Inside Katabatic Winds Over the Terra Nova Bay Polynya: 1. Atmospheric Jet and Surface Conditions

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Abstract The atmospheric and surface conditions during a late autumnal strong katabatic wind event were quantified using ship and rawinsonde measurements over the Terra Nova Bay coastal polynya. Wind speeds decreased from 35 to 18 m s⁻¹, while the wind direction decreased 18° over the distance from the Nansen Ice Shelf edge out 99 km eastward toward the Ross Sea. Maximum velocity winds (jet cores) at 173, 238, 179 and 144 m elevation were associated with atmospheric boundary layers capped by temperature inversions with bases at 294, 325, 226 and 196 m elevation. The tops of the inversion layers (near 700 m at all locations) also marked the top of the katabatic wind layer. Boundary layer air temperature and specific humidity increased from −31 to −21°C and 0.1 to 0.6 g kg⁻¹, respectively, in response to the warm polynya surface. The air at 15 ± 8 m elevation was saturated with respect to ice, causing supersaturation and snow growth where the air parcels become cooler in the upper atmospheric boundary layer. The surface was characterized by three zones, a fluid zone (open ocean, frazil, shuga and pancake floes), an accumulation zone (fused, rafted and compressed pancake floes) and a young floe zone (large floes). The surface temperature varied from freezing (−1.7°C) in the fluid zone to near air temperature (−20°C) in the floe zone with the largest horizontal surface temperature gradient occurring in the transition between the fluid zone and the accumulation zone, and at the edges of leads in the floe zone.

1. Introduction

Katabatic winds are a common feature of the Antarctic coastline (King & Turner, 1997; Parish & Bromwich, 1987). Longwave radiation from the Antarctic ice sheet cools the snow surface and removes heat from the near-surface air, creating ubiquitous, strong, low-level temperature inversions (Connolley, 1996). The shallow cold air flows downhill and is funneled into coastal valleys, where the sloping terrain allows the air parcels to gain momentum from gravity. In many situations, particularly on the East Antarctic Coast, the katabatic winds peter out quickly beyond the generating slopes due to surface friction, gravity wave dispersion (Renfrew, 2004; Vignon et al., 2020), opposing thermal winds from the build-up of cold air at the slope bottom (Vihma et al., 2011), opposing synoptic forcing (Seefeldt et al., 2007), or entrainment from above the jet (Manins & Sawford, 1979). In some situations, the synoptic pressure gradient forcing is favorable (downstream) or weak (Bromwich et al., 1990; Seefeldt & Cassano, 2012; Seefeldt et al., 2007) or the winds are decoupled from the surface (Andreas et al., 2000) and katabatic winds are able to propagate hundreds of kilometers across flat ice shelves and ocean surfaces.

The Antarctic katabatic winds have been the source of considerable attention from the time of the first explorers due to their major impacts on navigation, visibility and human survival. The katabatic winds often form strong jets that are characterized by low-level, high-wind-speed cores that are surrounded by regions of strong vertical and horizontal wind shears. Katabatic jumps, or rapid increases in wind speed associated with the onset of a katabatic wind event (Ball, 1956), are common phenomena in many Antarctic coastal regions (e.g., Vignon et al., 2020). Katabatic winds can extend over the ocean and trigger mesoscale cyclones and other atmospheric features that affect regions far from the initial katabatic wind source (Bromwich, Carrasco, et al., 1993; Carrasco & Bromwich, 1994; Gallée, 1995, 1997; Minnett & Key, 2007). These offshore winds play an important role in the overall Antarctic vorticity balance through their interactions with decaying synoptic cyclones (Egger, 1991; James, 1989).

Antarctic katabatic winds are strongly modulated by synoptic pressure patterns (e.g., Knuth & Cassano, 2011; Wenta & Cassano, 2020) and in many cases what are usually considered to be “katabatic” winds...
are, in reality, synoptically driven wind fields that have been channeled by topography, giving the appearance of being katabatic drainage flows (Parish & Cassano, 2001, 2003a, 2003b; Renfrew & Anderson, 2002). For this reason, it might be technically more correct to use the more general term “downslope” winds, rather than “katabatic”, when referring to topographically steered high-wind events in general. However, this paper will use the term “katabatic” loosely (as it has been in many references) to refer to any situation where a shallow cold air mass has gained downhill momentum.

An important effect of katabatic winds in Antarctica is the creation of coastal polynyas (e.g., Renfrew, 2006). Wind stress over the coastal ocean removes newly formed ice, creating open-water and thin-ice coastal polynyas which expose relatively warm surfaces to the cold air. The resulting gain of heat and moisture in the lower atmosphere can create phenomena such as plumes and mesoscale cyclones that can impact cloud cover and regional circulation patterns (e.g., Renfrew, 2002).

Katabatically forced coastal polynyas have two major impacts on the ocean that can extend well beyond the local region: (a) sea ice formation and (b) deep water formation. The large surface heat fluxes from the polynyas cause the formation of prodigious amounts of sea ice (Barber & Massom, 2007; Petrelli et al., 2008) which is transported into the Antarctic pack ice surrounding the continent; for this reason, these polynyas are often called “sea ice factories” (Gordon & Comiso, 1988). This additional sea ice can affect regional and global climate by changing the surface albedo and modifying ocean and atmospheric circulation patterns.

Another large-scale effect of the heat fluxes within the Antarctic coastal polynyas (ACPs) created by katabatic winds is the creation of dense water due to the brine rejection resulting from ocean freezing (Mathiot et al., 2012). This dense water flows out from the bottom of the continental shelves and, while undergoing some mixing on the shelf break, some of it sinks to the deep sea floor and spreads throughout the world ocean as Antarctic Bottom Water, AABW (Fusco et al., 2009; Gordon, 2001; Orsi & Wiederwohl, 2009; Rusciano et al., 2013). This is the deepest and most dense water mass on Earth, and a major factor in ocean thermohaline circulation and associated climate effects (Jacobs, 2009). The formation rate of AABW has changed in recent years; this is related to changes in ice formation in the Ross Sea ACP conditions (Abrahamsen et al., 2019; Silvano et al., 2020). However, understanding these changes in detail requires better quantification of the heat exchanges and other processes occurring within the coastal polynyas where the AABW formation is initiated.

Despite their potential global impacts, the actual surface areas of the ACPs that form in response to katabatic winds are relatively small (even when including areas covered with new ice), with areas on the order of 1,000 km² or less for a single polynya (e.g., Bromwich & Kurtz, 1984). This small size is below the resolution of current general circulation models (GCMs) that are used to predict climate change and potential ACP global impacts. Some of these processes can be resolved with regional forecast and research models such as the Antarctic Mesoscale Prediction System (AMPS) (Powers et al., 2003). Regional models with accurate and fine scale topography generally do well at predicting the timing and general characteristics of katabatic wind events. However, even with the higher spatial resolution of these models, they are still challenged by the representation of surface heat and moisture fluxes over ACPs for several reasons. The biggest challenge is accurately representing the surface conditions, in particular the surface temperature of the sea ice and exposed ocean $T_{se}$, which is highly variable in ACP regions. For example, Wenta and Cassano (2020) showed that AMPS had errors in the low-level atmospheric structure predictions due to incorrect $T_{se}$ calculations. Another challenge for numerical simulations of ACPs is the representation of a variety of microphysical processes such sea spray production, sea spray freezing and sea ice formation, all of which can potentially have major impacts on the surface fluxes.

Ever since Captain Robert Falcon Scott’s Northern Party was stranded on Inaccessible Island over the winter of 1912, the Terra Nova Bay (TNB) has been notorious for its frequent and intense katabatic wind events and the associated coastal polynya (Bromwich & Kurtz, 1982, 1984). Despite being one of the harshest environments on the surface of the Earth, there has been a considerable amount of geophysical research devoted to studying the TNB katabatic winds and the associated coastal polynya, using data from satellite remote sensing, unmanned weather stations, ocean moorings, manned and unmanned aircraft and a variety of numerical simulations.
The Terra Nova Bay Polynya (TNBP) is easily detectable using various satellite products, due to the distinctive thermal, visible and microwave signatures of the open water compared to the surrounding sea and land ice-covered areas. Early satellite studies verified the frequency and consistency of the TNBP openings (Kurtz & Bromwich, 1983) and their relation to katabatic winds (Kurtz & Bromwich, 1985), while more recent efforts have developed techniques for quantifying surface temperature (Aulicino et al., 2018; Clappa et al., 2012) and other surface characteristics such as ice type (Hauser et al., 2002; Hollands & Dierking, 2016) at high resolution spatial scales and frequency.

While satellites are excellent at providing detailed surface information, they are limited in their ability to detect state variable (wind vector, temperature, humidity, pressure) fields in the lowest parts of the atmosphere. Katabatic wind presence and horizontal dimensions can be detected by warm surface “streaks” (Bromwich, 1989b) or sastrugi orientations (Bromwich et al., 1990) on the glacial ice surface; however, for the most part, quantitative information on the wind, temperature, or various water phases in the atmosphere and associated surface fluxes over the TNB and the TNBP has come from modeling efforts (e.g., Bromwich et al., 1990; Cassano & Parish, 2000; Davolio & Buzzi, 2002; Gallée & Schaye, 1994; Van Woert, 1999), plus some strategically placed automated weather stations (Bromwich, 1989a; Bromwich, Parish, et al., 1993) and a few Sodar measurements (Argentini et al., 1995).

The only published atmospheric measurements above the surface layer in the TNB region have come from a small number of instrumented aircraft missions. Parish and Bromwich (1989) used a manned airplane in two flights to confirm the cold (bora-type) nature of the low-level jet air mass, which had a warm surface signature due to vertical mixing. The katabatic air was tracked for 250 km over the Ross Sea, as it gradually slowed and mixed vertically. The most detailed information available to date concerning low-level atmospheric features over the TNBP comes from a series of flights performed in two separate field programs using small unmanned aerial vehicles (UAVs) (Cassano et al., 2010, 2016; Knuth et al., 2013; Wenta & Cassano, 2020). These studies showed that the katabatic winds over the TNBP were concentrated in low-level jets, with maximum wind speeds occurring at 200–300 m elevation. During the stronger wind speed events, there were well-mixed atmospheric boundary layers (ABLS) that extended from the surface to near or slightly above the wind maximum level (jet core). The ABLS were significantly modified in the downwind direction due to surface heat fluxes.

The PIPERS (Polynya and Ice Production and seasonal Evolution in the Ross Sea) field program provided an opportunity to address some of the shortcomings in our knowledge of the atmospheric and surface conditions associated with the TNBP and ACPs in general. PIPERS was a multi-disciplinary effort to understand sea ice production in the Ross Sea and the various associated atmospheric, ice physics, oceanic and biogeochemical processes (Ackley et al., 2020). A variety of measurements were performed in the austral fall of 2017 during a two-month cruise in the Ross Sea on the ice-capable research vessel R/V Nathaniel B. Palmer. Virtually all of the data directly used for the results presented in this paper were from a 12-hour period while the Palmer was performing a downwind transect across the TNBP during an intense katabatic wind event.

The purpose of this paper is to describe the rarely sampled internal characteristics of the state variables (wind vector, temperature, humidity) within a katabatic wind event and how these variables evolve downwind in response to surface conditions. A companion paper (Guest, 2021) builds on the material presented in this paper and uses the same data to perform a dynamic and thermodynamic analysis of the katabatic wind event and will be referred to as “Part 2”. This paper, “Part 1”, is arranged as follows: after this introduction (Section 1) is an overview and measurement strategy (Section 2), measurement description (Section 3), results (Section 4), and conclusions and comparisons (Section 5).

2. Measurement Strategy and Case Overview

2.1. Measurement Strategy

The strategy for sampling the Terra Nova Bay katabatic wind events was to have the R/V Palmer steam as fast as possible downwind across the TNBP, while performing in situ ship (surface and near-surface) and rawinsonde (upper-air) measurements. The purpose was to approximate a 2-dimensional (by steaming downwind), steady-state (by transiting quickly) situation, to simplify the analysis and interpretations of the
events. This strategy was employed for three katabatic wind events: on May 1, May 5, and May 8, 2017. This paper will focus on the 8 May event, referred to as “this case”.

2.2. This Case - May 8, 2017

This case was chosen because it best met the desired criteria of being a steady-state and 2-dimensional situation (i.e., with vertical and downwind variations only). The operational AMPS 16-hour model run prediction of the surface wind field for this case (Figure 1) shows a channeled downslope and offshore flow pattern that is typical for katabatic wind events in the TNB region. The cold air that formed over Victoria Land ice sheet (to the left of Figure 1 scene) was funneled down the Reeves glacier valley, across the Nansen Ice Shelf and over the TNB. A synthetic aperture radar (SAR) image of the same scene, Figure 2 (zoomed slightly and rotated from Figure 1), shows more clearly the coastline and surface features within the TNBP during this case, including the Drygalski Ice Tongue to the south. The locations of the rawinsonde sounding stations, labeled P1, P2, P3 and P4, and connected with a red line, and the Manuela Automated Weather Station (AWS), P0 (Figures 1 and 2), were aligned with the surface wind vector. The Nansen Ice Shelf forms the western edge of the TNBP, and is indicated by a line (colored black in Figure 1 and white in Figure 2) that extends generally southwestward from location P0, which is on the south edge of Inexpressible Island. Note that there is an unreal “ghost” ice shelf edge image that appears in Figure 2 image just to the east of the indicated (white line) Nansen Ice Shelf edge, where the edge was before a recent calving event.

To the east (just off the right side of Figure 1), the streamlines merge with the northeastward outflow of the Ross Ice Shelf air stream (RAS), a persistent and common feature in this region (e.g., Parish et al., 2006; Steinhoff et al., 2009). The RAS forms the southwestern quadrant of a synoptic-scale cyclone centered on the outer fringes of the Ross Sea (not visible in Figure 1). This situation represented a good “default” case that was relatively unaffected by synoptic forcing in the key region over the TNBP. It was similar to many previously documented events (Bromwich, 1989a, 1989b; Parish & Bromwich, 1989, Wenta & Cassano, 2020).
Using the wind vector knowledge, the location of the measurements and knowledge of the ice shelf location (Figure 2), a new independent variable was created: parcel trajectory distance from the ice shelf edge, indicated by the symbol “x”. The modest curvature of the parcel trajectories meant that the variable x used in this paper was slightly larger (3% maximum) than the actual straight line distance to the ice shelf.

2.3. Ship Operations

The original cruise plan was to bring the R/V Palmer to within 1 km of the Nansen Ice Shelf edge before beginning a rapid downwind transect. However, while approaching the edge heading into the wind, extreme freezing sea-spray conditions overtaxed the Palmer bridge window wipers and de-icing systems, essentially blinding the ship crew and forcing the halt of progress at P1, 17.5 km away from the ice shelf edge (Figures 1 and 2). We were able to successfully launch a rawinsonde and collect good data at this location before beginning the downwind transect. We attempted several rawinsonde measurements at both 28 and 37 km from the ice edge; however, the launches were unsuccessful or did not produce accurate data due to the extreme conditions. However, at P2, P3 and P4 (x equals 54.5, 83.8 and 99.2 km) “clean” rawinsonde data were obtained. Due to the need for multiple rawinsonde launch attempts and the difficult ship maneuvering during these times, the entire transect (from P1 to P4) lasted almost 12 hr. However, throughout this period and domain, the AMPS numerical-produced fields of the state variables below 700 m elevation were virtually constant, providing confidence that the changes observed between the different locations could be attributed to spatial, not temporal, variations, within the estimated errors.

3. Measurements

The measurements used directly for this paper are listed in Table 1. Launching rawinsondes from the Palmer during near hurricane-force winds required innovative procedures. Just prior to launch, the ship was maneuvered so that the wind direction was 15 degrees off the starboard bow, which provided some protection from the wind blast at the helicopter deck launch location. Due to the intentional ship blockage, the measured wind speeds were suspect below 120 m elevation. When available, the ship measurements at 33 m (when available) provided lower-level wind vector information. The pressure levels were converted to height z levels using the hypsometric equation.
The extreme icing conditions due to sea spray and hydrometeors required the use of specially designed measurement equipment (Table 1). The Heitronics CT 15.10 IR surface temperature sensor was attached to a 1 m weather housing tube and pointed 50° downward from a boom that extended out from the starboard side of the ship. This housing worked well at keeping ice and water away from the sensor windows. This measurement was affected by sea spray and hydrometers in the path between the 9-meter sensor height and the surface. This contributed an additional sampling error of 1.1°C to the overall uncertainty of the instantaneous \(T_{\text{sfc}}\) measurements. Unfortunately, despite being heated to prevent icing, the sonic anemometer was nonfunctional due to icing at the start of this case study and did not become fully functional until the time of the second successful launch at P2, almost 6 hr later. The other measurements listed in Table 1 were operational during the entire period of this case.

The results of an error analysis of all the measurements, based on laboratory calibrations and intercomparisons of different sensors (during this case and other periods during PIPERS) with the results are summarized in Table 1. Because of natural variability, virtually instantaneous rawinsonde measurements may not represent mean conditions for a particular launch location. Therefore, in addition to the measurement uncertainty (based on measurement accuracy) an additional “sampling error” (based on environmental variability) is included in Table 1 for the rawinsonde measurements. Note that these sampling errors apply to estimates of mean values for a particular sounding. The actual measurement accuracies (i.e., no sampling

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**Table 1**  
**Measurements and Estimated Uncertainties**

| Instrument parameters measured | Uncertainty | Manufacturer | Model | Location elevation |
|-------------------------------|-------------|--------------|-------|--------------------|
| Rawinsonde 100 gm He balloons<sup>a</sup> | - | Vaisala | Digacora MW41 RS-41 sondes | Helicopter deck 12 m–10 km profiles |
| **Pressure** | 0.2 hPa | - | - | - |
| **Air temperature** | 0.8°C | - | - | - |
| **Specific humidity** | 0.078 gkg<sup>-1</sup> | - | - | - |
| **Wind vector** | 2.1 m s<sup>-1</sup> | - | - | - |
| IR radiometer 8–14 μm w/ weather housing<sup>a</sup> | - | Heitronics | CT 15.10 | boom, 9 m |
| **Surface temperature** | 1.4°C | - | - | - |
| Thermometer/Hygrometer radiation shielded<sup>b</sup> | - | Vaisala | HC2 | Boom 9 m |
| **Air temperature** | 1.0°C | - | - | - |
| **Specific humidity** | 0.05 gkg<sup>-1</sup> | - | - | - |
| Thermometer/Hygrometer radiation shielded<sup>b</sup> | - | Rotronic | HC2A-SS | ship tower 21 m ASL |
| **Temperature** | 1.0°C | - | - | - |
| **Specific humidity** | 0.05 gkg<sup>-1</sup> | - | - | - |
| Ultrasonic anemometer Heated<sup>b</sup> | - | Gill | Wind Observer II | ship tower 33 m |
| **Wind vector** | 0.5 m s<sup>-1</sup> | - | - | - |
| Barometer<sup>a</sup> | - | Vaisala | PTB110 | helo deck 9.5 m |
| **Surface pressure** | 0.5 hPa | - | - | - |
| Barometer<sup>b</sup> | - | Vaisala | PTB210B | dry lab 7 m |
| **Surface pressure** | 0.5 hPa | - | - | - |
| Manuela automated weather station (AWS)<sup>c</sup> | (various) | (various) | Inexpressible Island 3 m |
| **State variables** | (similar to ship) | - | - | - |
| Human observers | - | - | Bridge 15 m |
| **Ice characteristics** | - | - | - | - |
| **Visibility** | - | - | - | - |
| **Weather conditions** | - | - | - | - |

Note. Systems deployed and maintained by:
<sup>a</sup>Naval Postgraduate School.  
<sup>b</sup>R/V Palmer Antarctic Support Contract (ASC).  
<sup>c</sup>University of Wisconsin and partners (Lazzara et al., 2012).
error included) were similar to the ship measurement accuracies. Further details on the measurements and error analysis are available in the Supporting Information S1 associated with this paper.

4. Results

4.1. Surface Characteristics

4.1.1. Surface Temperature and Ice Cover Relationship

The availability of the R/V Palmer presented a unique opportunity to observe and measure in fine detail at close range the surface and near-surface atmospheric conditions in a coastal polynya during a katabatic wind event. This subsection will examine the surface conditions, in particular how the surface temperature, humidity and ice types varied along the downwind ship transect across the polynya. There was a strong relationship between surface temperature $T_{sfc}$ (Figure 3a) and surface ice conditions. The $T_{sfc}$ variability and associated surface characteristics can be categorized by three zones: (a) a fluid zone, (b) an accumulation zone and (c) a floe zone, as indicated in Figure 3. Within the fluid zone (x less than 41 km) sea ice was rapidly forming and becoming thicker and more concentrated due to intense heat loss to the atmosphere. Frazil (small suspended ice crystals) that formed within the upper water column would rise and concentrate at the surface, creating grease ice, with a characteristic smooth surface. The frazil and grease ice were concentrated in windrows (small bands) that were more or less aligned with the wind, a result of ocean convergence, due to Langmuir cells or other similar small-scale ocean circulations. The bands were 5–140 m wide and spaced 200–400 m apart. At the center of the ice bands, roughly spherical (0.5–4 cm diameter) slushy clumps of partially consolidated ice called shuga formed within the soupy grease ice. The larger shuga clumps sometimes rose above the more liquid surroundings, particularly at wave crests, pushing the surface temperature below the freezing point, reaching near −5°C in the most concentrated shuga regions. By x equals 28 km, the shuga had started to congeal into small pancake floes with surface temperatures as cold as −10°C. At the most upwind ship observations at P1 (x equals 17 km) the surface consisted of an approximately even mix between open ocean areas and ice bands. Downwind, the surface area covered by the ice bands increased until by 41 km most of the surface was covered with pancake floes surrounded by a soupy, slushy mix of frazil and shuga. The signature of the ice bands can be seen in the pattern of rapid variations in $T_{sfc}$ (Figure 3a). Within the accumulation zone (x between 41 and 59 km) most of the ice thickness growth was mechanical rather than thermodynamic. Here the ice was concentrated and thick enough to transmit internal ice stress from the ice pack outside the TNBP into the newly formed ice, causing ice convergence. In the upwind parts of the accumulation zone, near x equals 41 km, pancake floes were rafting and fusing into larger floes. This increased the ice thickness and decreased $T_{sfc}$ to −15°C (Figure 3a). Farther downwind, at x equals 47 km, the fused pancakes had become tipped sideways and compressed together, forming a distinctive rough surface that had the appearance of huge, overlapping scales. It was dubbed “dragon skin” and that will be the term used for this and other types of highly compressed and deformed young ice that existed in the accumulation zone. Within this zone, $T_{sfc}$ was relatively cold (closer to the air temperature than sea temperature) and with less horizontal variability because little open water or thinner ice was present.

At the x equals 60 km location, a few kilometers downwind of P2, the ship moved out of the ice “plume”, formed by the TNBP, into the floe zone. The ice in the floe zone consisted primarily of relatively large young ice floes that had formed in the Ross Sea south of TNB. Although the floe zone ice was much older than the ice in the accumulation zone, it was approximately the same thickness and therefore was in the same $T_{sfc}$ range, near −17.5°C. However, the pattern of $T_{sfc}$ variation was different. Within the floe zone, internal ice stresses caused leads to form. A particularly notable lead or small polynya was crossed between 67 and 71 km where open water was present and $T_{sfc}$ was equal to −1.7. This and other smaller leads created the upward spikes in $T_{sfc}$ shown in Figure 3a that characterize the floe zone. Toward the end of the transect at x greater than 71 m, some of the ice floes had up to 4 cm of snow on the surface. The snow effectively insulated the surface and kept the average $T_{sfc}$ within 3°C of the air temperature.

The fluid, accumulation, and floe zones in coastal polynyas are easily detectable in various satellite products and have been identified in other studies, with different terminology. For example, Hollands and Dierking (2016) used the terms “open water polynya zone” and “deformation belt”, while Fiedler et al. (2010) use “inner region” and “outer region”. In both cases, these are identical to “fluid zone” and “accumulation zone”
used in this paper. The fluid, accumulation and floe zones are present in most coastal polynyas regions, but the actual location and extent of the zones varies from case to case depending on the wind, current and temperature fields, existing ice thickness, geography, etc. Despite this variability in the specific locations of the zones for a particular coastal polynya situation, it is reasonable to expect that the same specific surface and ice types observed during this PIPER case within each zone occur in the same zones for most other cold-air high-wind coastal polynyas as well. The often-used Pease (1987) equation for estimating coastal polynya downwind size (using the Pease default collection depth 10 cm) and heat transfer coefficient $C_h = 2 \times 10^{-3}$,
and the PIPERS air temperature of ~25°C) predicts an equilibrium 17 km wide polynya, which closely matched the observed 50% ice concentration observed at location P1, also 17 km downwind.

4.1.2. Surface Humidity

The surface specific humidity \( q_{\text{sfc}} \) values (Figure 3b) assumed that over open ocean, \( q_{\text{sfc}} \) equaled 98% of the saturation value over pure water, defined as \( q_{\text{sfc sat}} \) (Fairall et al., 2003), while over solid snow or ice surfaces, \( q_{\text{sfc}} \) equaled the ice saturation value, \( q_{\text{sfc sat}} \) (Andreas et al., 2010). Pancake ice was observed to have partially wet surfaces due to wave breaking, wave crest compression (squirting) and sea spray. The effects of partially wet surfaces are represented by a wetness factor \( \alpha_{\text{wet}} \), defined as the portion of liquid water at the surface. The value of \( \alpha_{\text{wet}} \) for pancake ice varied from 5% to 100%, depending on the estimated thickness of the pancake ice. Ice floes thicker than fused pancakes (25 cm) were observed to have dry surfaces. Because the surface specific humidity \( q_{\text{sfc}} \) was, by specification, a function of \( T_{\text{sfc}} \), the horizontal \( q_{\text{sfc}} \) features (Figure 3b) were virtually identical to the \( T_{\text{sfc}} \) features (Figure 3a). The primary difference was because of the non-linearity in the \( q_{\text{sfc}} \) versus \( T_{\text{sfc}} \) relationship, and to a lesser degree, because the colder surfaces were drier (\( q_{\text{sfc sat}} \) is almost always less than \( q_{\text{sfc sat}} \)). Further details on the surface conditions encountered during this case are available as Supporting Information S1.

4.2. Characteristic Vertical Features

The data from the rawinsondes (Figure 4) showed that this case was characterized by several persistent vertical features in the lower atmosphere, including (a) a low-level wind jet, (b) well-mixed \( \theta \) and \( q \) layers, (c) inversion layers (strong \( \theta \) increases) and \( q \) jumps just above the mixed layers, and (d) supersaturated (with respect to ice) air in the mixed layers. The jets had maximum wind speeds \( U_{\text{max}} \) of 29.7, 25.3, 20.0 and 19.0 m s\(^{-1}\) (P1-P4) which occurred at elevations (“jet cores”) \( z_{\text{max}} \) of 173, 238, 179 and 144 m (Figure 4a). The capping temperature inversions had bases at elevations \( z_{\text{b}} \) of 294, 325, 226 and 196 m (Figure 4c) and horizontal lines in other Figure 4 panels). Because for this case the freezing temperature and surface temperature values were such important reference points, the potential temperature \( \theta \) is defined based on the current surface pressure (not 1,000 hPa) and expressed in degrees Celsius. The atmospheric boundary layer (ABL) is defined as the vertical region from the surface to \( z_{i} \). Note that although \( z_{\text{max}} \) and \( z_{i} \) vary between locations, they are correlated, with \( z_{\text{max}} \) being approximately two-thirds of \( z_{i} \) at all locations. The relative humidity with respect to ice \( RH_{i} \) (Figure 4e) was near saturation at the surface, becoming supersaturated in the upper ABL due to the adiabatically cooled air parcels. The wind directions WD backed modestly with height within the ABL, except for P1 which increased slightly (Figure 4b). Above \( z_{i} \), until ~700 m elevation, the actual air temperature \( T_{\text{sfc}} \) (not shown) generally increased with elevation. The top of this “inversion layer” \( z_{\text{top}} \) was also where the wind speeds leveled off and the wind directions became highly variable. This level \( z_{\text{top}} \) marks the top of the vertical region directly affected by the katabatic winds. Unlike \( z_{i} \) and \( z_{\text{max}} \), \( z_{\text{top}} \) did not significantly change in the downwind direction for this case.

4.3. Wind Vector

This subsection will examine the wind vector characteristics using data from (a) the four rawinsonde soundings at locations P1-P4 (Figures 4a and 4b), (b) the Manuela AWS and (c) the ship anemometer at locations downwind of P2 (Figures 3e and 3f). Although well away from the slope of the Reeves glacier, the wind field during this katabatic wind event resembled a gravity current, with a jet core occurring within a dense (cold) layer near the surface, similar in shape to jets over the Ross ice shelf (Seefeldt & Cassano, 2008). The wind speeds at the jet core measured with the rawinsondes \( U_{\text{max}} \) were near (1–2 m s\(^{-1}\)) the concurrent short-term mean 33 m ship tower anemometer \( U_{33} \) values at P2 – P4 (Figures 3e and 4a). All the raw wind speed data down to 30 m elevation are shown in Figure 4a, but below 120 m elevation, the measured values were affected by ship blockage and were too low, especially at P2 – P4. Despite this uncertainty regarding the exact vertical structure of the wind jet, the nearly equal values of \( U_{33} \) and \( U_{\text{max}} \) indicated that most of the positive vertical wind shear associated with the jet occurred below 33 m, in the atmospheric surface layer.

The average wind speeds near the surface decreased from 35 ms\(^{-1}\) at the ice shelf edge (PO Manuela AWS) to 18 ms\(^{-1}\) at P4, 99 km downwind (Figure 3e). The AMPS numerical simulation (Figure 1) indicates that
greater near-surface winds speeds existed at the base of the Reeves glacier; the air parcels slowed a few meters per second over the Nansen Ice Shelf before reaching Manuela and the coastal polynya.

Figure 4. (a) Wind speed U profiles from the rawinsondes (colored lines), ship wind speed $U_{33}$ (colored circles) and Manuela U value at the P1 launch time (cyan circle). The assumed tops of the atmospheric boundary layers $z_i$ are plotted with solid horizontal lines using the same color codes. Also shown is the top layer of the inversion layer $z_{\text{top}}$ (dashed black line). (b) Same for wind direction WD. (c) Same for potential air temperature $\theta$ using ship $\theta_{21}$. (d) Same for specific humidity q using ship $q_{9}$. (e) Same as previous for relative humidity with respect to ice RH$_i$. The Manuela RH$_i$ value was 51% - too low to show. The rawinsonde wind speeds below 130 m, particularly at P2-P4, were affected by ship blockage.

The smaller-scale (less than 5 km) variability in $U_{33}$ was due to local temporal gustiness and was not correlated with any specific local surface features (Figure 3e). The gustiness factor (the maximum wind speed divided by the mean wind speed) was 1.2, which was relatively low compared to similar high wind-speed measurements at lower latitudes, usually in tropical cyclones (e.g., Paulsen & Schroeder, 2005). The shallow ABL mixing layer, compared to the tropics, likely contributed to the low gustiness.
The wind speeds over TNB were much greater than could be explained by the downwind horizontal pressure gradient (Figure 1); therefore the momentum gained on the slope of the Reeves glacier (or higher) was able to carry the air parcels within the jet out over TNB and the Ross Sea for more than 100 km, while still maintaining the density current characteristics.

The average ABL wind direction from the rawinsondes $\text{WD}_{\text{ABL}}$ varied from $267^\circ$ at P1 to $293^\circ$ at P4, a backing of $26^\circ$ or $0.33^\circ$ per km (Figures 3f and 4b). A similar downwind backing occurred at all levels up to $\sim 600$ m. The one-minute-average wind direction from the ship anemometer $\text{WD}_{33}$ (Figure 3f) had a small and midscale variability range of $\sim 16^\circ$ that was not correlated with any known surface variable, similar to $U_{33}$, $\text{WD}_{\text{ABL}}$ was at the low end of the $\text{WD}_{33}$ variations, the former being $\sim 7^\circ$ less than the mean of the latter values (Figures 3f and 4b). The Manuela WD was an outlier that was lower in value (backed) than the other locations (Figure 3f), likely due to the topographic effects of its location on Inexpressible Island.

The moderate wind backing from the surface to the ABL and also at higher elevations within the ABL from P1 to P4 was consistent with other observations of Antarctic katabatic winds (e.g., Renfrew & Anderson, 2006) and is the expected result for a boundary layer in a rotating system (Ekman, 1905). The regional downwind backing of the wind vector was also in response to the Earth’s rotation and was very close to the curvature of an inertial oscillation, a result of weak crosswind pressure gradients, a topic to be addressed in Part 2.

### 4.4. Temperature

This subsection uses the data from the rawinsonde, ship and Manuela AWS to examine the vertical temperature structure and how it evolves in the downwind direction. At all the locations P1-P4, the potential temperatures measured from the ship sensors at 9 and 21 m elevations, $\theta_9$ and $\theta_{21}$, (collectively $\theta_{9,21}$) were closely correlated to each other and to the average potential temperature in the entire ABL $\theta_{\text{ABL}}$ (compare plotted circles with lines in Figures 3c and 4c). In particular, $\theta_9$ was always within the estimated sampling error (0.82°C, Table 1) of the rawinsonde $\theta_{\text{ABL}}$, indicating that $\theta_{\text{ABL}}$ was a good proxy for the entire ABL.

All the rawinsonde $\theta$ profiles (Figure 4c) showed the presence of three atmospheric layers: (a) a vertically well-mixed ABL layer below a well-defined $z_i$ (b) a vertically complicated inversion layer between $z_i$ and $z_{\text{top}}$ and (3) a vertically smooth free atmosphere above $z_{\text{top}}$. Near the surface, $\theta$ varied from $\sim 31^\circ$C at the shelf edge (Manuela) to $\sim 21^\circ$C at P4, 99 km downwind, (Figure 3c). This warming was in response to the surface heat fluxes, which are quantified in Part 2. Interestingly, a similar magnitude warming occurred in the free atmosphere between P1 and P4 (Figure 4c). However this warming was due to advection from the southwest (based on the AMPS analysis) and was not related to the surface fluxes. The close match between the ABL and free atmosphere warming was apparently a coincidence.

The inversion layer also warmed in the downwind direction by similar amounts, on average, as the ABL and free atmosphere at P1 – P4. However, the warming was unevenly distributed by elevation. Within the inversion layer proper, there were internal well-mixed layers with small $\theta$ vertical gradients and small gradient Richardson numbers (not shown) intermixed with very stable layers. These step-like structures could have been caused by intermittent wave breaking in this region of large (negative) magnitude wind shear. Some of the step structures appeared to be capped by cloud layers where cloud top longwave cooling could also contribute to local instability and mixing. The P2 location had a particularly prominent step feature, with a constant $\sim -23.1^\circ$C layer (between the P2 and P3 ABL values) in the region 350 m (just above $z_i$) up to 470 m, with an upper inversion capping that layer. The P2 pattern appears to be a deeper remnant ABL layer that was detrained from the surface when the ABL became cooler.

The same magnitude downwind increase in $\theta$ above $z_i$ (Figure 4c) and also the relative constancy of the wind vector between the inversion layer and the ABL (Figures 4a and 4b) was strong evidence that there was a direct physical connection between these layers. However, the gradient Richardson numbers just above $z_i$ were large (not shown), indicating that turbulent mixing was not occurring between the ABL and inversion layer at the times of the rawinsonde launches. Apparently, intermittent mixing processes such as wave-breaking, local plume penetration or entrainment from longitudinal rolls were occurring at times and locations that were not sampled by the rawinsondes. Subsidence also may have played a role in warming the inversion.
Although generally "well-mixed", the ABL layers had considerable vertical variation in \( \theta \) (\(-0.5^\circ\text{C}\)) on small vertical scales (Figure 4c). The ship measurements also show the influence of turbulent (less than 1 km) and midscale (1–5 km) processes with \( \theta_{z,1} \) fluctuating by 1.8°C within the general overall downwind trend of increasing values (Figure 3c). Although displayed in Figure 3 as a function of \( x \), in reality, most of the \( \theta_{z,1} \) fluctuations were advective, and not directly related to any observed local surface features. An exception was over the large lead at \( x \) between 67 and 71 km, where \( \theta_{z,1} \) was enhanced by 1.5°C compared to the background trend. As with all cold air outbreaks over warm water with capping inversions, longitudinal rolls existed in the ABL (Liu, 2004; Liu et al., 2006) and likely contributed to much of the observed midscale variability. Due to the ship course (downwind instead of crosswind) the signal of the variations caused by the rolls was not distinguishable from other possible causes of midscale variability such as gravity waves, convective features, crosswind variations or variable upwind surface conditions not detected by the shipboard measurements.

The depth of the ABL \( z_1 \) increased from P1 to P2 and decreased downwind of that (Figure 4c). Downwind convergence (wind speed slowing) and entrainment from increased TKE due to surface heat fluxes likely caused \( z_1 \) to increase initially. However, downwind of P2, the sideways spreading out of the jet air mass (as it was no longer as constrained by topography) and the decrease in surface fluxes counteracted the downwind convergence and entrainment effects enough to cause \( z_1 \) to decrease between P2 and P4. The possible presence of gravity waves, longitudinal rolls and/or penetrating convective plumes introduced sampling error to the rawinsonde \( z_1 \) measurements. Therefore a quantitative assessment of the effects of the various processes affecting the value of \( z_1 \) was not possible.

### 4.5. Atmospheric Water

#### 4.5.1. Sea Spray, Hydrometeors and Steam Fog

In the regions with open water at \( x \) less than 27 km, there were large quantities of frozen and liquid hydrometeors from sea spray, steam fog, and precipitation, which reduced visibility to less than 200 m. At bridge level, \(-5\%\) of the sea spray (mostly smaller particles) had frozen into sleet before impacting the bridge windows. However, most of the sea spray was in liquid form, which created severe icing issues for ship operations as the super-cooled liquid spray froze upon impact. What appeared to be snow (white and light, coming from higher up) was distinguishable from the darker, larger sea spray droplets which were more concentrated below bridge level. The generation of sea spray was reduced by a factor of approximately two over the grease/shuga areas compared to the open ocean areas and was virtually zero where the pancake ice became fused into larger floes and the liquid surface was no longer exposed.

#### 4.5.2. Water Vapor

The specific humidity \( q \) had many similar characteristics as \( \theta \), including (a) being generally well mixed within the ABL (Figure 4d), (b) having generally increasing values downwind at all levels (Figures 3d and 4d), and (c) having an almost identical downwind variation pattern (compare Figures 3c and 3d). The close relationship between \( q \) and \( \theta \) was because \( q \) was always near to the saturation value with respect to ice \( q_\text{s} \) in the surface layer (and \( q_\text{s} \) is a function of temperature (Buck, 1981)). The ship measurements of air humidity with respect to ice at 9 m \( RH_{9} \) were always within 1% of 97.5%, while the 21 m value \( RH_{21} \) was almost always within 1.5% of 104% (based on data shown in Figure 3 but not plotted), indicating the supersaturation level (a lifting deposition level for water vapor) was 15 ± 8 m elevation, the uncertainty due to possible bias errors in the humidity measurements.

The ice saturation effect can also be seen in the rawinsonde data (Figure 4e), where the relative humidity with respect to ice \( RH_{12} \) was very close to 100% near the surface at P2, P3 and P4. The P1 profile had an unusual low level moisture spike below 55 m elevation. However, above that, \( RH_{12} \) at P1 was similar to the other locations (between 101% and 110%) up to the respective ABL tops at \( z_1 \). The adiabatic cooling of \( T_\text{air} \) in the ABL caused \( RH_{12} \) to increase with elevation, with the largest supersaturation values occurring near \( z_1 \). The \( RH_{12} \) increase with elevation in the ABL was most apparent at P2, which had the deepest ABL, allowing the greatest adiabatic cooling and therefore the largest increase in \( RH_{12} \) in the ABL. That fact that ice-saturated low-level air was detected by the two ship measurements and all the rawinsondes (for this case
and another PIPERS case, not shown here) provides confidence that this feature was not just a measurement fluke, caused, for example, by sea spray or icing contamination of the sensors.

The presence of liquid droplets in the form of sea spray was an additional moisture source that allowed rapid adjustment of q toward ice saturation. The supersaturation in the upper ABL caused ice crystals (snow) to form and grow. The removal of moisture by snow was just enough to counteract the surface moisture fluxes and keep the value of q near $q_{\text{sati}}$ in the surface layer. The saturation values for both ice $q_{\text{sati}}$ and liquid water $q_{\text{satw}}$ surfaces are nonlinear functions of surface temperature, with higher relative values of $q_{\text{sati}}$ at the higher temperatures. This means that the warm surface will always provide more moisture from surface fluxes than can be counteracted by the increase in temperature (saturation vapor pressure) within the ABL due to sensible heat fluxes unless there is some other process such as advection or entrainment that is drying or warming the air.

The specific humidity in the upper part of the surface layer $q_{21}$ was significantly greater than the average value of q in the rest of the ABL (compare lines and circle symbols in Figures 3d and 4d). This was in contrast to $\theta$, which had surface layer versus ABL differences within the sampling errors (Figures 3c and 4c). This behavior in $q_{21}$ was related to how snow growth and/or entrainment affected q and $\theta$ differently. Temperature (sensible heat) had sources at the surface and in the upper ABL where snow was forming, and potentially from entrainment of the relatively warm inversion layer above the ABL. In contrast, water vapor also had a surface source, but snow growth and entrainment at the upper ABL levels were potential sinks. These ABL sinks, which do not have temperature equivalents, caused the observed differences between mean ABL and upper surface layer (21 m) q values that were not observed in the $\theta$ data.

5. Discussion and Conclusions

The measurements described in this paper provided several insights on the structure and characteristics of a katabatic wind event over the TNBP. For this case study, particular attention was devoted to surface conditions and how they influenced the surface temperature, which is a key parameter controlling the surface fluxes of sensible and latent heat. There were three zones that could be readily distinguished from the ship observations or satellite SAR images: (a) a fluid zone, (b) an accumulation zone and (c) a floe zone, each characterized by specific types of sea ice and with different patterns of surface roughness and surface temperature. Many remote sensing techniques are available that could be used to identify these various surface types and zones, from which the important characteristics such as thickness, surface temperature and roughness could be inferred.

Most of the atmospheric (ABL) warming and moistening occurred over the fluid zone, where the surfaces were near the water temperature. Within the relatively narrow accumulation zone, the thicker ice greatly reduced the surface temperatures. Within the floe zone, leads allowed more ocean heat loss and atmospheric warming than in the accumulation zone, but only enough to increase the temperature and humidity by approximately one-third as much as over the fluid zone, even though the ABL air parcels were over the floe zone for a longer period.

To judge how representative this case was, it was compared other known similar in situ measurements, in particular, the 19 September downwind transect case from Went and Cassano (2020), hereafter referred to as W&C, and also the two other TNB transects performed during the PIPERS cruise on May 1, 2017 and May 5, 2017 (“1May” and “5May”), all of which shared many similarities. For all four transects, at all downwind locations, there were low level jets with cores occurring between 150 and 300 m elevation and enhanced wind speeds occurring up to ~700 m elevation. In most cases, well-mixed ABLs were present, with distinctive inversion caps located at (W&C, 1May) or just above (this case, 5May) the jet core.

All the PIPERS cases exhibited moderate backing of the wind direction and gradual slowing of the mean wind speed in the downwind direction, which implied horizontal pressure gradients were weak and the air parcels within the ABLs were responding primarily to inertial and frictional forcing only. This is in contrast to W&C and other studies (referenced in the introduction) that found synoptic scale processes and associated pressure-gradient wind fields had major impacts on the katabatic winds over the TNBP (and other ACPs).
The large surface heat fluxes and turbulent kinetic energy generation from wind shear that occurred for the TNB cases (i.e., this case, 1May, 5May, W&C) normally would be associated with ABL depths that are well over 1,000 m, such as within polar lows or tropical cyclones. However for these cases, the ABL depths were relatively low and remained low or even decreased in the downwind direction. This was because the injection of cold air at low levels provides an external stabilization (represented by the strength of the temperature inversion above $z_i$) that counteracts the internal generation of TKE (from wind shear above the jet core) that would normally lead to entrainment and growing ABL depths. Also sideways divergence (sinking air) contributed to keeping ABL depths low. When the katabatic winds slacken, the (adverted) stabilization decreases, and the ABL over the TNB deepens, as evidenced by residual upper level inversions in many of the PIPERS and W&C temperature profiles.

For all of the three PIPERS cases, the average temperature in the inversion layer increased at the same approximate rate as within the ABL. It was not clear what the role of subsidence and intermittent mixing with the ABL was in warming the inversion layer. The process by which the ABL and inversion layer communicate in this case, and similar situations, will remain a mystery that needs attention in future studies, ideally those involving large eddy simulations that could resolve the suspected processes.

A notable result from this case (and also from the 1May case) that had not been previously published was the observation that the humidity near the surface was always very close to the ice saturation value at approximately 15 m elevation, with supersaturation occurring higher in the ABL. The supersaturation in the upper parts of the ABL leads to deposition (vapor to solid transition in the form of ice crystals, as snow or forming on existing frozen particles). The snow or ice then falls to the surface, removing moisture and converting moisture enthalpy to sensible heat in the ABL. This is not a universal result, for example the 5May case and the 22 September case reported by Cassano et al. (2016) had near surface humidities below ice saturation. These latter cases had warmer temperatures than this case and the 1May case, which meant the nonlinear saturation humidity versus temperature effects were less important and the moisture increase was never enough for the air to reach ice saturation. It is likely that there is a critical temperature, somewhere between $-25$ and $-30^\circ C$, below which the exposure to warm ocean surfaces will inevitably lead to saturation with respect to ice in the surface layer and to supersaturation and snow growth in the upper ABL.

This concludes the description of the atmospheric jet and surface conditions associated with the katabatic wind event which occurred on May 5, 2017 over the Terra Nova Bay Polynya. A detailed analysis of the sensible heat, latent heat and momentum surface fluxes and ABL budgets for this case is contained in Part 2.

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 paper are available at the U.S. Antarctic Program Data Center: [http://www.usap-dc.org/view/project/p0010032](http://www.usap-dc.org/view/project/p0010032). The 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](http://amrc.ssec.wisc.edu/aws/index.php?region=Reeves%20Glacier&year=2017&mode=uw).

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