 \)) could be a reasonable first-order approximation, at least on scales greater than a few kilometers. The downwind variations in \( \tau \) would cause ice in free drift to vary in amount (ice crystal thickness) by plus or minus 30%, with the overall average decrease in wind speed (stress) creating a 15% increase in ice amount for these regions. In the accumulation zone (\( x \) between 41 and 60 km), the compressive effects of internal ice stress far surpassed the local average wind stress gradients and the ice thickened by a factor of 3 times or more than it would have from thermodynamic growth alone. Outside of the accumulation zone (\( x \) greater than 60 km), the young ice floes were thicker than 30 cm and resisted compression; therefore thermodynamic growth again became important, although rafting, ridging, lead formation, and other deformations caused by wind stress and internal ice stress variations remained significant factors that could alter regional mean and local ice thicknesses.

Previous studies have demonstrated the importance of synoptic forcing in the regions of katabatic wind jets (e.g., Knuth & Cassano, 2011; Parish & Bromwich, 1987), including the TNB (Parish & Cassano, 2003a, 2003b; Wenta & Cassano, 2020). However the results from this case indicated that any pressure gradients resulting from synoptic events were weak and the airflow was dominated by local forcing. Some other studies of the TNB regions had similar results. For example, Bromwich and Kurtz (1984) stated that while synoptic scale motions may modify the extent of the TNB katabatic wind events; they are not the primary cause of its formation or maintenance. The aircraft flight measurements over the TNB described by Parish and Bromwich (1989) revealed light winds above the jet regions, similar to this case and another example of weak synoptic forcing in the region of the katabatic wind “runout” over TNB. Bromwich et al. (1993) conclude that internal dynamics primarily controlled the TNB katabatic wind behavior, based on an AWS analysis of the region. These results suggest that simple formulas such as Equations 45 and 46 can serve as reasonable first-order approximations of the wind field over TNB during many katabatic wind events.
For this case, the ice breeze effect counteracted the frictional slowing by 12% in the most upwind (high heat flux) regions where the air-sea temperature difference (and SHF) was greatest. Lower wind (stress) cases would have greater relative ice breeze effects. The wind speed prediction (Equation 45) could be adjusted for ice breeze effects by including another factor in the exponent (using Equation 44), in this case that would be 0.89; it would be smaller (indicating greater relative sea breeze effects) in lower wind cases.

Despite the progress that has been made from this and previous research efforts, considerable uncertainty remains regarding the dynamic and thermodynamic forcing over the TNBP and ACPs in general, particularly with regard to (a) the sea spray microphysics and associate surface and droplet freezing effects, (b) the role of coherent structures such as rolls and gravity waves within the jet, (c) the importance of entrainment, subsidence, and convergence in controlling the vertical extent of the jet and associated ABL depth, and (d) the ocean and internal sea ice processes controlling the surface conditions. Efforts to address these issues will need to rely on various numerical, laboratory, and remote sensing techniques, in addition to the few available in situ measurements. For these reasons, the results presented in this study could serve as a “control” case for use in numerical or remote-sensing studies that will address the unknowns regarding the physics of katabatic winds and ACPs.

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

Data Availability Statement
The PIPERS data used to support the results and conclusions of this study are available at the U.S. Antarctic Program Data Center: http://www.usap-dc.org/view/project/p0010032 Manuela AWS data are available from the University of Wisconsin at http://amrc.ssec.wisc.edu/aws/index.php?region=Reeves%20Glacier&year=2017&mode=uw.

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