# AIR-SEA INTERACTIONS IN THE TERRA NOVA BAY REGION OF ANTARCTICA AS MEASURED BY UNMANNED AERIAL VEHICLES

by

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# This thesis entitled: Air-sea interactions in the Terra Nova Bay region of Antarctica as measured by unmanned aerial vehicles written by Shelley L. Knuth has been approved for the Department of Atmospheric and Oceanic Sciences

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The final copy of this thesis has been examined by the signatories, and we Find that both the content and the form meet acceptable presentation standards Of scholarly work in the above mentioned discipline

#### Abstract

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# Air-sea interactions in the Terra Nova Bay region of Antarctica as measured by unmanned aerial vehicles

Thesis directed by Associate Professor John J. Cassano

In September 2009, several unmanned aerial vehicles (UAV) flew over Terra Nova Bay (TNB), Antarctica to collect measurements of the three dimensional properties of the atmospheric boundary layer. TNB has important implications on the atmosphere, ocean, and cryosphere due to an open water polynya in the region. Within the area of lower sea ice concentrations, significant air-sea interactions occur. Until September 2009, observations of the wintertime atmospheric boundary layer over the polynya had not been collected. The UAVs captured important information on the structure, moistening, and warming of the atmosphere due to the presence of the polynya.

This study seeks to increase our understanding of the air-sea interactions in TNB by estimating and assessing the primary forcing mechanisms of heat exchange in the region. Three flights in September 2009 collected measurements designed to capture the evolution of the atmospheric boundary layer as the continental air passed over the relatively warmer and moister surface of the polynya. This dissertation first analyzes climatological observations to put the field season observations into a broader context, showing September 2009 was an anomalous year. Next, a new and innovative methodology, based only on atmospheric data, estimates heat fluxes over TNB. Finally, UAV and satellite data are synthesized to identify the primary forcing

mechanisms controlling the fluxes in TNB. The structure of the atmospheric boundary layer is also studied as the continental air mass moves over the polynya.

This work is significant for several reasons. First, the UAV data collected are the first insitu observations gathered during the wintertime months over the polynya. This information is key for understanding the significant air-sea interactions that occur during this season. Second, the methodology developed to estimate heat exchange from the in situ UAV data provides a unique approach for quantifying air-sea interactions without bulk flux algorithms that have large uncertainty. Finally, a better understanding of the forcing mechanisms of energy exchange in this region has widespread applicability toward other polynyas in the Arctic and Antarctic, and in areas of cold air outbreaks. This work also has potential to improve the uncertainty of bulk flux algorithms in model output.

#### Acknowledgments

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In the middle of my PhD career, my father passed away suddenly and unexpectedly, and this event has changed my worldview profoundly. He was so proud that I was working toward my degree, and I am stricken with sadness to think he will not be here through its completion. I miss him every day, and I dedicate this work to him because without his support, I would not have even had the courage to try.

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### **Chapter 1: Introduction**

The Terra Nova Bay (TNB) region of Antarctica, located in the Ross Sea sector of the continent, is an area of complex topography that profoundly affects dynamics of the atmosphere, ocean, and cryosphere on a local, regional, and global scale (Kurtz and Bromwich 1985; Bromwich 1989; van Woert 1999; Morales Maqueda et al. 2004) (Figure 1). The Antarctic continent to the west and north, the Drygalski Ice Tongue to the south, and the Ross Sea to the east surrounds TNB. Approximately 1000 km inland of the coast lies the East Antarctic polar plateau, with gently sloping terrain and an elevation of approximately 2-3 km. The polar plateau gives way to mountainous terrain, where the Priestley, Reeves, and David Glaciers connect the plateau to TNB (Figure 1).

During the winter months on the polar plateau, strong radiational cooling occurs as longwave radiation is emitted from the surface without compensating shortwave radiation (King et al. 1998). This cold, high-density air overlying the surface is forced toward lower elevations under the influence of gravity with the presence of even a gentle slope, reaching speeds in excess of 40 m s<sup>-1</sup> (Bromwich 1989; Hauser et al. 2002; Knuth and Cassano 2011). These katabatic winds lose their forcing at sea level, but have been shown to propagate outwards from the coast by 30 km or more (Kurtz and Bromwich 1985). Synoptic pressure systems in the Ross Sea can enhance the katabatic drainage and aid in their propagation eastward (Bromwich and Kurtz 1984). Most air movement between the polar plateau and TNB occurs through the Reeves Glacier, due largely to its orientation (Bromwich and Kurtz 1984) (Figure 1). The katabatic winds are particularly strong in TNB due to a confluence zone just upstream of the Reeves Glacier (Parish 1982; Bromwich and Kurtz 1984; Parish and Bromwich 1987).



Figure 1. Map of the TNB region (top panel) within the Ross Sea sector of Antarctica (bottom panel). MLA represents Manuela AWS station.

When this drainage flow reaches the western coast of TNB, the strong winds advect sea ice away from the coast, producing an area of open water, termed a polynya, along its edge (Morales Maqueda et al. 2004). The Drygalski Ice Tongue to the south prevents additional sea ice from moving into the bay. As the cold continental air passes over the polynya, the water quickly freezes, but the resulting ice is advected to the east as long as the winds persist. For this reason, coastal polynyas are known as sea ice factories due to the large amount of sea ice that is continuously produced and advected offshore (Morales Maqueda et al. 2004). The TNB polynya has been projected to produce approximately 40-60 m of ice per winter, which is approximately 10% of the total annual ice production in the Ross Sea (Morales Maqueda et al. 2004). Very dense water forms due to brine rejection when the cooled ocean waters produce sea ice (Galleé 1997; Morales Maqueda et al. 2004). This high-density water eventually becomes part of Antarctic Bottom Water, with TNB producing approximately 10% of the Ross Sea dense water contribution (Kurtz and Bromwich 1985; van Woert 1999).

As the cold, dry air originating from the continental interior is modified by the warmer and moister surface of the TNB polynya, strong sensible and latent heat fluxes (SHF and LHF) occur (Galleé 1997; van Woert 1999; Morales Maqueda et al. 2004 and others). The TNB polynya varies in size largely with the strength of the offshore winds (Bromwich and Kurtz 1984; Hauser et al. 2002; Ciappa and Budillon 2012), but has been found to have an average area of approximately 1000 km<sup>2</sup> and a maximum area of 3-5000 km<sup>2</sup> (Kurtz and Bromwich 1985; Morales Maqueda et al. 2004). The polynya generally oscillates with a period of 15-20 days. During the winter months, when the katabatic winds are strongest and the polynya is of greatest size, the heat fluxes are similarly largest. As the atmospheric boundary layer (ABL) that originates from the continent passes over the polynya, its properties are modified by the input of energy through heat fluxes at the air-sea interface (Renfrew and King, 2000; Heinemann 2008; Raddatz et al. 2011). The ABL is warmed and moistened with increasing fetch. Along with changing surface temperatures due to changing ice conditions, the vertical temperature and moisture gradients between the atmosphere and surface decrease in magnitude as the air mass moves downstream over the polynya (Heinemann 2008). Heat fluxes are weakened as a result. This simplification assumes the polynya is the primary source of air mass modification, and that entrainment or radiational sources have comparatively smaller impacts on ABL properties.

The structure of the ABL is also modified as the air mass is advected over the polynya. Over the high polar plateau during the winter months, the ABL consists of a strong temperature inversion due to radiational cooling at the surface and a downward sensible heat flux, which cools the atmosphere. Once the air mass moves over the open water, a convective internal boundary layer (CIBL) is generated, wherein the bottom of the inversion layer is slowly eroded as convection ensues due to instability (Chang and Braham 1991; Renfrew and King 2000; Heinemann 2008; Raddatz et al. 2011). Once the continental air mass has passed a sufficiently large distance over the polynya, the remnants of the continental inversion layer are no longer present, and the ABL consists entirely of the CIBL. Entrainment of air at the interface of the ABL and free atmosphere cause the depth of the ABL to increase with increasing distance from the coast (Chang and Braham, 1991; Renfrew and King, 2000; Heinemann 2008). The modifications of the ABL in cold-air outbreaks have been studied in many locations across the globe, including other Antarctic (Kottmeier and Engelbart 1992; Roberts et al. 2001; King et al. 2008) and Arctic (Serreze et al. 1992; Walter et al. 2006; Raddatz et al. 2011) polynyas, the

eastern United States (Grossman and Betts 1990), and during lake effect snow events over the Great Lakes (Chang and Braham 1991).

Despite the importance of the TNB region, few observational studies have been conducted to understand the impacts of the strong heat fluxes on the ABL. Kurtz and Bromwich (1985) estimated monthly mean heat fluxes over open water using a combination of remote sensing and surface meteorological data. They found the average SHF for September to be 615 W m<sup>-2</sup> and the LHF to be 193 W m<sup>-2</sup>, which was the second highest month for both heat fluxes (behind August). The AWS used for estimating the heat fluxes was at Inexpressible Island (now Manuela site, Figure 1), which is on the western coast of TNB and primarily influenced by winds from the Reeves and Priestley Glaciers.

Several modeling studies have also attempted to quantify the heat flux out of the polynya, but have not been confirmed with observations (Morelli 2011; Fusco et al. 2002; van Woert 1999 and others). Van Woert (1999) in particular used a simple polynya model to examine the response of the TNB polynya to local winds. His study found an average SHF of 770 W m<sup>-2</sup> and an average LHF of 250 W m<sup>-2</sup> for May-August 1988-1990 with a 30-40% open water fraction (positive numbers indicate atmospheric heat gain, as is the standard throughout this paper). Fusco et al. (2009) used a combination of observational data and output from the European Centre for Medium-range Weather Forecast (ECMWF) 40-year reanalysis (ERA-40) and ECMWF operational analysis to calculate fluxes over TNB for the period from 1990-2006. This study found an average SHF of 145 W m<sup>-2</sup> over the 17 year period, with an average LHF of 57 W m<sup>-2</sup>, assuming a polynya width of 50 km (the approximate length of the Drygalski Ice Tongue). The Fusco et al. (2009) study uses weekly sea ice concentration values from 1994 to estimate the presence or absence of ice cover throughout the entire 17 year period. It was

presumed in the study that fluxes were smaller in part due to an underestimation of wind intensity from ERA-40. Morelli (2011) uses a mesoscale Eta model to simulate the ABL over the polynya on 15-16 July 2006, finding SHFs of 600 W m<sup>-2</sup> and LHFs of 290 W  $m^{-2}$  with a sea ice fraction of 0.6.

The importance of air-sea interactions on the ABL during the winter months and the lack of observations to quantify these interactions drive the focus of this dissertation. In September 2009, several unmanned aerial vehicles (UAVs) were flown over TNB to estimate the threedimensional properties of the atmosphere. A subsequent field season in September 2012 was also flown with the same mission. These measurements were the first collected over the polynya during the winter months, and provide invaluable information regarding the atmospheric state. Eight missions were flown to TNB, with varying mission goals for each flight. On 18, 23, and 25 September, the mission purpose was to estimate the downstream evolution of the ABL as the air masses originating from the continental interior pass downstream over the polynya. For each of these flights, a south-north transect along the coast and over the polynya was made to search for the area of strongest winds from the continent. The UAV was then flown downstream of the coast within this jet to estimate the downstream evolution of the ABL along its strongest area.

The data from these flights were used to understand the spatial and temporal response of the ABL to the underlying warm polynya during the winter months. To put the observations into context, a climatology of the TNB region using satellite and AWS data was conducted. This study, published in the *Journal of Applied Meteorology and Climatology* ("An analysis of near-surface winds, air temperature, and cyclone activity in Terra Nova Bay, Antarctica, from 1993 to 2009"), composes Chapter 2 (Knuth and Cassano 2011). This work shows that September 2009 is an anomalous year in terms of surface winds, cyclone activity, and upper-level patterns. A

higher number of strong wind events, a shift in cyclone activity eastward, and a corresponding shift in a deeper upper level trough led to these conclusions. As part of this work, an extensive cyclone climatology was conducted over the region. This climatology was used in an effort to evaluate the Antarctic Mesoscale Prediction System (AMPS) cyclone forecasts in the region. This study, entitled "Evaluation of Antarctic Mesoscale Prediction System (AMPS) cyclone forecasts using infrared satellite imagery" was published in *Antarctic Science*. While not a part of this dissertation, the work from Chapter 2 led to important discoveries in model evaluation (Nigro et al. 2011).

Air-sea interactions in the ABL are best examined by understanding the amount of energy transferred between the ocean and the atmosphere, as this exchange will modify the cold, dry continental air in the ABL. Mesoscale and numerical weather prediction models typically use the bulk flux equations to estimate surface heat fluxes, where an eddy diffusivity coefficient for heat or moisture is applied (Fairall et al. 2003; Walter et al. 2006). While these transfer coefficients are based on measurements, they represent a large amount of uncertainty in the bulk flux estimates (Walter et al. 2006). These formulas also rely on surface temperature information, which is highly variable. As such, estimates of the heat fluxes without the use of bulk algorithms have the potential to be more robust and less uncertain. The development of a methodology to estimate SHFs and LHFs from only atmospheric measurements shows this to be true. Chapter 3, entitled, "A methodology for estimating sensible and latent heat fluxes from in situ aircraft measurements", estimates these fluxes by considering adiabatic and non-Lagrangian processes and correcting for unobserved changes in heat and moisture below the minimum flight level of the UAV. Several variations of a single method are tested by estimating the change in the heat and moisture content as the UAV moves downwind over the polynya. Each method produces

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results that differ by less than 5 W m<sup>-2</sup>, showing the robustness of this method. This study has been submitted to the *Journal of Atmospheric and Oceanic Technology*, and is currently under review (Knuth and Cassano 2014).

Chapter 4, entitled, "The primary mechanisms of air-sea heat exchange over Terra Nova Bay, Antarctica", uses the UAV data, satellite data, and a bulk flux algorithm to understand the primary forcing mechanisms of heat transfer in TNB. Changes in the air temperature, atmospheric moisture, and wind speed are compared to surface temperatures, surface saturated mixing ratios, and sea ice fraction to understand the forcing for the surface heat fluxes estimated. The Tropical Ocean-Global Atmosphere (TOGA) Coupled Ocean-Atmosphere Response Experiment (COARE) bulk flux algorithm, which incorporates both surface and atmospheric data, is used to estimate the sensitivity of the heat fluxes to surface state. The impacts on the ABL are also explored by examining the evolution of the temperature structure over the polynya. This chapter is currently in preparation for submission to a peer review journal.

The appendix holds useful information on the design and execution of the sixteen UAV missions from McMurdo to TNB in September 2009, including the three missions used in the studies of Chapters 3 and 4. This chapter also discusses the quality control performed on the data, and is a good reference for the flights throughout this dissertation. The appendix was published in *Earth System Science Data*, and is entitled, "Unmanned aircraft system measurements of the atmospheric boundary layer over Terra Nova Bay, Antarctica" (Knuth et al. 2013). Another study describing the flights and data collected entitled "Observations of Antarctic polynya with unmanned aircraft systems", published in *Eos* can be found in Cassano et al. 2010, although is not included in this dissertation.

# Chapter 2: An analysis of near-surface winds, air temperature, and cyclone activity in Terra Nova Bay Antarctica from 1993-2009

**ABSTRACT.** In September 2009, the first unmanned aerial vehicles were flown over Terra Nova Bay, Antarctica to collect information regarding air-sea interactions. Prior to the field season, wind and temperature data from a local automatic weather station (AWS) were collected from 1993-2007 and compared to an August-October 2006-2008 satellite cyclone analysis to place the September 2009 observations into a broader context. AWS wind data revealed a strong tendency toward downslope flow in the region regardless of season, as the majority (55%) of winds were from the west to northwesterly directions. Most winds observed at the site were less than 20 m s<sup>-1</sup>, but 83% of the stronger winds were associated with downslope flow. Of fifteen strong wind events (greater than 20 m s<sup>-1</sup> for more than 10 hours) evaluated during the cyclone analysis period, 100% occurred in the presence of a cyclone in the adjacent Ross Sea. Winter experienced the greatest number of strong wind events (68%), while summer had the lowest (4%). Most temperatures were between -15 and -25°C, with temperatures influenced by wind fluctuations. The cyclone analysis revealed 64% of systems were comma-shaped, and most cyclones (84%) within the Ross Sea were mesocyclones. A comparison of AWS data for Septembers 1993-2007 and September 2009 showed more strong wind events during 2009, while the cyclone analysis revealed a shift in cyclonic activity eastward. Reanalysis data comparing September 1993-2007 and September 2009 shows an eastward shift in a deeper upper-level trough, indicating September 2009 was an anomalous year.

#### 1. Introduction

Since the early 1900s, when Robert Falcon Scott's Northern Party spent an unfortunate winter on Inexpressible Island, the Terra Nova Bay region of Antarctica has been recognized as an area of harsh winds and brutally cold temperatures. These meteorological extremes make this a region where scientific research is both difficult to conduct and of great importance. The severe winds, caused by offshore downslope flow through nearby mountain valleys, transport sea-ice eastward from the coast and create a relatively ice-free area within Terra Nova Bay. These ice-free windows into the ocean, termed polynyas, can exist throughout the winter season, making the region important for the alteration of oceanic currents, the production of phytoplankton, the congregation of oxygen-breathing sea mammals, and the movement of ships carrying supplies for various Antarctic scientific stations (Buffoni et al. 2002; McMahon et al. 2002; Arrigo and van Dijken 2003 and 2004).

Despite the small size of Terra Nova Bay, the interaction between local atmospheric and oceanic forcing can have significant effects on the local and large-scale circulation of the Southern Hemisphere. The atmospheric effects – largely from downslope drainage from the interior of the continent – can alter existing weather patterns or enhance energy fluxes between the ocean and atmosphere (Bromwich 1989; Bove and Paolo 2009). The oceanic effects – largely caused by the open ocean water – can change currents in the southern oceans (Buffoni et al. 2002). A well-documented coastal polynya that is highly sustainable during the winter season is located at the base of the Nansen Ice Sheet in Terra Nova Bay, and is largely maintained by strong downslope flow from air rushing through the valleys of the Reeves, and to Priestley, and to some extent David Glaciers (Bromwich and Kurtz 1984; Petrelli et al. 2008; Kern 2009) (Figure 2).



Figure 2. Map of Terra Nova Bay, with a broader area of the Antarctic continent in the inset.

Oceanic sensible heat fluxes are also thought to contribute to maintaining the ice-free polynya (van Woert 1999).

During the winter season, when the downslope flow is the strongest, intense winds blowing across the open water can cause large fluxes of energy out of the ocean and into the atmosphere. The injection of warmer ocean air from the polynya into the atmosphere can cause a convective surface layer to form, initiating the development of mesoscale dynamic systems, such as land-breezes or cyclones (Gallée 1996; Carrasco et al. 2003; Bove and Paolo 2009). Cold atmospheric temperatures and strong winds can further the development of this convective layer.

As well, the polynya is an important area for the production of sea ice, as the denser, saltier water left behind when the ice forms and moves eastward sinks to the bottom of the ocean and adds to the formation of Antarctic Bottom Water (Hauser et al. 2002; Kern 2009). Despite the relatively small area of the polynya (during the winter the size varies between 3000 and 7000 km<sup>2</sup>), the impacts on the climate system are significant, as the formation of Antarctic Bottom Water influences the Southern Ocean thermohaline circulation (Hauser et al. 2002).

There is a subtle but important difference between katabatic and downslope flow, as the two are often mistaken as the same. Any flow down sloping terrain can be described as downslope flow. Winds blowing down the terrain can be caused by a number of factors, including a background pressure gradient force or gravity acting on negatively buoyant air adjacent to the slope. Katabatic flow is downslope flow of radiatively cooled, negatively buoyant air (Parish and Cassano 2003). Radiational cooling from the surface, particularly during the winter, will generate a strong surface inversion, which will in turn produce a pressure gradient force directed down the fall-line. It is the presence of radiatively cooled, negatively

buoyant air that distinguishes katabatic flow from the more general downslope flow. As such, katabatic flow is always downslope flow, but downslope flow is not always katabatic. In this paper, we will attempt to discriminate between the two types of flow, although it is important to note that from wind measurements alone, it is not possible to determine whether a flow is katabatic or simply downslope (Parish and Cassano 2001). In these cases, the more general term of "downslope flow" will be used. A detailed analysis of whether the flow is downslope or katabatic requires detailed knowledge of the thermodynamic evolution of the near surface air over the Antarctic continent and is not possible from the limited observational data available over the continent and as such is beyond the scope of this work. Since the development of the coastal polynya is due largely to the presence of strong offshore (and downslope) winds the distinction between katabatic and the more general downslope flow is inconsequential for this study.

Despite the impacts of the Terra Nova Bay region on the local and regional climate system, few in situ observations are available to examine the significance of these atmospheric effects. Several automatic weather stations (AWS) are available near Terra Nova Bay, but these stations are widely spaced and limited to within a few meters of the surface (Stearns et al. 1993). Satellite data coverage has improved dramatically in recent years, and while the temporal and spatial scales of this data are high, the vertical resolution of the data is limited, particularly near the surface. Aircraft flights flown during field campaigns can provide excellent data throughout the atmospheric layer, but can be expensive and logistically difficult. To date and to the authors' knowledge, there have only been two field campaigns consisting of instrumented aircraft over the Terra Nova Bay region (Parish and Bromwich 1989; Davis et al. 2008). The first field campaign took place during November 1987, and consisted of two successive flights through a

katabatic layer originating from Reeves Glacier, with the purpose being to study the boundary layer dynamics within a katabatic flow field. The second took place in several transects across the polar plateau, Ross Island, and Terra Nova Bay during November and December 2003, collecting measurements of temperature and pressure as part of a study aimed at understanding the processes that control trace chemicals across Antarctica (Davis et al. 2008).

In September 2009, the first unmanned aerial vehicles (UAV), which are more easily maneuverable and require less of a logistical component than larger, manned aircraft, were flown from the Pegasus White Ice runway near Ross Island to Terra Nova Bay, with the purpose being to examine the atmospheric mechanisms responsible for the formation and modification of dense shelf water within the area polynya. The UAVs measured wind speed and direction, temperature, humidity and pressure over Terra Nova Bay. As well, information about the surface state of the polynya, such as digital images and skin temperature measurements, were collected during the UAV flights at high vertical and temporal resolutions (Cassano et al. 2010). Prior to the September 2009 UAV field season, an examination of the local weather patterns of Terra Nova Bay from 1993-2008 was conducted using the only two readily available data sources in the region – AWS and satellite data. Prevailing wind speeds and directions, as well as changes with temperature, were studied from the AWS between 1993 and 2007, and synoptic and mesoscale cyclones from a manual satellite analysis for the months of August through October of 2006-2008. The purpose of this examination is to explore the characteristics of the local atmospheric forcing, namely downslope flow and cyclogenesis, and use these tendencies to place the data from the September 2009 field season in a broader context. The UAV observations of the atmosphere over Terra Nova Bay can be used to determine whether expected shelf water

formation from the polynya during September 2009 is stronger or weaker than the long-term average, based on our analysis.

Section 2 describes the data sources used for this analysis, including explanations regarding specific criteria for strong wind events and the use of Rita AWS. Section 3 describes the results from the 1993-2008 analysis years. Section 4 relates the 1993-2008 analysis to the meteorological pattern of September 2009. A summary is presented in Section 5.

### 2. Data Description

#### a. AWS observations

AWS data from Rita site, operated by the Italian Antarctic Research Programme (PNRA), was used as the primary dataset to study atmospheric changes in Terra Nova Bay. Rita AWS is located just west of Terra Nova Bay, downstream of both Reeves and Priestley Glaciers, at an elevation of 268 meters (Figure 2). The station measures temperature, wind speed and direction, relative humidity, and atmospheric pressure on an hourly basis. The temperature, pressure, and relative humidity data are instantaneous measurements reported at the top of the hour, while the wind measurements are the average of the data from the ten minutes prior to the start of the hour (Paolo Grigioni, pers. comm.). The Vaisala wind instruments are operational at a minimum temperature of -50°C, while the Vaisala temperature and humidity measurements are operational to -40°C. The tower stands 10 meters tall on bare rock, and is powered by two 48-Watt solar panels and six lead acid batteries. The data from the station is transmitted in real-time via the ARGOS data collection system (DCS), and has been in operation since January 1993. Data from

January 1993 through November 2007 have been quality-controlled by PNRA, and are used for this analysis. Data from September 2009 were quality-controlled by the authors for comparison with the 1993-2007 dataset to place the September 2009 observations into a broader perspective. Unfortunately, due to a loss of the radiation shield, the temperature sensor was not transmitting reasonable temperatures during September 2009, and was omitted for this period.

Analysis of the AWS data focused on wind and temperature observations. These observations were examined separately to determine the general flow pattern over Terra Nova Bay as well as the basic atmospheric properties of the region. Wind and temperature observations were then combined to determine the potential offshore interaction between the atmospheric and oceanic state. Changes in wind patterns, temperature, and wind-temperature fluctuations were examined using the Rita AWS data both over the entire 1993-2007 period and the September 2009 UAV time period, as well as by season, with the seasons defined as being spring (October – November), summer (December – January), autumn (February – March), and winter (April – September). The selection of these seasons were based on Seefeldt and Cassano (2008), where April was chosen as part of the winter period instead of autumn to closer match known Antarctic seasonal patterns.

Further analysis was conducted using the Rita AWS wind observations by identifying strong wind events over the study area. A strong wind event was identified as being one in which the wind speeds reported were in excess of 20 m s<sup>-1</sup> for at least a 10-hour duration. The selection of a 10-hour duration was based on Seefeldt and Cassano (2007) in which prevailing wind regimes over the Ross Ice Shelf were examined using AWS and model data to identify corresponding wind events. The selection of a minimum wind speed threshold of 20 m s<sup>-1</sup> was based on several factors, including manual analysis of the dataset. Several thresholds were tested,

including 15, 30, 40, and 50 m s<sup>-1</sup>, and it was found that using 20 m s<sup>-1</sup> captured a representative number of wind events that were maintained for longer than 10 hours. In addition to this analysis, Morales Maqueda et al. (2004) indicate the need for a wind speed threshold of at least 20 m s<sup>-1</sup> to exist in order to maintain ice-free polynyas around the Antarctic coast.

To account for fluctuations in wind speed due to wind gusts, winds were allowed to drop below 20 m s<sup>-1</sup> for no greater than 10 consecutive hours during an event. Below threshold durations of 5 and 10 hours were tested and compared to the criteria listed above. An example of the changes between the 10- and 5-hour duration analyses during a sample period (September 2002) can be seen in Figure 3. During this month, event 1 was eliminated during the 5-hour analysis, while event 2 was shortened by 25 hours. Event 1 was removed as there were more observations below the wind speed threshold than above, while event 2 was shortened due to the occurrence of a sharp peak in wind speed at the beginning of an event, followed by an immediate drop for more than five hours. A manual analysis of these events indicated that the first event should be removed, but the second should have the duration as in the 10-hour analysis. From this analysis and others throughout all months of September between 1993-2007 and 2009, it was concluded that a maximum below-threshold duration of ten hours would be used, while limitations would be imposed to eliminate those wind events where a greater period of time during the event was spent below the wind speed threshold. The criteria for the specifications of strong wind events are summarized in Table 1.

The AWS analysis was not extended through 2008 for two reasons. First, the PNRA quality-controlled dataset only included data through November 2007, and maintaining a uniform dataset throughout the analysis period was critical. Secondly, on 21 September 2008 the Rita AWS fell down, and data was unavailable until the following field season. Given that a



Figure 3. Comparison of strong wind events during September 2002 allowing 10 and 5 hours below the wind speed threshold. Gray boxes indicate the strong wind events for each threshold level.

significant portion of 2008, and in particular the month of highest interest to this study – September - was unavailable, 2008 was disregarded from the AWS analysis entirely. Additionally, there were several months during the 1993-2007 analysis period during which data was unavailable at Rita. These periods of data loss largely extended from winter through spring season, as issues with the AWS that occurred during winter were unable to be repaired until the summer field season. Periods of data loss occurred from July-October 1993, May-October 1996, and August-November 2001. No data was available for December 2007, as the data had not been quality-controlled during that time. This loss of data accounted for a 25% total loss of observations in spring, a 14% loss in summer, an 8% loss in autumn, and a 21% loss in winter.

Criteria	Value
Wind Speed Threshold	$20 \text{ m s}^{-1}$
Minimum Duration of Time at or Above	
Wind Speed Threshold	10 hours
Maximum Continuous Amount of Time	
Allowed Below Wind Speed Threshold	10 hours
Ratio of time above threshold to time below	≥1
threshold	

Table 1. Selection criteria for strong wind events.

#### b. Selection of Rita AWS

Rita site was chosen over other area AWS due to its representativeness of the meteorological conditions just upstream of the polynya. To confirm this, comparisons of the winds and potential temperatures at three other AWS upstream of Terra Nova Bay (Eneide, Sofia, and Zoraida) were made to the Rita AWS observations (Figures 4 and 5). To appropriately compare the four datasets, only times when observations were available at all four stations were used in the analysis. As Sofia AWS ceased transmissions in 2002, the dataset of the four stations was between 1993 and 2002. It was expected that Eneide site would have the lowest wind speeds of all four sites, based on its low elevation (approximately 91 meters) and being furthest from the terminus of the Reeves and Priestley Glaciers. Due to location, winds at Sofia site were expected to reflect influences from both Reeves and Priestley Glaciers, and Zoraida was expected to experience the strongest wind speeds due to its elevation (884 m) and location on Priestley Glacier. Figure 4 confirms this. Prevailing winds at Eneide are almost equally dominated from the west and north-northwest, and exhibit the weakest wind speeds overall and the least number of strong wind events (Figure 4). Winds at Sofia are mostly from the west, but also strong from the north, indicating influences from both Reeves and Priestley Glaciers (Figure 4). Winds at Zoraida reflect strong influences from Priestley Glacier in terms of directional flow, and also show the highest number of strong wind events (Figure 4).

It was expected that temperatures would be potentially coldest at Zoraida site, reflecting an unstable atmosphere consistent with katabatic drainage from higher elevations. However, Figure 5 shows temperatures to be potentially coldest at Sofia site. With a stable atmosphere present between Zoraida and Sofia, mean katabatic forcing cannot exist between these sites on Priestley Glacier without an additional external force, such as that from a background synoptic pressure gradient. Further analysis is needed to confirm this.



Figure 4. Wind roses from 1993-2002 for Eneide, Rita, Sofia, and Zoraida AWS.



Figure 5. Potential temperature histogram from 1993-2002 for Eneide, Rita, Sofia, and Zoraida AWS.

This analysis shows that Zoraida and Eneide sites are not the most ideal locations to characterize the off-continent air mass, while Sofia and Rita are. Zoraida's location inland and on Priestley Glacier is only an accurate representation of the air flowing down the glacier from the high polar plateau. Southwesterly winds at Eneide site (Figure 4) show an influence from local topography that will not necessarily be representative of the flow pattern that travels further east over the polynya. Additionally, Eneide's location is too far north of where the Priestley and Reeves Glaciers' wind jet extends to depict these air masses accurately. Sofia and Rita sites,
however, are ideally placed to characterize the air mass upstream of the polynya. Sofia's location on the Nansen Ice Sheet demonstrates flow from both Reeves and Priestley Glaciers, but the shortened observational record (which extends until 2002) is not ideal, and the lack of data in September 2009 makes this AWS ineffective to this study. Rita site, located on the windward side of the Northern Foothills and further south than Eneide, has a long data record, including wind observations during September 2009, and is the most reasonable AWS to use for this analysis.

Using all four sites for the analysis presented in this paper was not feasible, as the time period when all stations have coincident data is severely limited. The availability of coincident data was considered critical for this study, since missing data could easily bias the high wind event analysis. For the 1993-2002 time period, only two winters could be sampled due to a loss of transmissions from at least one of the four sites. Closer examination of Figure 6 shows that when using all four stations, upwards of 80% of the observations during the winter months were missing, whereas using only Rita AWS results in only a 30% loss in observations during those months. Given the above analysis between regional AWS, and the optimization of data during the winter months, Rita is the ideal AWS to use for this study.

#### c. Cyclone Climatology

The cyclone analysis was produced using the National Oceanic and Atmospheric Administration (NOAA) Local Area Coverage (LAC) 1-km resolution data provided by the Antarctic Meteorological Research Center (AMRC) at the University of Wisconsin. A time frame of August to October 2006 to 2008 was chosen as the analysis study period to acquire a reasonable sampling of cyclones around the time of year of the UAV flights in September 2009. For years 2006-2008, LAC data from the NOAA-18 satellite was used; for the September 2009 comparison dataset, NOAA-19 LAC data was used. Cyclones were identified using the infrared data over a domain area centered at 74°S 175°W (Figure 7), which covers the entire Ross Sea including Terra Nova Bay, from one satellite image per day, typically around 3 UTC. This dataset was expanded to include the entire Ross Sea for two main reasons: to include all systems that would have an impact on Terra Nova Bay, and to match previous climatological datasets that have included the entire Ross Sea region (Carrasco and Bromwich 1994; Gallée 1996; Carrasco et al. 2003).

The cyclones found from this data were manually identified by a thorough examination of each image (Figure 7). The analysis included any system showing cyclonic rotation over the domain, with information regarding position, time, size, and shape recorded. The size of the systems were identified by examining both the short and long dimensions of the system, with those smaller than 1000 km identified as mesocyclones, while those larger classified as synoptic systems (Heinemann 1990; Bromwich 1991; Carrasco et al. 2003). The shape of the system was divided into four categories: comma, merry-go-round, single-cyclonic band, and spiral-form, following previous studies of cyclones in the Antarctic (Carleton and Fitch 1993; Carrasco et al. 2003). Merry-go-round systems have several small cyclones attached to a much larger system, single-cyclonic band systems are those that are in the early stages of cyclone formation, and a spiral-form system has several bands that form a cyclone, as opposed to the comma-shaped systems which are a single mass of clouds. Information on the size and shape of cyclonic systems was included in this dataset to coincide with other cyclone climatologies conducted across the Antarctic (Carleton and Fitch 1993; Carrasco et al. 2003).



Figure 6. Percentage of strong wind events and percentage of missing data occurring during each month at a) Eneide, Rita, Sofia, and Zoraida AWS between 1993 and 2002, and b) Rita AWS between 1993 and 2007.



Figure 7. Example infrared satellite image from 2 October 2006 at 3 UTC used to conduct the cyclone analysis. The lines indicate the dimensions of a cyclone in the eastern Ross Sea, while the "L1" indicates the cyclone center. The box over Ross Island shows an anti-cyclonic vortex to the east of the island.

## 3. Climatology of Selected Variables in Terra Nova Bay

a. Wind speed and atmospheric temperature tendencies

Wind rose plots from Rita site show that the majority (55%) of winds are from the west, west-northwest, and northwest directions, indicating a dominant flow regime that is downslope (Figure 8). Eighty six percent of all wind speeds at Rita site are less than 20 m s<sup>-1</sup>; however, of

the wind speeds greater than 20 m s<sup>-1</sup>, 83% are from the west to northwesterly direction, indicating that most of the extreme wind speeds are from downslope flow. Between 1993 and 2007, a total of 418 strong wind speed events (as defined in Section 2) were identified, with the maximum winds found during the 1993-2007 time frame being 58 m s<sup>-1</sup>. Most strong wind events occurred in July (at 14%), with the least occurring during January (1%) (Figure 6b). The winter and autumn seasons showed the highest percentages of strong wind events at 68% and 17% respectively, while spring and summer had only 11% and 4% of the total wind events.

An examination of the temperatures at Rita site shows 39% of the observations are within the -15 to -25°C range, with a peak that exhibits a sharper decrease toward colder values and a gradual decline as temperatures get warmer (Figure 9). Analysis of the seasonal contributions to temperatures in the -15 to -25°C bin shows that the majority (73%) of temperatures in this bin occur during winter season, while 0% occur during summer. Spring and autumn contribute to 11% and 16%, respectively. The majority of observations colder than -25°C also occur during the winter season (97%), with 0% of the observations occurring during summer and the remainder evenly split between spring and autumn. Temperatures warmer than -15°C see contributions from all four seasons, with most temperatures occurring during summer (39%), and the least during winter (14%). Spring and autumn contribute 21% and 26%, respectively.

Histograms of temperature for various wind speed ranges are shown in Figure 10. Relatively few observations are found at the most extreme wind speeds, as the frequency of these events is much lower. Intermediate winds, such as the 10-20, 20-30, and 30-40 m s<sup>-1</sup> categories exhibit a Gaussian distribution, while the 0-10 m s<sup>-1</sup> category exhibits a bimodal distribution of temperature, with two peaks occurring near -20°C and 0°C. Further analysis comparing



#### Rita AWS Wind Rose for 1993-2007

Figure 8. Wind rose plot from Rita AWS for 1993-2007.



Figure 9. Temperature histogram for Rita AWS from 1993-2007.

## b. Synoptic and mesoscale cyclone characteristics

The identification of meso- and synoptic-scale cyclones from a manual cyclone analysis during the August-October 2006-2008 time period indicated several interesting patterns of cyclone locations across the Ross Sea (Figure 11). In particular, two prominent features are easily visible. The first is the high density of mesocyclones (cyclones with sizes less than 1000 km) located on the lee (northern) side of Ross Island, as indicated by Box 1. The large number of mesoscale cyclones in this area are likely cyclonic shear vortices that are produced by southerly flow around Ross Island. Previous studies suggest that some of these cyclones are spawned from cyclonic shear vortices that occur on the lee (north) side of Ross Island, which in

turn arise from the persistent low-level southerly winds that flow around the island topography (e.g. Monaghan et al 2005). Within the area outlined by Box 1 are three dense regions of mesoscale cyclones. The areas to the western and eastern sides of Ross Island (on the left and right-most portions of Box 1) are the shear vortices produced by the southerly wind flow, which are enhanced by the topography of the island. This topography alters the flow around the barrier such that cyclonic shear vortices on the western side, and anti-cyclonic vortices on the eastern side, no larger than 300 km, are produced (see example, Figure 7).

In the middle of Box 1 is a predominantly cyclone-void region, with the exception of a few larger (300-500 km) mesocyclones. The few number of cyclones produced directly north of Mt. Erebus, a nearly 4000 meter volcano that sits atop Ross Island, is due to the inability of flow that originates from the southern portion of the island to climb the barrier and flow over the obstacle instead of around (Seefeldt et al. 2003). The few cyclones that do bridge this barrier assist in the production of slightly larger cyclonic shear vortices than those that are produced when the flow traverses the westerly and easterly edges of the island. This cluster of cyclonic activity on the lee-side of Ross Island is in direct correlation with several previous studies that examined cyclonic flow in the southwestern corner of the Ross Sea (Bromwich 1991; Carrasco and Bromwich 1994; Carrasco et al. 2003). It is possible for the cyclones that are produced in the lee of Ross Island to propagate northward to later impact Terra Nova Bay. A second region of interest, outlined by Box 2 of Figure 11, depicts a noticeable lack of cyclones that not only form but track through this region. This lack of cyclonic activity will be discussed further in Section 4.



Figure 10. Temperature histograms by wind speed for Rita AWS from 1993-2007.



Figure 11. Map of cyclone locations from August to October 2006-2008. A high-density region of mesocyclones is located in the southwestern corner of the Ross Sea (Box 1). A low-density region of cyclonic activity is located in the center of the Ross Sea (Box 2).

To determine the influence of cyclonic activity on downslope winds, a comparison between strong wind events and cyclones in and around Terra Nova Bay was conducted for August-October 2006-2007 (the only time period during which both datasets coincided). This analysis showed that one hundred percent of the 15 strong wind events occurred in the presence of a cyclone located in the Ross Sea. This is a rather important result that will need to be further addressed as part of future analysis of the potential linkages between cyclones in the Ross Sea and strong winds in Terra Nova Bay.

In addition to the location of the cyclones, other characteristics, such as size and shape, were collected as part of the satellite analysis. Figure 12 depicts the size distribution of the various cyclones during the study period. Most cyclones (63%) found in the Ross Sea are smaller than 500 km, with 84% of all cyclones being mesocyclones. Only 16% of all cyclones found were classified as synoptic-scale systems. Examination of the shape of the cyclones identified a majority (64%) of the systems, either synoptic or mesoscale, were of a comma shape, with 25% spiral-form, 6% single-cyclonic band, and 3% merry-go-round (Figure 13). While the shape of a system will not change the system dynamics, it can be important in terms of understanding the regional circulation. For example, comma and merry-go-round systems tend to develop in regions of strong background flow, while spiral-form systems develop in synoptically quiet regions (Turner and Pendlebury 2000). Information regarding the shape and concurrent size of the systems can provide vital information about the individual systems and background synoptic environment that can impact the weather on a regional level, such as in Terra Nova Bay.

Knowledge of the typical atmospheric state over Terra Nova Bay is critical for understanding the UAV field measurements collected during September 2009. Changes in wind or temperature patterns can lead to changes in the opening of the polynya, which in turn can lead to changes in sea-air fluxes. Variations in cyclonic activity can lead to changes in the direction and magnitude of winds, and can advect warm or cold air into the region. Interactions between two air masses of different temperatures can generate mesoscale systems, which can also impact the size of the polynya. From this analysis, it can be seen that the primary air source that is advected over the polynya originates from the glacial valleys upstream of Terra Nova Bay, with a temperature typically observed at -20°C. In future analysis, observations of this air mass from the UAV collected during September 2009 will be compared to the findings from this study to put the UAV observations in a broader perspective. While the small number of UAV flights are not necessarily representative of the September 2009 monthly mean conditions, it is important to understand not only how the overall September 2009 conditions fit into the broader perspective, but how the conditions sampled by the UAV relate to the broader perspective as well. Inferences can then be made as to changes in the size of the polynya and in turn sea-air fluxes, which can impact changes in the shelf water of the polynya.



Figure 12. Histogram of cyclone size for August 2006 to October 2008. Bins range from 0-500 km, 500-1000 km, etc.



Figure 13. Number of cyclones of different shape for August 2006 to October 2008.

## 4. Comparison of AWS winds and cyclone activity for September 2009

An examination of wind fields and cyclone activity in Terra Nova Bay between September 1993-2007 and September 2009 shows that September 2009 experienced stronger winds and fewer cyclones, with more winds from the north-northwest, than the analysis months. As during the 1993-2007 analysis, the majority of winds originate from the west to northwesterly directions, indicating that downslope winds are the primary flow field over the polynya during September 2009. During the Septembers of 1993 to 2007, 60% of the winds were west to northwesterly, as compared with 75% of the winds during September 2009 (Figure 14). This shift in wind direction, particularly to the north-northwest during September 2009, indicates a displacement of surface cyclones in the Ross Sea and Ross Ice Shelf regions, accompanied by variations in an upper-level large scale feature, as described below.



Figure 14. Wind roses from observations at Rita AWS for all Septembers from 1993-2007 and September 2009.

During September 2009, 39% of winds were greater than 20 m s<sup>-1</sup>, as compared with 14% of wind observations during 1993-2007. As during the analysis years, 83% of those winds greater than 20 m s<sup>-1</sup> were from the west to northwesterly directions, indicating that the strongest winds correspond with downslope flow. Five strong wind speed events were recorded in September 2009, while the 1993-2007 Septembers experienced an average of 3.7 strong wind speed events per month (Figure 15). As during the analysis years, 100% of the strong wind events that occurred during September 2009 arose in the presence of a Ross Sea cyclone.

Most of the cyclone activity shifted toward the eastern Ross Sea during September 2009, while much of the activity during the previous Septembers (2006-2008) was located in the western to central Ross Sea (Figure 16). A cyclone-void region in the area between approximately 71-75°S and 165-175°W, as outlined in Box 2 of Figure 11 and Box 1 of Figure 16, shows that during September 2009 this area had 150% more cyclones (13 vs. 5) than during September 2006-2008, with four of the five cyclones observed within the box occurring in 2008. During 2008, nearly double the number of cyclones were observed throughout August-October, which was due to a strong upper-level trough at 500 mb located in the same area as the trough in 2009 (not shown).



Figure 15. The number of strong wind events during September between the years of 1994 and 2009. Years excluded were years with no data.



Figure 16. Cyclone locations for all Septembers from 2006-2009. A region void of cyclone activity in the Ross Sea in years 2006-2008, and with increased activity in 2009, is identified by Box 1.

A comparison of the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis 500 mb geopotential height composite means for September 1993-2007 to September 2009 shows an upper-level trough that was situated over the center of the Ross Ice Shelf in previous Septembers had deepened and shifted further east over western Marie Byrd Land during September 2009 (Figure 17). These upper-level troughs, which will tilt toward colder air, are indicative of surface cyclones to the north and northeast. Hence, typical Septembers show more cyclones in the center to just east of center portions of the Ross Sea, while during September 2009 the majority of surface cyclones would have been further east of the Ross Sea and north of Marie Byrd Land. This shift explains the lower number of cyclones found in Terra Nova Bay during September 2009.

A comparison of Figures 17a and 17b indicates that an enhanced west to east pressure gradient force, which will strengthen downslope surface winds in the vicinity of Terra Nova Bay, is present in 2009 compared to the multi-year mean. This enhanced pressure gradient is the result of both a deeper trough in the eastern Ross Sea as well as a stronger ridge over the East Antarctic plateau. An analysis of Figure 15 shows that the years of 1994, 2000, 2003, 2005, and 2009 had an above average number of strong wind events, while 1999, 2002, and 2007 were below average. 1995 and 1997 had an average number of strong wind events during the year. A closer examination of the NCEP/NCAR reanalysis 500 mb geopotential height fields for these years (not shown) indicates that during the average and above average years (with the exception of 1997), a ridge could be found on the high polar plateau. During the years when the number of strong wind events were below average, no ridge was found on the polar plateau, but rather a trough at 500 mb was found just west of the Ross Ice Shelf on the high polar plateau. Turner et al. (2009) suggests that a strong ridge pattern on the high polar plateau accompanied by a sea level low pressure anomaly promotes an ideal situation for generating strong wind events off the East Antarctic coast. The ridge during September 2009 was more intense than for the other strong wind event years, as was the trough over western Marie Byrd Land, implying strong downslope flow along the glacial valleys of the Transantarctic Mountains, including in Terra



b)

Figure 17. NCEP/NCAR Reanalysis monthly means of 500 mb geopotential height (m) for a) September 1993-2007 and b) September 2009.

Nova Bay. The shifts in the 500 mb geopotential height pattern described above would also be expected to influence surface low pressure centers. The eastward shift in cyclone locations during September 2009 is consistent with an eastward shift in the 500 mb trough during 2009 (Figure 16).

One potential explanation for the shift in the 500 mb geopotential height pattern is related to the El Niño/Southern Oscillation (ENSO) pattern of 2009, which experienced a transition from a cold ENSO (La Niña) to warm ENSO (El Niño) during the latter half of the year (Arndt et al. 2010). Correlations between ENSO events and Antarctic climate patterns are highly variable, but studies have shown that ENSO events can lead to changes in the Rossby wave pattern of the Southern Hemisphere (Turner 2004). Previous studies (Kidson 1999; Turner 2004; Fogt and Bromwich 2006) have shown that during warm ENSO events there typically exists a high pressure anomaly in the Amundsen Sea region, while in cold ENSO events a low pressure anomaly is present. These anomalies are particularly present during the September-October-November (SON) months (Fogt and Bromwich 2006). An examination of the difference between 500 mb heights in 2009 and the analysis years (1993-2007) shows the existence of a high pressure anomaly in this area (Figure 18a). The intensity and location of the anomaly in SON 2009 is similar to years in which an El Niño pattern was present, and is particularly comparable to the 2002 warm ENSO event (not shown), which corroborates findings from the 2009 State of the Climate Report (Arndt et al. 2010).

While the mean SON high pressure anomaly for 2009 coincides with previous studies that indicate a similar upper level pattern to other El Niño years, the September 2009 month does not represent the El Niño signal. Instead, a low pressure anomaly is present in the Amundsen Sea during this month, indicating that the Amundsen Sea region during September 2009 was still experiencing the effects of La Niña (Figure 18b). Later in 2009, as Figure 18a shows, a wave train of positive and negative height anomalies consistent with the El Niño pattern can be seen in the South Pacific and South Atlantic (Fogt and Barreira 2010).



Sep to Nov: 2009 to 2009 minus 1993 to 2007

Sep: 2009 to 2009 minus 1993 to 2007

# Figure 18. NCEP/NCAR reanalysis monthly means of 500 mb geopotential height (m) difference between a) SON 2009 and SON 1993-2007 and b) September 2009 and September 1993-2007.

The Southern Annular Mode (SAM) indices in the latter half of 2009 were highly variable. Generally speaking, a negative (positive) SAM index is correlated with higher (lower) pressures over the Antarctic continent (Turner 2004; Fogt and Bromwich 2006). The SAM index of September 2009 was -0.78 (Fogt and Barreira 2010), indicating higher pressures over the interior. Indeed, an examination of Figure 18 shows these higher pressures.

The eastward shift in the upper-level flow pattern and surface cyclone activity, a stronger than average ridge on the polar plateau and stronger than average low in the Ross Sea, and stronger than average winds at Rita site indicates that September 2009 was anomalous. An understanding of the September 2009 meteorological pattern, and its relationship to previous patterns, is imperative for understanding the data collected from the UAV flights in September 2009. As discussed in several studies, including Bromwich and Kurtz (1984) and Bromwich et al. (1993), strong wind events aid in opening Antarctic polynyas, both in Terra Nova Bay and elsewhere across the continental coast. In a year such as September 2009, when winds were stronger than previously observed, the ice-free polynya will occur more frequently and exist for longer periods of time. A combination of the strong winds, colder temperatures from the westerly winds originating from the high polar plateau, and open ocean water will increase ocean to atmosphere heat fluxes during this time. An understanding of the strength of the wind flow during this month, as well as its origin (i.e., downslope) and how it compares to previous years will provide a basis to determine how the anomalous year sampled by the UAV will impact the Terra Nova Bay climate system in September 2009.

#### 5. Summary

The climate of the Terra Nova Bay region of Antarctica was studied from 1993-2009 using AWS and satellite data to establish an understanding of the local flow regimes and atmospheric conditions of this region. This work was in support of an observational field campaign conducted during September 2009 that included the use of UAVs to collect information regarding the atmospheric structure of Terra Nova Bay. The analysis of weather patterns during the years prior to 2009 was used to place the AWS and satellite observations during September 2009 into a broader context.

Wind and temperature data from Rita AWS, located to the west of Terra Nova Bay on the Northern Foothills region of the Transantarctic Mountains, were analyzed during 1993-2007 and 2009 to determine dominant weather conditions. As several studies have shown that downslope flow is strongly driven by a pressure gradient force induced by sea level cyclonic activity, a manual satellite analysis was produced from August to October 2006-2008 to study the density of cyclones throughout Terra Nova Bay and the nearby Ross Sea. The wind, temperature, and cyclone datasets from 1993-2008 were then compared to AWS and satellite datasets collected during September 2009 to determine the similarities of the field season month to years prior.

Results indicate typical winds in Terra Nova Bay originate from glaciers just west of Terra Nova Bay. The majority of the winds are less than 20 m s<sup>-1</sup>, but of the winds greater than 20 m s<sup>-1</sup>, 83% are from directions consistent with downslope flow. An analysis of strong wind events found 418 events between 1993 and 2007, with 100% of these events during 2006-2007 associated with cyclonic activity in the Ross Sea. Winter experienced the greatest number of strong wind events (68%), with 14% of the total events occurring during July. Summer season exhibited the least number of strong wind events, with only 1% of events occurring during January.

Temperature histograms at Rita site from 1993-2007 show that 39% of temperatures fall in the ten-degree band between -15 and -25°C, with 45% of temperatures between -15 and 10°C, and only 16% of temperatures below -25°C. A satellite analysis over Ross Sea found a highdensity of mesocyclones occurring on the lee-side of Ross Island, including the southwestern corner of the Ross Sea as well as Terra Nova Bay. The majority of systems classified during this

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study period were found to have a size extent of less than 1000 km, with the remaining 16% identified as synoptic systems. Most of these cyclonic systems had a comma shape.

A comparison of the AWS and satellite data from the 1993-2007 Septembers and September 2009 showed an increased number of strong winds and an eastward shift of cyclonic activity during September 2009. NCEP/NCAR Reanalysis 500 mb geopotential heights show an eastward shift in an upper level trough in September 2009, as compared to Septembers 1993-2007, over the eastern Ross Sea. This eastward displacement was due to changes in the Southern Hemisphere Rossby wave pattern, which was consistent with previous documented impacts of ENSO on high southern latitude circulation (Turner 2004; Fogt and Bromwich 2006). The intensification of the 500 mb trough and a strong ridge center over the high polar plateau enhanced the downslope directed pressure gradient force and thus downslope flow in the region. The strong flow pattern, change in cyclone activity, and shift in synoptic flow patterns, indicates that September 2009 was an anomalous year. Data collected and analyzed in future work from the UAV flights during this month will provide exciting new insights into uncharacteristic flow patterns over the region.

#### 6. Acknowledgments

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## Chapter 3: A methodology for estimating sensible and latent heat fluxes from in situ aircraft measurements

ABSTRACT. In September 2009, several Aerosonde® unmanned aerial vehicles (UAVs) were flown from McMurdo Station to Terra Nova Bay, Antarctica, with the purpose of collecting three-dimensional measurements of the atmospheric boundary layer (ABL) overlying a polynya. Temperature, pressure, wind speed, and relative humidity measurements collected by the UAVs were used to calculate sensible and latent heat fluxes (SHF and LHF) during three flights. Fluxes were calculated over the depth of the ABL using a robust and innovative method in which only measurements of mean atmospheric state (no transfer coefficients) were used. The initial flux estimates assumed that the observations were Lagrangian. Subsequent fluxes were estimated that included modifications to incorporate adiabatic and non-Lagrangian processes as well as the mass below flight level.. The SHF ranged from 12 to 485 W m<sup>-2</sup>, while the LHF ranged from 56 The importance of properly measuring the variables used to calculate the to 152 W  $m^{-2}$ . adiabatic and non-Lagrangian processes is discussed. Uncertainty in the flux estimates is assessed both by varying the calculation methodology and by accounting for observational errors. The SHF proved to be most sensitive to the temperature measurements, while the LHF was most sensitive to relative humidity. All of the flux estimates are sensitive to the depth of the boundary layer over which the values are calculated. This manuscript highlights these sensitivities for future field campaigns to demonstrate the measurements most important for accurate flux estimates.

## 1. Introduction

The coastal landscape of Antarctica is dominated by fierce winds that originate over the continental interior, introducing cold, dry air to a relatively warmer and moister atmospheric boundary layer (ABL) at the outlet of glacial valleys. Coastal polynyas are formed when this strong off-continental flow pushes ice offshore, leaving an area of open water or thin sea ice adjacent to the coast. In Terra Nova Bay (TNB), located in the western Ross Sea (Figure 1), strong flow that can be greater than 40 m s<sup>-1</sup> (Bromwich 1989; Hauser et al. 2002; Knuth and Cassano 2011) produces a polynya that is a dominant feature throughout the winter months. As cold and dry air is advected over the polynya, the ensuing large ocean-atmosphere temperature and humidity differences can lead to large sensible and latent heat fluxes (SHF and LHF). This modification of the near surface air above the polynya in TNB has been shown to enhance mesocyclone development, sea ice production, and Antarctic Bottom Water (Budillon et al. 2003; Carrasco et al. 2003; Petrelli et al. 2008; Orsi and Wiederwohl 2009).

To understand the conditions within the ABL over the polynya and to quantify air-sea interactions in TNB from observational data, several unmanned aerial vehicles (UAV) were flown over TNB in September 2009 to collect information on the three-dimensional state of the atmosphere (Cassano et al. 2010; Knuth et al. 2013). Using the data collected during these flights, SHF and LHFs are calculated to quantify air-sea interactions in the region.

A methodology to calculate the heat fluxes is described, where only mean atmospheric state data from the UAV are utilized and a bulk transfer coefficient is not used. This method initially considers a Lagrangian approach to estimating the fluxes, with corrections to incorporate other processes considered. The purpose of this paper is to describe this method of calculating

heat fluxes while highlighting the sensitivity of the necessary corrections on estimating the fluxes. Understanding this sensitivity will be critical for other researchers flying scientific missions to ensure the data most sensitive to the flux estimations are observed well. As such, the uncertainty in the flux calculations will also be assessed from both a methodological as well as an instrumental perspective. Section 2 describes the UAV flights, Section 3 discusses the methodology used to calculate the fluxes, Section 4 describes the results and uncertainty, and Section 5 provides a summary.

#### 2. Aerosonde Flights

In 2009, sixteen Aerosonde® UAV missions were flown from McMurdo Station, Antarctica, with six local test missions, two failed missions, and eight missions to TNB (Cassano et al. 2010; Knuth et al. 2013). The UAV had a wingspan of 3 m, a weight of 15 kg, and a payload capacity of 2-5 kg. The aircraft had a range of over 1000 km, and flew at approximately 150-3000 m altitude during the flights. The flights were flown as low as possible to ensure the most accurate flux estimates, but could not be flown below 150 m due to concerns about icebergs as well as inaccuracies in the Global Positioning System (GPS) data. Data was telemetered back to the field team in McMurdo in real time by an on board data logger (Knuth et al. 2013).

The UAV carried several instruments on board, including temperature, pressure, and relative humidity sensors that recorded data with a 10 second temporal resolution (Knuth et al. 2013). Wind speed measurements were made using a combination of air speed calculated by the on-board pitot tube and ground speed observations from the GPS. Due to the lack of a magnetometer, periodic aircraft maneuvers were performed to find wind direction. These wind

finding maneuvers involved periodic sweeps across the incoming air stream where the maximum air speed measured by the pitot tube was found. Specific humidity was calculated from the relative humidity and pressure measurements.

On days when the flight mission was to estimate the turbulent fluxes over the polynya, the UAV flew north from McMurdo toward TNB along the coast. Once over the Drygalski Ice Tongue (DIT) the UAV would descend to ~150-240 m, and a roughly south-north coastline parallel flight leg across TNB was flown to find the location of maximum wind speed (the downslope wind jet) (Figure 19). Once this determination was made, the UAV was then flown downwind of the coast within this jet. Flying downwind from the coast within the jet achieved two goals: 1) to sample the area of TNB that would have the strongest air-sea fluxes; and 2) to document the downstream evolution of the ABL as the continental air mass was modified by the air-sea fluxes. Along this approximately west-east cross section, vertical profiles were collected approximately every 20 km across TNB (Figure 19). The UAV ascended and descended within the profiles by spiraling with a 1 km width; however, not all profiles contained useable data within each ascent or descent.

Of the sixteen UAV flights, three flights – 18 September, 23 September, and 25 September – were used to calculate heat fluxes as part of this study. Three profiles were collected on 23 and 25 September (Profiles 1-3), where Leg 1 uses data between Profiles 1 and 2, and Leg 2 between Profiles 2 and 3 (Figure 19). On 18 September, four profiles were collected, but only the furthest downstream pair – Profiles 3 and 4 (corresponding to Leg 3), are used in this study. Using the data from these flights, a methodology for calculating the fluxes was developed, and is described in the following section.



Figure 19. Flight paths on 18, 23, and 25 September 2009. Markers indicate profile locations along flight path. Profiles 1-4 are labeled as P1, P2, P3, and P4. Eneide AWS is labeled as "END".

## 3. Flux Calculation Methodology and Corrections

The methodology described below uses atmospheric data (temperature, pressure, wind speed, and moisture) to estimate changes in the energy present in the atmosphere over the UAV flight path. The general approach, described in 3a, uses methods similar to those of Kottmeier and Engelbart (1992) and Serreze et al. 1992, as well as those in the ocean community to estimate carbon dioxide flux through inverse modeling (Tarantola 1987; Enting et al. 1993; Bosquet et al. 1999). However, the methodology outlined in this work provides a much more extensive analysis of the non-Lagrangian correction methods (outlined in 3c), making this work unique.

#### a. SHF formulation

Using a Lagrangian approach, the heat fluxes are estimated by examining changes in the state of an air parcel along a UAV flight path. An estimation of the SHF can be found from the thermodynamic energy equation (TEE) written in finite difference form:

$$Q_s = c_p \frac{\Delta T}{\Delta t} - \frac{1}{\rho} \frac{\Delta p}{\Delta t} \tag{1}$$

where  $c_p$  is the specific heat of dry air at constant pressure (1004 J kg<sup>-1</sup> K<sup>-1</sup>), *T* is the air temperature (K), *t* is time it takes the air parcel to travel between concurrent profiles,  $\rho$  is the density of the air (kg m<sup>-3</sup>), and *p* is the pressure (Pa). The density is calculated using virtual temperature and pressure from the UAV data.  $c_p \frac{\Delta T}{\Delta t}$  is proportional to the time rate of change of

temperature and  $\frac{1}{\rho} \frac{\Delta p}{\Delta t}$  is the adiabatic term, which are both estimated from the UAV observations. It is assumed that changes in  $Q_s$ , the diabatic heating term (here in units of J kg<sup>-1</sup> s<sup>-1</sup>), are only due to surface fluxes from the polynya into the air parcel. All other diabatic processes, such as entrainment, subsidence, and radiation, are assumed to be negligible as part of this study. While neglecting these processes may impact the results, the necessary observations to quantify these terms are not available for these flights.

Equation 1 is modified by multiplying by  $\rho dz$ , the mass per unit area of the column of interest, to put the equation in terms of W m<sup>-2</sup> such that:

$$\rho \Delta z Q_s = \frac{c_p \rho \Delta T \Delta z - \Delta p \Delta z}{\Delta t}$$
(2)

where z is the UAV height (m) at some point in the ABL profile. The term on the LHS is the SHF in units of W m<sup>-2</sup>, while the RHS represents the change in heat content (HC) adjusted by adiabatic processes over the ABL over time. For these calculations the air parcel is assumed to extend over the depth of the ABL, since this is the portion of the atmosphere directly influenced by the surface. The depth of the ABL is estimated from the UAV observed potential temperature profiles along the flight path.

Considering the change in HC throughout the depth of the ABL at a single point in time and (for the moment) neglecting adiabatic processes, the RHS becomes:

$$HC = \sum c_p \rho T \Delta z \tag{3}$$

where the HC is being summed over the depth of the ABL (UAV flight level to the ABL top). This term represents the amount of energy present over the entire depth of the ABL within each profile measured along the UAV flight path. The SHF (in W m<sup>-2</sup>) is then calculated by examining the change in HC over the amount of time,  $\delta t$ , it takes the air parcel to travel between the upstream and downstream profiles, or

$$SHF = \frac{HC_{down} - HC_{up}}{\delta t}$$
(4)

Because the SHF in Equation 4 is valid between two concurrent profiles, there is one value estimated for each flight leg. For example, if a flight leg has three profiles, there will be two SHF values – one valid between Profiles 1 and 2, and one valid between Profiles 2 and 3.

Equations 3 and 4 are critical to this study, as these indicate how changes in the atmospheric state will alter the HC of the air parcel, which will in turn be used to estimate the SHF. The derivation of these equations assumes the observations are exactly Lagrangian and neglect adiabatic changes in temperature between pairs of profiles. We will use this simplest estimate of the SHF as a baseline for comparing more realistic flux estimates that include processes such as adiabatic temperature changes and the non-Lagrangian nature of the observations. To account for this, the HC estimated in Equation 3 will be altered for each process that influences the SHF, and corrections to the purely uncorrected HC (given in Equation 3) will be made to estimate a new flux. Each of these corrections (described below) will be compared to the uncorrected SHF to estimate the relative impact of each correction on the SHF.

#### b. Adiabatic Processes

As shown in Equation 1, as the air parcel travels from the upstream to the downstream profile, changes in the temperature of the column occur due to both diabatic and adiabatic processes. When calculating the SHF, only diabatic changes in temperature should be considered and adiabatic changes removed. To account for and remove changes in the temperature due to adiabatic processes, Poisson's equation, derived from the TEE, can be used to estimate the adiabatic temperature of the air parcel:

$$T_{adb} = T_{1obs} \left(\frac{p_2}{p_1}\right)^{R/c_p}$$
(5)

where  $T_{adb}$  is the adiabatic temperature the upstream air column would have if it moved downstream at the new pressure  $p_2$ .  $T_{1obs}$  is the UAV observed mean temperature of the upstream air column,  $p_1$  and  $p_2$  are the average pressures of the upstream and downstream columns, and R is the gas constant for dry air (287 J kg<sup>-1</sup> K<sup>-1</sup>). Then, to estimate the diabatic temperature of the downstream air column, we use:

$$T_{2dia} = T_{2obs} - (T_{adb} - T_{1obs})$$
(6)

where  $T_{2obs}$  is the UAV observed mean temperature of the downstream air column, and the term in parentheses on the RHS is the adiabatic temperature change experienced as the air parcel moves from the upstream to the downstream profile location. Finally, to estimate a SHF that excludes adiabatic processes, the temperature in Equation 6 is used to estimate the downstream HC, so that Equation 3 now becomes

$$HC_{down} = \sum c_p \rho T_{2dia} \Delta z \tag{7}$$

and Equation 4 is used to estimate the SHF.

## c. Non-Lagrangian Processes

The above formulations describe fluxes that use a Lagrangian approach, assuming the UAV measures the same air parcel at the location of the downstream profile as it did at the upstream. In reality, the UAV does not fly at the same speed as the air parcel, and this non-Lagrangian nature of the observations needs to be accounted for in the flux estimations within the TEE. Non-Lagrangian processes will be used to correct the temperature and pressure variables to consider the conditions that should have been measured if the UAV had followed the air parcel. This correction will consider two perspectives: time and space, with the space perspective separated into two distinct methodologies. Both perspectives are equally viable, but differ in the observations used to implement them.

#### *i. Time Perspective*

Figure 20a shows a schematic diagram of the time perspective used to account for the non-Lagrangian nature of the UAV observations. In this schematic, the u subscript represents the

UAV, and the *p* subscript represents the parcel. Consider two profile locations, P1 (at position  $x_1$ ) and P2 (at position  $x_2$ ), separated by a distance  $\Delta x = x_2 - x_1$ . At time  $t_1$ , both the UAV and air parcel are located at P1 ( $x_{u,1} = x_{p,1}$ ,  $t_{u,1} = t_{p,1}$ ). As the parcel and UAV move toward P2, they travel at different speeds, reaching the location of P2 (at distance  $x_2$ ) at different times ( $x_{u,2} = x_{p,2}$ ,  $t_{u,2} \neq t_{p,2}$ ). This time mismatch needs to be corrected with (in terms of temperature):

$$T_{corr} = -\frac{\partial T}{\partial t} \delta t \tag{8}$$

where  $T_{corr}$  represents the correction to the Lagrangian temperature (or the non-Lagrangian correction),  $\frac{\partial T}{\partial t}$  is the Eulerian time rate of change of temperature, and  $\delta t$  is the difference in time  $(t_{u,2}-t_{p,2})$  between when the UAV and air parcel reach the location of P2. It is important to note that not only is temperature corrected in this way, but also pressure.

The Eulerian time rate of change term  $(\frac{\partial T}{\partial t})$  is estimated using data from Eneide AWS (Figure 19) because there is no stationary data over the UAV flight path that allows for an estimation of the Eulerian time rate of change term. In an analysis conducted by Knuth and Cassano (2011), Rita AWS was shown to be the most representative weather station for conditions over TNB. However, due to instrument failures and a loss of data during September 2009 at that site, Eneide was chosen for this study. Unfortunately, Eneide (as well as all other AWS in the region) is a poor choice to use for estimating the  $\frac{\partial T}{\partial t}$  term. Eneide is approximately 35-70 km from the profiles being collected along the polynya, and in a completely different topographical regime. As well, the time resolution of the AWS data is hourly, while it takes approximately 15-20

minutes for the UAV to fly between profiles. The lack of appropriate data to estimate the time correction for the 2009 UAV flights raises concerns about the reliability of time perspective corrections for this field campaign, and as such, a discussion of the time corrected fluxes is not included in this work. In a subsequent set of UAV flights over TNB in September 2012, Eulerian measurements of the atmosphere were collected by the UAV over the polynya that make an appropriate estimate of the time-only corrections possible.

### ii. Space Perspective

There are two space perspective scenarios to consider when correcting for the non-Lagrangian nature of the UAV observations. The first considers a scenario where the air parcel and UAV start at P1 at the same time but at some later time are not collocated. The second considers a scenario where the air parcel and UAV are at P2 at the same time and are not collocated at the initial time. Each method should provide similar answers, and both are described below.

Figure 20b is a schematic illustrating the first space perspective scenario. Again, P1 and P2 are separated by a distance  $\Delta x$ , and both the UAV and air parcel are located at P1  $(x_{u,1} = x_{p,1})$  at the initial time  $(t_{u,1} = t_{p,1})$ . As the parcel and UAV move toward P2, they travel at different speeds. In the example illustrated in Figure 20b, the UAV travels slower than the air parcel. Therefore, when  $t_{p2}=t_{u2}$  and the UAV is located at P2, the air parcel is located at some point downstream, and  $x_{u,2} \neq x_{p,2}$ . This space mismatch needs to be corrected with (in terms of temperature):



a)

Space Perspective – P1 Equal



b)




c)

Figure 20. Schematic of the time (a) and space (b and c) perspectives of the non-Lagrangian approach to the heat flux calculations. The circle represents the air parcel, and the plane outline represents the UAV. P1 and P2 represent Profile 1 and Profile 2, x indicates distance, t indicates time, u represents the UAV, and p represents the parcel. The "1" and "2" subscripts indicates two different points in space or time. The equal P1 space perspective is shown in panel b and the equal P2 space perspective is shown in panel c.

$$T_{corr} = -\frac{\partial T}{\partial x} \delta x \tag{9}$$

where  $\frac{\partial T}{\partial x}$  is the temperature gradient found from the difference in the mean column temperatures at the start and end of the flight path, and  $\delta x$  is the difference in space  $(x_{p,2}-x_{u,2})$  between the air parcel location and the UAV location (P2).

Figure 20c is a schematic illustrating the second space perspective scenario where the UAV and air parcel are located at P2 ( $x_{u,2} = x_{p,2}$ ) at the same time ( $t_{u,2} = t_{p,2}$ ). Because the air parcel and UAV travel at different speeds, they must have originated at different locations

upstream. In the example illustrated in Figure 20c, the air parcel, located at P1 at  $t_{u,1}$ , travels faster than the UAV. At this same time, the UAV is located some distance downstream of P1 so that  $x_{u,1} \neq x_{p,1}$ . This space mismatch, much like the equal P1 scenario described in Figure 20b, is also corrected with Equation (9).

A SHF that accounts for the non-Lagrangian processes illustrated in Figure 20b can be estimated by altering the downstream HC in Equation 3 such that

$$HC_{down} = \sum c_p \rho T_{2LG} \Delta z \tag{10}$$

where  $T_{2LG}$  is the downstream air column temperature that only incorporates Lagrangian processes. This temperature is found by adjusting the observed downstream temperature by the non-Lagrangian temperature given by Equation 9. Equation 4 is then applied to estimate the SHF. Alternately a correction for the upstream HC can be applied based on the case illustrated Figure 20c.

In the space correction detailed above, the estimated temperature gradient based on the UAV measurements is considered instantaneous over the amount of time it takes the air parcel to fly to the downstream profile. An adjustment to the temperature gradient that considers changes over time should also be considered. However, an appropriate correction would need to consider data from a local AWS site such as Eneide, and this data has already been deemed inappropriate for the 2009 flights. Therefore, for the final SHF (and LHF) calculations, a correction to the estimated  $\frac{\partial T}{\partial r}$  for changes due to time along the flight path is not considered.

#### d. Changes in the Atmosphere Below Flight Level

The flux estimates described above neglect changes in the heat content in the atmosphere below flight level, which likely underestimates the fluxes described previously. To account for this, we estimate the change in heat content below the flight level by modifying Equation 3 to give:

$$\Delta HC = \Sigma c_p \rho \Delta T \Delta z \tag{11}$$

where the summation is evaluated over the depth of the layer between the lowest UAV flight level and the surface and  $\Delta T$  is the change in temperature between pairs of profiles and includes the non-Lagrangian and adiabatic corrections described previously. Since observations of the temperature below flight level are not available we assume that  $\Delta T$  below the UAV flight level is the same as that found between the flight level and ABL top. Given that the ABL is usually well mixed this is a reasonable assumption. The additional heat content calculated with Equation 11 is added to the previously estimated heat content value from flight level to the top of the ABL and a new SHF is calculated. This SHF is the final, fully corrected flux (hereafter termed the "final flux").

The above discussion highlights several adjustments to the uncorrected flux necessary for providing the most accurate estimate of the SHF. In this study, the following fluxes will be discussed: uncorrected flux, adiabatic-only (described in Section 3b), space-only (Section 3cii), space-adiabatic (for both described scenarios), and the final flux estimate. The final fluxes adjust

temperatures to account for both adiabatic and non-Lagrangian processes, as well as the heat content below flight level.

#### e. Latent Heat Fluxes

Calculation of the LHF will follow a similar formulation to the SHF. The water vapor conservation equation (in units of  $J \text{ kg}^{-1} \text{ s}^{-1}$ ) in finite difference form is:

$$Q_l = L_v \frac{\Delta q}{\Delta t} \tag{12}$$

where  $L_v$  is the latent heat of vaporization at 0°C (2.5x10<sup>6</sup> J kg<sup>-1</sup>) and q is the specific humidity. Equation 12 is converted to units of W m<sup>-2</sup> by multiplying both sides by the mass per unit area of the column ( $\rho \Delta z$ ). The diabatic moisture term with respect to latent heating  $Q_l$  incorporates surface fluxes, entrainment, and any latent heat release due to changes in phase, but we will assume that all changes to  $Q_l$  are due to surface fluxes.

Following on the methods of calculating the SHF, it is also useful to define the latent heat content (LHC) of the air parcel in each profile within the ABL to be:

$$LHC = \sum \rho \, L_v q \Delta z \tag{13}$$

The LHF is then calculated by considering the difference in LHC measured within each profile along the UAV flight path over the amount of time  $\delta t$  it takes the air parcel to travel between the upstream and downstream profiles,

$$LHF = \frac{LHC_{down} - LHC_{up}}{\delta t}$$
(14)

Equations (12-14) yield the uncorrected LHF for this study.

In order to make the most accurate estimate of the LHF non-Lagrangian corrections similar to the SHF must be applied. In September 2009, there was no relative humidity data from the AWS, and thus a time-only correction cannot be made. Following on Section 3cii, the space-only correction to the specific humidity data is:

$$q_{corr} = -\frac{\partial q}{\partial x} \delta x \tag{15}$$

where the terms in the equation are as listed above. Equation 13 is then modified to become

$$LHC = \int \rho L_{\nu} q_{2LG} \Delta z \tag{16}$$

and Equation 14 is used to estimate the LHF that incorporates non-Lagrangian processes. The correction to account for the LHC below flight level is implemented in a manner similar to what was described in Section 3d.

# 4. Results

## a. *ABL Selection*

Because the HC and LHC, and thus estimated fluxes, depend greatly on the depth of the ABL, the top of the ABL was determined to be the point at which the upstream and downstream potential temperature profiles are no longer clearly distinguishable from each other. This criterion was used since we expect that the surface fluxes experienced by an air parcel will modify the atmospheric state in the lowest portion of the profile, while conditions above the influence of the surface fluxes should remain relatively constant between pairs of profiles. However, some flight days displayed clear features, such as inversions, capped mixed layers, or changes in mixing ratio, which indicate the presence of a convective ABL (CBL), and in these instances the ABL height is chosen accordingly.

On 18 September (Figure 21), the top of the ABL was defined at 457 m. Below this point, the lapse rate in the downstream profile (Profile 4) is decreasing at the dry adiabatic lapse rate, whereas above 457 m, the lapse rate shifts to a positive value. The specific humidity profiles also support an ABL depth of 457 m as a shift in the profiles from decreasing specific humidity to a more well-mixed value is apparent, particularly in Profile 4. Given the complexity of the ABL profiles, alternative ABL heights (and impacts on flux estimates) are explored in Section 4d.

On Leg 1 on 23 September (Figure 22), an ABL height of 720 m was determined. Above this point, the upstream and downstream profiles are similar, while below 720 m the downstream profile (Profile 2) is clearly warmer throughout the layer than the upstream profile (Profile 1). In Profile 2, the layer between 420 and 720 m exhibits sharp spikes in the potential temperature, where the colder temperatures show conditions similar to Profile 1, and the warmer temperatures similar to Profile 2. We suspect that these spikes are not noise but an example of the UAV sampling the convective plumes of the ABL alongside the conditions in the profile that are not yet impacted by convection. This indicates the top of the ABL on this day is not at one level, but varies vertically over this 300 m distance. The specific humidity profiles also show a clearly moister ABL in the downstream profile below 720 m with little difference above this height.



Figure 21. The upstream (blue) and downstream (red) potential temperature (top), temperature (bottom left), and specific humidity (bottom right) profiles for Leg 3 on 18 September. The black horizontal line represents the best estimate of the ABL height.



Figure 22. The upstream (blue) and downstream (red) potential temperature (top), temperature (bottom left), and specific humidity (bottom right) profiles for Leg 1 on 23 September. The black horizontal line represents the best estimate of the ABL height.



Figure 23. The upstream (blue) and downstream (red) potential temperature (top), temperature (bottom left), and specific humidity (bottom right) profiles for Leg 2 on 23 September. The black horizontal line represents the best estimate of the ABL height.

The height of the ABL for Leg 2 on 23 September (Figure 23) was determined to be at 540 m. This flight leg exhibited most clearly the characteristics of a CBL, with a well-mixed layer present from flight level to 540 m in the downstream profile (Profile 3). This layer is topped by a capping inversion at 540 m that extends to 600 m. The specific humidity profiles are also well mixed in this layer with a rapid decrease in specific humidity above 540 m, consistent with the ABL top being located at this height.

On 25 September, an ABL depth of 240 m was determined for Leg 1 (Figure 24), with Profile 2 being warmer than Profile 1. Above this level, the two profiles exhibit similar characteristics to approximately 525 m, above which another air mass is likely impacting conditions within the profiles. The specific humidity profiles also show clearly distinct values below 240 m, although the values don't appear well-mixed until approximately 280 m.

For Leg 2 on 25 September, the two profiles exhibit similar characteristics throughout the entire depth of the UAV profile (Figure 25). However, a nearly well-mixed layer topped by a capping inversion in Profile 3 shows that the ABL top is likely at 250 m on this day. The specific humidity profiles support this contention. The downstream profile shows nearly constant specific humidity up to 250 m with a rapid drop above this height, consistent with an ABL depth of 250 m.

#### b. Sensible Heat Fluxes

From the methods described in Section 3, five variations of SHF have been calculated – the uncorrected, adiabatic-only, space-only, space-adiabatic, and final fluxes for the P1 and P2 scenarios. This section will briefly examine the uncorrected fluxes, then describe each



Figure 24. The upstream (blue) and downstream (red) potential temperature (top), temperature (bottom left), and specific humidity (bottom right) profiles for Leg 1 on 25 September. The black horizontal line represents the best estimate of the ABL height.



Figure 25. The upstream (blue) and downstream (red) potential temperature (top), temperature (bottom left), and specific humidity (bottom right) profiles for Leg 2 on 25 September. The black horizontal line represents the best estimate of the ABL height.

correction made to understand how the addition of other processes change the original, uncorrected flux estimate.

The estimates of the SHF are influenced by the temperature difference between the two profiles, the depth of the ABL, and the amount of time it takes the air parcel to travel between profiles (which is influenced by the wind speed) (Table 2). A larger temperature difference indicates a larger amount of energy being added to the air parcel as it moves downstream, which will lead to a larger SHF. A deeper ABL will also yield a larger flux due to a greater amount of mass being heated in the air column. An increased wind speed will yield a faster air parcel travel time that will also increase the flux.

Examining the uncorrected fluxes for each flight day (Table 2) shows a range of SHF between -1 and 331 W m<sup>-2</sup>, with the largest flux occurring in Leg 1 on 23 September and the smallest on Leg 2 of 25 September (Table 2). Despite Leg 1 on 23 September having a higher SHF than all other flight days, Leg 1 on 25 September exhibits the strongest wind speed and largest difference in temperature between the upstream and downstream profiles (Table 2). This seeming disparity occurs due to the depth of the ABL over which the HC is calculated – the ABL depth on the  $23^{rd}$  is over 475 m greater than on the  $25^{th}$  (Table 2). As such, the difference in HC per unit meter between the two concurrent profiles is over  $6.6x10^5$  J m<sup>-2</sup> greater on the  $23^{rd}$  than the  $25^{th}$ , leading to a larger SHF being estimated on the 23rd.

The adiabatic-only correction will decrease (increase) the flux due to an increase (decrease) in pressure along the flight path. An increase in pressure downwind will lead to a heating of the air parcel due to adiabatic effects. Because this adiabatic heating needs to be removed to estimate an accurate diabatic SHF, this increase in pressure will act to decrease the flux. The opposite is true for a decrease in pressure downwind. In Table 2,  $\Delta P$  describes the

	18 Sept.	23 Sept.	23 Sept.	25 Sept.	25 Sept.		
	Leg 3	Leg 1	Leg 2	Leg 1	Leg 2		
$WS (m s^{-1})$	19.6	18.2	16.4	22.2	18.7		
ABL depth	220	564	383	76	86		
above flight							
level (m)							
ABL depth	237	157	157	160	164		
below flight							
level (m)							
Total ABL	457	721	540	236	250		
depth (m)							
	I	Uncorre	cted				
$\Delta T(\mathbf{K})$	0.4	0.7	0.4	1.1	0		
SHF (W $m^{-2}$ )	49	331	121	75	-1		
Adiabatic-only correction							
$\Delta T(\mathbf{K})$	0.3	0.6	0.4	1.1	0.1		
$\Delta P$ (Pa)	45	90	3	101	-146		
SHF (W $m^{-2}$ )	44	297	120	70	25		
$\Delta$ SHF (W m <sup>-2</sup> )	-5 (-10%)	-34 (-10%)	-1 (0%)	-5 (-7%)	26 (26%)		
	Space-	only correc	<u>tion – P1 ec</u>	lual			
$\Delta T(\mathbf{K})$	0.9	0.8	0.4	1.9	0		
SHF (W m <sup>-2</sup> )	123	412	129	128	0		
<b>⊿SHF (W m<sup>-2</sup>)</b>	74 (151%)	81 (24%)	8 (7%)	53 (71%)	-1 (1%)		
Space-only correction – P2 equal							
SHF (W $m^{-2}$ )	124	414	129	128	0		
$\Delta$ SHF (W m <sup>-2</sup> )	75 (153%)	83 (25%)	8 (7%)	53 (71%)	-1 (1%)		
Space-adiabatic correction – P1 equal							
$\Delta T(\mathbf{K})$	0.8	0.7	0.4	1.8	0		
SHF (W m <sup>-2</sup> )	111	371	128	120	11		
<b>⊿SHF (W m<sup>-2</sup>)</b>	62 (126%)	40 (12%)	7 (6%)	45 (60%)	12 (12%)		
Final correction – P1 equal							
SHF (W $m^{-2}$ )	232	472	185	385	12		
$\Delta$ SHF (W m <sup>-2</sup> )	183	141 (43%)	64 (53%)	310 (413%)	13 (13%)		
	(373%)						

Table 2. Values for average wind speed over the ABL (WS), various ABL depths, temperature difference between pairs of profiles ( $\Delta T$ ), estimated sensible heat flux (SHF), pressure difference between pairs of profiles ( $\Delta P$ ), change in sensible heat flux from uncorrected value ( $\Delta SHF$ ), and the correction to the observed temperature ( $T_{corr}$ ) for flight legs on 18, 23, and 25 September 2009. Percent difference comparing the correction to the uncorrected flux is given in parentheses.

difference in pressure between the downstream and upstream profiles ( $\Delta P = P_{down} - P_{up}$ ). Table 2 shows that on most flight legs, pressures increased downwind (positive  $\Delta P$ ), and the adiabatic-only correction was negative, leading to a decrease in the uncorrected flux. On Leg 2 on 25 September, however, the pressure decreased, leading to a positive adiabatic-only correction.

The adiabatic-only corrections for all five flight legs ranged from -34 to 26 W m<sup>-2</sup>. Considering only the relative magnitude of the corrections, the largest magnitude correction was on Leg 1 on 23 September (34 W m<sup>-2</sup>) (Table 2), and the smallest (1 W m<sup>-2</sup>) on Leg 2 on 23 September (Table 2). The adiabatic correction had the largest relative impact on Leg 2 on 25 September, altering the uncorrected fluxes by 26%. The pressure change was also largest on this flight leg. The adiabatic correction had the smallest relative difference on Leg 2 on 23 September, where the adiabatic correction changed the uncorrected flux 0% (Table 2).

The space-only correction depends on two main factors: the temperature gradient along the flight leg  $(\partial T/\partial x)$ , and the space difference  $\delta x$  between the UAV and air parcel locations (see Section 3cii). A positive temperature gradient indicates an increasing temperature with increasing distance downwind. A positive  $\delta x$  indicates a situation where the UAV is further downstream than the air parcel – in other words, the UAV is flying faster than the air parcel. It is the product of these terms, as given in Equation 9, that controls whether the space-only correction will be positive or negative, and thus act to increase or decrease the uncorrected flux.

The space-only corrections for the P1 equal scenario ranged from -1 to 81 W m<sup>-2</sup> (Table 2), with the correction term on Leg 1 on 23 September being the largest, and Leg 2 on 25 September the smallest. The space-only correction had the largest impact on the 18<sup>th</sup>, however, adjusting the uncorrected fluxes by 151%. On Leg 1 on 25 September, the relative impact of the space-correction altered the uncorrected fluxes by half that of the 18<sup>th</sup>, and the other three days

had a relatively small adjustment. For all flight days, the  $T_{corr}$  term is positive, indicating an increase in the uncorrected flux. The space corrections for the P2 equal scenario are nearly identical to the P1 equal scenario, with values ranging from -1 to 83 Wm<sup>-2</sup>, and the relative impacts ranging from 1% to 153%. This indicates that either of the space correction methods is appropriate.

Perhaps the most important result from the space-only corrections described above is the similarity of both the P1 and P2 space-only fluxes, showing the robustness of this methodology. Each perspective is a different approach to an equally scientifically viable method, which should, and does, provide similar answers. For the P1 and P2 equal scenarios, the space-only fluxes vary by at most 2 W m<sup>-2</sup>, which is well within the uncertainty of the fluxes due to observational uncertainty (see below). Because of this, the remaining discussion in the text assessing individual corrections (such as the space-adiabatic and incorporating the HC and LHC below flight level), will only include a discussion of the P1 estimates. The final range of fluxes given at the end of this study, however, will include both the P1 and P2 estimates.

The space-adiabatic correction incorporates both the space-only and adiabatic corrections and uses the terms that impact each individual correction to adjust the uncorrected flux. The relative strengths of each correction term will determine how much the uncorrected flux is adjusted. Over all flight legs but Leg 2 on 25 September, the space-only correction is larger than the adiabatic correction, indicating that the non-Lagrangian term is more important than the adiabatic term on these flight legs for correcting the fluxes. The space-adiabatic correction terms for both the P1 and P2 equal scenarios are nearly identical, with the P1 equal scenario ranging from 7 to 62 W m<sup>-2</sup> and the P2 equal scenario ranging from 7 to 63 W m<sup>-2</sup>. For both scenarios, the largest correction term is on 18 September and the smallest on Leg 2 on 23 September. The largest and smallest impacts on the uncorrected flux occurred on these days at 126-129% and 6%, respectively.

The final corrected fluxes, which include the space-only, adiabatic, and heat content below flight level corrections, range from 12 to 472 W m<sup>-2</sup> (Table 2). Incorporating the ABL HC below the flight level has a large impact on the fluxes for most of the flight legs. The largest impacts are on Leg 1 on the 25<sup>th</sup> and the 18<sup>th</sup>, where this correction changes the uncorrected flux by 413% and 373%, respectively. This is a ~350% and ~250% change from the space-adiabatic correction (Table 2). These large corrections from incorporating the HC below flight level are because the depth of the ABL below the flight level (160 and 237 m) is a large percentage of the total ABL depth (236 and 457 m) (Table 2). The ABL depth is also small on Leg 2 on the 25<sup>th</sup> (250 m), but the changes in the atmospheric variables between the two profiles are comparatively smaller, and so the addition of the HC from below flight level does not have as large of an impact.

#### c. Latent Heat Fluxes

Similar to the SHF, the estimates of the LHF are influenced by the specific humidity difference between the two profiles, the depth of the ABL, and the amount of time it takes the air parcel to travel between profiles (which is influenced by the wind speed) (Table 3). A larger specific humidity difference indicates a larger amount of moisture being added to the air parcel as it moves downstream, which will lead to a larger increase in the amount of LHC in the column and thus larger LHF. The uncorrected LHF estimates range from 11 to 98 W m<sup>-2</sup> for each of the five flight legs, with the largest flux occurring in Leg 2 on 25 September, and the smallest on

Leg 1 on 25 September (Table 3). The largest difference in specific humidity between concurrent profiles occurs in Leg 2 on 25 September, with the smallest occurring on 18 September (Table 3). The difference in LHF between 18 September and Leg 1 on the  $25^{th}$  is only 8 W m<sup>-2</sup>, with the smaller ABL depth on the  $25^{th}$  leading to a smaller flux despite a slightly higher wind speed and specific humidity difference.

	18 Sept.	23 Sept.	23 Sept.	25 Sept.	25 Sept.		
	Leg 3	Leg 1	Leg 2	Leg 1	Leg 2		
<i>WS</i> (m s <sup>-1</sup> )	19.6	18.2	16.4	22.2	18.7		
ABL depth	220	564	383	76	86		
above flight							
level (m)							
ABL depth	237	157	157	160	164		
below flight							
level (m)							
Total ABL	457	721	540	236	250		
depth (m)							
Uncorrected							
$\Delta q (\mathrm{kg \ kg^{-1}})$	$5.7 \times 10^{-5}$	$7.8 \times 10^{-5}$	$8.2 \times 10^{-5}$	$6.7 \times 10^{-5}$	$1.7 \times 10^{-4}$		
LHF (W $m^{-2}$ )	19	96	68	11	98		
Space-only correction – P1 equal							
$\Delta q (\mathrm{kg \ kg^{-1}})$	$1.4 \times 10^{-4}$	9.7x10 <sup>-5</sup>	$8.7 \times 10^{-5}$	$1.1 \times 10^{-4}$	$8.2 \times 10^{-5}$		
LHF (W $m^{-2}$ )	46	119	72	19	51		
⊿LHF (W m <sup>-2</sup> )	27 (142%)	23 (24%)	4 (6%)	8 (73%)	-47 (-48%)		
Space-only correction – P2 equal							
LHF (W $m^{-2}$ )	56	121	73	20	43		
$\Delta$ LHF (W m <sup>-2</sup> )	37 (195%)	25 (26%)	5 (7%)	9 (82%)	-55 (-56%)		
Final correction – P1 equal							
$LHF (W m^{-2})$	98	151	100	56	130		
$\Delta$ LHF (W m <sup>-2</sup> )	79 (416%)	55 (57%)	32 (47%)	45 (409%)	13 (33%)		

Table 3. Values for average wind speed over the ABL (WS), ABL depth (ABL depth), specific humidity difference between pairs of profiles ( $\Delta q$ ), estimated latent heat flux (LHF), change in latent heat flux from uncorrected value ( $\Delta LHF$ ), and the correction to the observed specific humidity ( $q_{corr}$ ) for flight legs on 18, 23, and 25 September 2009. Percent difference comparing the correction to the uncorrected flux is given in parentheses.

The space-only LHF correction depends on two main factors: the specific humidity gradient downwind  $(\partial q/\partial x)$ , and the space difference  $\delta x$  (see Section 3cii), where the product of these terms given by  $q_{corr}$  (Equation 15) determines whether this correction will increase (positive  $q_{corr}$ ) or decrease (negative  $q_{corr}$ ) the uncorrected LHF. The space-only LHF corrections for the P1 equal scenario range from -47 to 27 W m<sup>-2</sup>, with the largest flux correction on 18 September and the smallest on Leg 2 25 September (Table 3). The smallest magnitude flux correction, 4 W m<sup>-2</sup>, occurred on Leg 2 23 September. The space-only correction also had the largest relative impact on the 18<sup>th</sup>, where the space-only correction adjusted the uncorrected LHF by 142%, compared to only 6% on Leg 2 on the 23<sup>rd</sup>, which had the smallest relative impact. Four of the five flights showed a positive  $q_{corr}$  value, leading to an increase in the uncorrected flux (Table 3). Leg 2 on 25 September had the only negative  $q_{corr}$  term, which led to a decrease in the uncorrected flux. The P2 equal scenario showed similar results, with the space-only correction ranging from -55 to 37 W m<sup>-2</sup>. Again, the largest relative impact on the uncorrected flux (195%) occurred on 18 September, and the smallest (7%) occurred on Leg 2 on 23 September. The space correction also reduced the uncorrected flux on Leg 2 on 25 September.

Again, because the space-only P1 and P2 LHF are so similar, the correction incorporating the LHC below flight level is only applied to the space-only P1 corrected fluxes. These final fluxes are the best estimate of the LHF, and vary from 98 to 151 W m<sup>-2</sup> (Table 3). The largest correction terms come on the 18<sup>th</sup> and on Leg 1 on the 25<sup>th</sup> at 416% and 409%, respectively, which is an increase of 274% and 336% from the space-only corrected fluxes. On the 18<sup>th</sup> and Leg 1 on the 25<sup>th</sup> the depth of the ABL below the flight level (237 and 164 m) is large relative to the total depth of the ABL (457 and 250 m) and is the reason for the large correction on these flight legs.

#### *d.* Uncertainty

There are three main areas of uncertainty associated with estimating the fluxes as described above. This uncertainty comes from using time or space corrections for the non-Lagrangian nature of the UAV observations, from measurement uncertainty (instrument errors), and from the choice of ABL depth. As discussed above, the time versus space will not be considered here, but the other two uncertainties are detailed below. Since it has been shown above that the P1 and P2 equal methods provide very similar flux estimates, only the P1 equal method fluxes will be discussed below when assessing the remaining two of these uncertainties.

All instruments operate with a certain degree of uncertainty. It is important to assess how this uncertainty impacts the calculated fluxes. To do this, we randomly perturbed the UAV observations of temperature, pressure, relative humidity, and wind speed within the degree of uncertainty of each instrument. The uncertainty was based on specifications provided by each manufacturer, with accuracies of  $\pm - 0.1^{\circ}$ C for temperature (Vaisala HMM213),  $\pm - 0.3$  hPa for pressure (Vaisala PTB110), and  $\pm - 3\%$  for relative humidity (Vaisala HMM213). Because of the way wind speed is measured, there is no manufacturer specified uncertainty, but McGeer and Holland (1993) and Holland et al. (1992) found Aerosonde® to have a wind speed accuracy within  $\pm - 1$  m s<sup>-1</sup>. Kocer et al. (2011) found the UAV (not an Aerosonde®) in their study to be accurate to within  $\pm - 0.7$  m s<sup>-1</sup>. A value of  $\pm - 1$  m s<sup>-1</sup> was used in our assessment of instrument uncertainty.

Five combinations of perturbations of the measured atmospheric variables were tested. The first involved perturbing all four variables within each instrument's degree of uncertainty. This most closely follows reality when the UAV collects measurements. The remaining four combinations involved only perturbing one of the atmospheric variables at a time while leaving the remaining three unchanged (i.e., as measured by the UAV). Testing these individual perturbations allows an examination of the sensitivity of each measurement on the calculated flux. Each flux was calculated 500 times using random perturbations, and an average flux and standard deviation across all trials was found. This average flux and standard deviation were then compared to the unperturbed SHF and LHF to determine the uncertainty.

Table 4 shows values of the SHF and LHF for all flight legs for each of the five instrument uncertainty combinations. The average fluxes for the uncertainty calculations show that for all combinations both the SHF and LHF do not vary by more than 2 W m<sup>-2</sup> from the final best estimate flux values. Most flight days had a standard deviation within 5-15 W m<sup>-2</sup> for the SHF and less than 5 W m<sup>-2</sup> for the LHF. On 18 September, the standard deviation was over 30 W m<sup>-2</sup> for the SHF and this was due primarily to sensitivity to the imposed pressure errors on this leg. The standard deviations also show that the perturbed SHF is least sensitive to changes in relative humidity, and generally most sensitive to changes in temperature and pressure, which is consistent with the way the flux is calculated. The LHF is most sensitive to changes in relative humidity, which is also consistent with how this flux is estimated.

Because the fluxes are sensitive to the selection of ABL depth, it is important to assess the uncertainty in the fluxes due to different ABL heights by choosing other scientifically reasonable heights to calculate the fluxes. Changes in the fluxes when changing the ABL depth will result from two effects – the positive or negative change in the heat (moisture) content due to increasing or decreasing the mass of air being warmed (moistened) as the ABL depth changes, and changes in the atmospheric conditions within that ABL depth, which will alter the heat and

	18 Sept.,	23 Sept.,	23 Sept.,	25 Sept.,	25 Sept.,		
	Leg 3	Leg 1	Leg 2	Leg 1	Leg 2		
Sensible Heat Flux							
			1				
Best estimate SHF	232	472	185	385	12		
SHF (all)	230 +/- 33	471 +/- 13	185 +/- 11	385 +/- 6	12 +/- 6		
SHF (T only)	232 +/- 10	472 +/- 9	185 +/- 8	385 +/- 4	13 +/- 4		
SHF (p only)	234 +/- 32	471 +/- 9	185 +/- 8	385 +/- 5	12 +/- 5		
SHF (RH only)	232 +/- 0	472 +/- 0	185 +/- 0	385 +/- 0	12 +/- 0		
SHF (WS only)	232 +/- 0.8	472 +/- 1	185 +/- 0.8	385 +/- 1	12 +/- 0.1		
Latent Heat Flux							
Best estimate LHF	98	151	100	56	130		
LHF (all)	98 +/- 5	151 +/- 4	100 +/- 4	56 +/- 2	130 +/- 3		
LHF (T only)	98 +/- 0.6	151 +/- 0.4	101 +/- 0.5	56 +/- 0.1	130 +/- 0.2		
LHF (p only)	98 +/- 0.3	151 +/- 0.1	100 +/- 0.1	56 +/- 0	130 +/- 0.1		
LHF (RH only)	98 +/- 5	151 +/- 4	100 +/- 4	56 +/- 2	130 +/- 3		
LHF (WS only)	98 +/- 0.3	151 +/- 0.5	100 +/- 0.4	56 +/- 0	130 +/- 0.4		

Table 4. Average fluxes (W m<sup>-2</sup>) found when varying atmospheric parameters based on instrument specifications, plus one standard deviation. The best estimate flux values listed are the P1 equal final fluxes for the SHF and LHF. All refers all observations being perturbed. T only refers to perturbations of temperature, p only refers to perturbations of pressure, RH refers to perturbations of relative humidity, and WS refers to perturbations of wind speed.

layer. If the change in ABL depth alters the atmospheric conditions within the layer, then the flux will increase or decrease depending on whether the change in the atmospheric variables increases or decreases the temperature (moisture) difference between pairs of profiles. The individual sensitivity of each effect to the final flux estimation will vary for each case. Alternative selections for the ABL depth outlined in Section 4a and the impact of changing the height of the ABL on the best estimates of the SHF and LHF (space-adiabatic and space-only) are assessed below.

On 18 September (Figure 21), the top of the ABL could be 625 m. Above this point, the potential temperatures in both the upstream (Profile 3) and downstream (Profile 4) profiles are very similar, whereas below this level the upstream profile is generally colder than the downstream. This layer is unlikely to be the top of the ABL, due to the distinctness of the dry-adiabatic layer below, but an assessment of this ABL height can still be made. The increase in the depth of the ABL from 457 m to 625 m yields a SHF of 321 W m<sup>-2</sup> and a LHF of 111 W m<sup>-2</sup>, which is a 90 and 18 W m<sup>-2</sup> (39% and 13%) increase, respectively. In this instance, the increase in fluxes is due to both an increase in the ABL depth and an increase in the temperature and specific humidity differences between the pair of profiles.

The ABL top on 18 September could also be ~790 m, the level above which the upstream and downstream potential temperatures are nearly identical. This 333 m increase in the ABL depth leads to a SHF of 212 W m<sup>-2</sup> and a LHF of 109 W m<sup>-2</sup>, which is a 19 W m<sup>-2</sup> decrease and a 11 W m<sup>-2</sup> increase (-8% and 11%) in the SHF and LHF, respectively. The slight increase in these fluxes is likely due mostly to the increase in ABL depth, as the temperature and humidity differences will not change much between the profiles from 457 to 790 m (Figure 21).

On Leg 1 on 23 September, it is possible to consider an ABL top of approximately 280 m (Figure 22). Below this level is the presence of a mixed layer, with a capping inversion at the top of this layer. Calculating the flux with an ABL top of 280 m on this day yields a SHF of 337 W m<sup>-2</sup> and a LHF of 82 W m<sup>-2</sup>. This decreases the fluxes by 135 and 69 W m<sup>-2</sup> (-29% and -46%) respectively. Despite an increase in the temperature difference, wind speed, and specific humidity difference the SHF and LHF still decrease, which is solely due to a decrease in the ABL top. This ABL height was not chosen over the 720 m level because there is a warming between 280 and 720 m that is unexplained if we assume the ABL top is at 280 m.

For Leg 2 on 23 September, a ABL height of 500 m could be chosen as the level above which the downstream profile becomes colder than the upstream profile (Figure 23). This could indicate that surface processes are no longer controlling the atmosphere at this point, because the atmosphere should continue warming throughout the ABL. The clear well-mixed layer on this flight leg, however, yields the 540 m level to be the appropriate ABL height rather than at 500 m. When the SHF and LHF are calculated using 500 m as a ABL top, the fluxes become 207 and 93 W m<sup>-2</sup>, increasing the SHF by 22 W m<sup>-2</sup> and decreasing the LHF by 7 W m<sup>-2</sup> (12% and -7%). The wind speed change was negligible, but the temperature difference increased by 0.1K while the specific humidity difference decreased very slightly. Because there was only a 40 m difference in ABL height between the two cases, the fluxes were most sensitive to the changes in atmospheric conditions on this flight leg.

For Legs 1 and 2 on 25 September, no other scientifically reasonable values could be chosen for the ABL height outside of that selected for this study (Figures 24-25). On Leg 1, 240 m is the location of both an inversion as well as where the two profiles exhibit similar characteristics.

The areas of uncertainty described above aid our understanding of the accuracy of the SHF and LHF estimated as part of this work. Final estimates of the fluxes for the UAV flights in September 2009 with error bars based on the uncertainty assessment described above are provided in Table 5. Not all estimates described above, however, are appropriate to include in this analysis. For example, the varying ABL heights are not appropriate to include in the assessment because those levels are deemed to be less scientifically justifiable compared to the levels described in Section 4a. It is appropriate to include the standard deviations from the instrument uncertainty assessment because these are viable uncertainties in the fluxes. Including the uncertainty associated with the P1 and P2 equal space methods is appropriate as well. The uncertainty associated with the time corrections are not included due to the issues with the local time rate of change term for the 2009 data, but if used in other field campaigns it is appropriate to include this uncertainty as well.

In Table 5, it is clear that overall there is not much uncertainty in the fluxes calculated using the methods described in this paper, with variations of no more than 22 W m<sup>-2</sup> for the SHF and 14 W m<sup>-2</sup> for the LHF. The greatest overall uncertainty is on 18 September, where the instrument uncertainty is relatively large.

	18 Sept.	23 Sept.	23 Sept.	25 Sept.	25 Sept.
	Leg 3	Leg 1	Leg 2	Leg 1	Leg 2
SHF (W m <sup>-2</sup> )	209 to 231	472 to 485	176 to 185	370 to 385	12 to 30
	+/- 33	+/- 13	+/- 11	+/- 6	+/- 6
LHF (W $m^{-2}$ )	98 to 112	151 to 152	100 to 101	56 to 58	117 to 130
	+/- 5	+/- 4	+/- 4	+/- 2	+/- 3

Table 5. Final values for the SHF and LHF including all relevant uncertainties.

## 5. Summary and Conclusions

In the Terra Nova Bay region of Antarctica, cold, dry air flowing from the continental interior over the relatively warm and moist coastal polynya introduces the potential for strong air-sea fluxes to occur. In September 2009, several flights were conducted using Aerosonde® UAVs to collect three-dimensional measurements of the atmosphere overlying the polynya. SHF and LHF were estimated by assessing the change in heat and moisture content over the depth of the ABL from pairs of vertical profiles collected along the flight path as the UAV moved downwind over TNB. Spatial and temporal corrections for the non-Lagrangian nature of the observations were considered, but due to the observational data available it was concluded that only the spatial correction was reasonable for the 2009 flights. It was determined that, for the SHF, a flux correction accounting for both adiabatic and non-Lagrangian spatial processes (space-adiabatic), as well as the HC below flight level, was most appropriate, while the spatial non-Lagrangian fluxes (space-only) incorporating LHC below flight level were most appropriate for the LHF. Over the three flight days used to calculate the heat fluxes in 2009, the final SHF ranged from 12 to 485 W m<sup>-2</sup>, with the final LHF ranging from 56 to 152 W m<sup>-2</sup>. The methodology developed shows that the space correction can be implemented in two different but scientifically appropriate ways that provide very similar final answers.

One of the main goals of this work is to highlight sensitivities in the UAV observations that are important to the methodology used in estimating the fluxes. This will enable future field campaigns to better design and implement measurement collection. The methodology discussed here requires a few key measurements to accurately estimate the SHF and LHF. First, accurate collection of profile data within and higher than the atmospheric ABL is critical. This data should be acquired as frequently as possible along the flight path to ensure correction terms are small. Second, the data needs to be corrected considering adiabatic (for SHF) and non-Lagrangian processes (for both SHF and LHF), as well as for the HC and LHC below flight level, and this work has shown that it is imperative to acquire accurate measurements of the atmospheric variables included in these calculations. For the adiabatic correction, appropriate pressure measurements were critical, and for the non-Lagrangian correction, accurate estimates of the along path gradients of temperature, pressure, and specific humidity. To minimize the correction due to the unsampled HC and LHC below flight level, ensuring flight data as close to the surface as possible is necessary. Missing from this study were appropriate measurements of changes in the atmosphere over time, such as from a nearby AWS or from the UAV data itself. The work conducted during the 2009 flights highlighted this critical need, and a future field campaign over TNB in 2012 ensured appropriate time-varying measurements from the UAVs. For both campaigns, work is currently underway to assess why the fluxes vary from day to day and over different locations within TNB, and what processes impact the depth of the ABL.

#### 6. Acknowledgments

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# Chapter 4: The primary mechanisms of air-sea heat exchange over Terra Nova Bay, Antarctica

ABSTRACT. In September 2009, unmanned aerial vehicles (UAVs) were flown over Terra Nova Bay (TNB) Antarctica to sample the atmospheric boundary layer (ABL) overlying a polynya. In TNB, a relatively warmer and moister surface associated with the open water interacts with cold, dry air originating from the high polar plateau to generate large sensible and latent heat fluxes. Atmospheric measurements collected during these flights were used to estimate these fluxes in a previous study. Through the use of data from the UAVs and satellites, this study examines the primary forcing mechanisms for heat exchange during the flights. It is shown that, in September 2009, surface conditions vary much more than atmospheric conditions, and the heat fluxes respond primarily to these changes. The UAV estimates are compared to the Tropical Ocean-Global Atmosphere (TOGA) Coupled Ocean-Atmosphere Response Experiment (COARE) bulk flux algorithm, which incorporates both atmospheric and surface data. The COARE fluxes have a much higher degree of uncertainty compared to the UAV fluxes due to uncertainty in surface temperatures, showing the necessity for accurate measurements of surface conditions. The impacts of heat exchange on the temperature structure of the ABL are also examined. It is shown that a convective well-mixed layer is formed on most flight days, which increases in height as the air passes over the polynya. The well-mixed layer erodes a residual stable layer originating from the polar plateau, generating a new ABL over the open water that is distinct from its previous structure.

## 1. Introduction

The Terra Nova Bay (TNB) region of Antarctica is located in the western Ross Sea (Figure 26). TNB is unique for its topographical, atmospheric, cryospheric, and oceanographic influences (Kurtz and Bromwich 1985; Bromwich 1989; van Woert 1999; Morales Maqueda et al. 2004). A large drainage area of katabatic winds to the west of TNB forms a confluence zone where winds sharply increase just upstream of the Reeves Glacier (Parish 1982; Bromwich and Kurtz 1984; Parish and Bromwich 1987). This confluence zone leads to an increase in winds that can aid in the formation of a polynya at the coast, where sea ice is formed at its western edge and pushed further into the Ross Sea by the strong winds. The Drygalski Ice Tongue, an extension of David Glacier that extends approximately 50 km into TNB, prevents sea ice from advecting into the region from the south, leaving open water that remains even during the winter season (Kurtz and Bromwich 1985) (Figure 27). As cold and dry air moves over the polynya from the continent, air-sea interactions result in large heat and momentum fluxes (Knuth and Cassano 2014). During winter, when winds are strongest and the atmosphere is coldest and driest (Knuth and Cassano 2011), the largest sensible and latent heat fluxes (SHF and LHF) will occur.

The atmospheric boundary layer (ABL) over the polynya is modified by the SHF and LHFs such that heat added at the surface warms and moistens the ABL (Renfrew and King, 2000; Heinemann 2008; Raddatz et al. 2011). The entrainment of air at the top of the ABL causes the depth of the ABL to increase with distance from the coast as well as alter the heat and moisture content (Chang and Braham, 1991; Renfrew and King, 2000; Heinemann 2008). With increasing fetch, the ABL warms and moistens such that the temperature and moisture gradient



Figure 26. Map of the TNB region (top panel) within the Ross Sea sector of Antarctica (bottom panel).



Figure 27. MODIS Terra Visible 0.25 km satellite image from 23 September 2009 at 21:40 UTC. This image is valid approximately four hours before the UAV flights were flown, due to the lack of a 0.25 image during the flight.

between the atmosphere and the surface diminishes and heat fluxes decrease (Heinemann 2008). In TNB, the cold air modification over the polynya is akin to cold air outbreaks occurring over other areas across the globe (Grossman and Betts, 1990; Chang and Braham 1991).

In September 2009, Aerosonde® unmanned aerial vehicles (UAVs) were flown over TNB to collect information on the three-dimensional properties of the atmosphere overlying the polynya (Cassano et al. 2010; Knuth et al. 2013). These measurements were the first in situ measurements of the TNB air mass during the winter months, and provided invaluable information on the ABL. Knuth and Cassano (2014) (hereafter KC2014) used the data collected by the UAVs to develop a methodology of estimating SHF and LHF between the ocean and atmosphere without the use of a bulk aerodynamic formula. Estimated SHFs over TNB in September 2009 from this study ranged from 12 to 485 W m<sup>-2</sup>, while the LHFs ranged from 56 to 152 W m<sup>-2</sup> (Tables 6 and 7).

The purpose of this study is to understand the primary forcing mechanisms of oceanatmosphere surface heat exchange acting over the TNB polynya. This paper applies UAV and satellite observations to examine how changing atmospheric and surface conditions over TNB impact SHFs and LHFs estimated from the UAV data as part of KC2014. The UAV estimates, which incorporate only atmospheric data, are then compared against a bulk flux algorithm (which uses both surface and atmospheric data) to understand differences between the two flux estimates. The three flight days examined – 18, 23, and 25 September – represent varying physical processes that are explored in this study. Section 2 briefly summarizes the flux methodology developed by KC2014, the data collected during the UAV flights, and the Tropical Ocean-Global Atmosphere (TOGA) Coupled Ocean-Atmosphere Response Experiment (COARE) bulk flux algorithm. Section 3 discusses the atmospheric and surface conditions on each of the three flux days to understand the primary forcing mechanisms for heat exchange, and Section 4 shows comparisons of COARE to the UAV fluxes. A summary is found in Section 5.

	COARE-	COARE-	UAV	COARE-E	COARE-M
	E Flux	M Flux	Flux	mean	mean
				comparison	comparison to
				to UAV	UAV
Leg 1 on 23		530	472 to		50 (11%)
September		+/-50	485 +/-		
			13		
Leg 2 on 23		120	176 to		45 (24%)
September		+/-33	185 +/-		
			11		
Leg 1 on 25	305	500	370 to	75 (20%)	125 (33%)
September	+/- 63	+/- 45	385 +/-		
			6		
Leg 2 on 25	20	88	12 to 30	<5 (25%)	65 (325%)
September	+/- 42	+/- 33	+/- 6		
Leg 3 on 18	-20	100	209 to	245	90 (72%)
September	+/- 35	+/- 35	231 +/-	(108%)	
			33		

Table 6. COARE-E, COARE-M, (bulk flux algorithms) and UAV SHF (from KC2014) are given. Also shown are the comparisons between the COARE and UAV fluxes. Values are in W m<sup>-2</sup>. Values in parenthesis are percent differences from the UAV values. Bolded values show the COARE and UAV flux comparisons that are within the range of uncertainty of the pair of flux estimates.

## 2. Data and Methods

# *a.* UAV fluxes methodology

The UAVs flown over TNB in September 2009 were manufactured by Aerosonde®, and had a wingspan of 3 m, a payload capacity of 2-5 kg, and a weight of 15 kg (Knuth et al. 2013). Temperature, pressure, and relative humidity sensors were on-board the UAVs, while wind speed measurements were made using the plane's pitot tube and GPS sensors. Wind direction was found during aircraft maneuvers.

	COARE	COARE	UAV	COARE-E	COARE-M
	-E Flux	-M Flux	Flux	mean	mean
				compariso	comparison to
				n to UAV	UAV
Leg 1 on 23		200	151 to		50 (33%)
September		+/- 22	152 +/- 4		
Leg 2 on 23		58	100 to		45 (45%)
September		+/- 10	101 +/- 4		
Leg 1 on 25	115	195	56 to 58	60 (107%)	140 (212%)
September	+/- 22	+/- 12	+/- 2		
Leg 2 on 25	33	50	117 to	90 (73%)	50 (41%)
September	+/- 15	+/- 10	130 +/- 3		
Leg 3 on 18	19	52	98 to 112	85 (81%)	55 (52%)
September	+/- 12	+/- 8	+/- 5		

Table 7. COARE-E, COARE-M, (bulk flux algorithms) and UAV LHF (from KC2014) are given. Also shown are the comparisons between the COARE and UAV fluxes. Values are in W m<sup>-2</sup>. Values in parenthesis are percent differences from the UAV values. Bolded values show the COARE and UAV flux comparisons that are within the range of uncertainty of the pair of flux estimates.

Three UAV missions (on the  $18^{th}$ ,  $23^{rd}$ , and  $25^{th}$ ) were flown during September 2009 that collected useful data for estimating air-sea fluxes (Cassano et al. 2010). On these flights, the plane flew parallel to the coastline, searching for the area of strongest winds. The UAV was then flown downstream from the coast within the strong katabatic jet to sample the area where the heat fluxes would be largest (KC2014; Knuth et al. 2013). Along this downstream flight leg, vertical profiles were collected at ~20 km intervals within the jet (Figure 19). The flights on the  $23^{rd}$  and  $25^{th}$  both had three profiles, while the flight on the  $18^{th}$  had two. The  $18^{th}$  originally had four profiles collected, but only Profiles 3 and 4 were useful for this study.

KC2014 developed a methodology for estimating SHF and LHF using only atmospheric state observations collected during these flights. The basic premise behind this method was to use the UAV observations within the vertical profiles to estimate how the heat and moisture content within the ABL changes as the air mass moves over the polynya. This approach

provides one estimate of heat flux between pairs of profiles. For example, on 23 September, when three profiles were collected, two fluxes are estimated – one between Profiles 1 and 2, and one between Profiles 2 and 3 (Figure 19). The convention of this paper is that Leg 1 refers to data collected or fluxes estimated between Profiles 1 and 2, Leg 2 between Profiles 2 and 3, and Leg 3 between Profiles 3 and 4. A summary of the SHF and LHF found on each flight day, including the range of uncertainty in the estimates, is given in Tables 6 and 7.

#### b. Data Description

The surface temperature ( $T_{sfc}$ ) data used within this study comes from two sources: an infrared thermometer that flew on the UAVs (Everest Interscience, Inc., model 3800ZL), and satellite data from the Moderate Resolution Imaging Spectroradiometer (MODIS) (Hall et al. 2004). The data produced from the Everest was found to have issues related to the housing temperature that were corrected in post-processing (Knuth et al. 2013). The post-processing calibration method found the uncertainty of the Everest infrared T<sub>sfc</sub> to be within 2 K. Everest T<sub>sfc</sub> data was not available on 23 September.

Data acquired from the National Aeronautics and Space Administration (NASA) Level 1 and Atmosphere Archive and Distribution System website (LAADS) was used to calculate the  $T_{sfc}$  values from MODIS. One MODIS product was acquired per day over limited time frames to minimize the effects of sun angle on reflectance. Thermal bands 33 and 34 were used to generate a  $T_{sfc}$  image, which was reduced to include only data from the TNB region, with a mask applied to exclude land areas. The data was geo-located to the UAV flight path using Level 3 1km data for each pixel of the Level 1B data. In general, the accuracy (root mean square error) of the MODIS  $T_{sfc}$  data is found to be within +/- 1.3 K (Hall et al. 2004). MODIS 0.25 km visible satellite images were also acquired from the LAADS website to allow visual inspection of sea ice cover across the western Ross Sea region.

Satellite observations from the Advanced Microwave Scanning Radiometer for Earth Orbiting System (AMSR-E) (Lobl 2001) are used to calculate sea ice concentration. The NASA Team 2 and Bootstrap algorithms were applied to the 12.5 km AMSR-E data to determine sea ice concentration. Finally, digital photographs from a Canon SX110 IS digital camera mounted on the UAVs provided images of the surface conditions below the UAV during the flight on 23 September. Low light prevented the availability of these images for the flights on 18 and 25 September.

#### c. COARE bulk flux algorithm

The flux methodology as part of KC2014 estimates the SHF and LHF using only atmospheric state information. In contrast, the COARE bulk flux algorithm uses both atmospheric and surface conditions, and through its use, the sensitivity of the heat fluxes to the surface state can be further explored.

SHF and LHF can be parameterized using the bulk aerodynamic formulae through the use of mean quantities of atmospheric and surface conditions together with transfer coefficients (Pond et al. 1974). The SHF and LHF (in W m<sup>-2</sup>) can be expressed by

$$SHF = \rho c_p C_H U(T_{sfc} - T_{air}) \tag{1}$$

$$LHF = \rho L_{\nu} C_E U(q_{sat} - q_{air}) \tag{2}$$
where  $\rho$  is the density of the air (kg m<sup>-3</sup>),  $c_p$  is the specific heat of dry air at constant pressure (1004 J kg<sup>-1</sup> K<sup>-1</sup>),  $C_H$  and  $C_E$  are the dimensionless bulk transfer coefficients for heat and moisture, U is the wind speed,  $T_{sfc}$  and  $T_{air}$  are the surface and ambient temperatures,  $L_v$  is the latent heat of vaporization at 0°C (2.5x10<sup>6</sup> J kg<sup>-1</sup>), and  $q_{sat}$  and  $q_{air}$  are the surface saturation and air specific humidities.

In this work, the COARE 3.0 algorithm represents the bulk flux formulae. The COARE algorithm was initially developed based on measurements taken during a field program over the western Pacific, where the algorithm took into account low wind speeds and tropical conditions (Fairall et al. 2003). COARE has since been updated so that version 3.0 is "globalized" in the sense that, whenever possible, variables are calculated rather than estimated from tropical conditions. As well, data outside the tropics have been used to estimate inputs to the transfer coefficients such that the COARE 3.0 algorithm is valid up to 20 m s<sup>-1</sup>. It is important to note that the wind speeds during September 2009 in TNB are on the upper end or outside of this range, which may impact the quality of the COARE estimated fluxes.

To estimate SHF and LHF using COARE, the ambient atmospheric conditions (temperature, humidity, pressure, and wind speed) measured by the UAV and collected at heights ranging from  $\sim$ 150-240 m were input into the COARE algorithm via the bulk algorithm package. Specific humidity was calculated from the UAV observations of relative humidity, temperature, and pressure. Separate COARE estimates were made using T<sub>sfc</sub> observations from the Everest and MODIS datasets.

The accuracy of the COARE algorithm is largely dependent on the quality of the MODIS and Everest  $T_{sfc}$  data. It is anticipated that the Everest data, despite some of the issues with the instrument (Knuth et al. 2013), is more accurate than the MODIS data over areas where sea ice

changes rapidly (such as Legs 1 and 2). This is largely due to the finer spatial resolution of Everest compared to MODIS. The MODIS data has a 1 km resolution, while the Everest data were obtained with a field of view of approximately 30 m, with one observation recorded approximately every 10 seconds. All other atmospheric observations from the UAV were also recorded approximately every 10 seconds. The MODIS data is collected at slightly different times than the UAV flights (about one hour later on the 18<sup>th</sup>, an hour and thirty minutes later on the 23<sup>rd</sup>, and about 30 minutes earlier on the 25<sup>th</sup>) and this increases the uncertainty in the MODIS T<sub>sfc</sub> values for estimating fluxes coincident with the time of the UAV flights. Over areas with more uniform sea ice, such as parts of Leg 2 and Leg 3, MODIS could produce a better T<sub>sfc</sub> estimate than Everest due to a smaller uncertainty (+/- 1.3K compared to the Everest +/- 2K), and due to issues in post-processing making Everest T<sub>sfc</sub> biased toward open water cases (Knuth et al. 2013).

# 3. Primary Forcing Mechanisms of Heat Exchange – Comparing Atmospheric and Surface Conditions

This section compares the atmospheric and surface conditions to determine which of these is dominant in controlling the energy exchange between the ocean-atmosphere interface. Generally, it is expected that the atmosphere will warm, moisten, and lose momentum as the air mass originating from the high polar plateau passes over the open water of the polynya (Roberts et al. 2001). It is also anticipated that sea ice concentrations will increase and  $T_{sfc}$  will decrease as sea ice is generated in the polynya and pushed east by the winds (Kurtz and Bromwich 1985; Walter et al. 2006; Fiedler et al. 2010). Both factors should work together to produce a

decreasing SHF and LHF with increasing distance along the polynya downwind of the coast (Kurtz and Bromwich 1985; Walter et al. 2006; Fiedler et al. 2010; KC2014). To quickly and succinctly identify changes between the atmospheric and surface state, two terms, the "vertical temperature difference" and "vertical moisture difference" are defined, representing the difference between  $T_{sfc}$  and the atmospheric temperature ( $T_a$ ;  $T_{sfc} - T_a$ ) and the difference between the surface saturation mixing ratio ( $q_{sfc}$ ) and the atmospheric mixing ratio ( $q_a$ ;  $q_{sfc} - q_a$ ), respectively.

It is important to also understand how the data from this study compares to that used in estimating the UAV fluxes from KC2014. Figures 28 and 29 show observations from the UAV during low altitude flight legs (Table 8) between concurrent profiles. The data used to estimate the SHFs and LHFs in KC2014 are from the profiles and not from this along flight leg data. However, using this data allows for a better examination of changing conditions between the atmosphere and the surface both spatially and temporally. These changing conditions will impact the entire ABL, and thus are representative of the conditions measured in the vertical profiles.

Figure 28 shows the UAV SHF,  $T_a$ ,  $T_{sfc}$ , wind speed, and sea ice concentration for 18, 23, and 25 September. The observations of  $T_a$  and wind speed are from the UAV as it flies along a generally constant altitude between profiles (Table 8). Figure 29 is similar to Figure 28 but with LHF and mixing ratio in place of SHF and temperature.



Figure 28. SHF and relevant meteorological parameters for the  $18^{th}$  (blue),  $23^{rd}$  (red), and  $25^{th}$  (black) of September, 2009. X-axis is distance along the flight leg. The data from 18 September actually begins ~40 km downstream of the coast, but is plotted to the left of the graph for aesthetic purposes.  $T_{sfc}$  and corresponding data is given from both the MODIS and Everest datasets, where the MODIS data is represented by a  $\otimes$  symbol, and the Everest data by a solid circle.



Figure 29. LHF and relevant meteorological parameters for the  $18^{th}$  (blue),  $23^{rd}$  (red), and  $25^{th}$  (black) of September, 2009. X-axis is distance along the flight leg. The data from 18 September actually begins ~40 km downstream of the coast, but is plotted to the left of the graph for aesthetic purposes.  $q_{sfc}$  and corresponding data is given from both the MODIS and Everest datasets, where the MODIS data is represented by the  $a \otimes$  symbol, and the Everest data by a solid circle.

	Altitude (m)	Average Wind Speed
		$(m \ s^{-1})$
18 September, Leg 3	237	19.6
23 September, Leg 1	157	18.2
23 September, Leg 2	157	16.4
25 September, Leg 1	164	22.2
25 September, Leg 2	164	18.7

# Table 8. The altitude the UAV was flown and the wind speed averages for each flight leg in September 2009.

#### a. 23 September

On the  $23^{rd}$  (red lines and symbols on Figures 28 and 29), the surface and atmospheric conditions behave as outlined above, with T<sub>a</sub>, q<sub>a</sub>, and sea ice concentration increasing and T<sub>sfc</sub>, q<sub>sfc</sub>, and wind speed decreasing with increasing fetch. Considering first the surface conditions, Figure 28 (red symbols) shows the MODIS T<sub>sfc</sub> data decreases from 270 to 256 K along the entire flight leg, while Figure 29 shows the MODIS q<sub>sfc</sub> data decreasing from 3.0 g kg<sup>-1</sup> to 1.0 g kg<sup>-1</sup>. Only MODIS (no Everest) data is available on this flight. Sea ice concentrations increase from 0.59 to 0.83 along the leg. In Leg 1, digital photographs (Figure 30) show surface conditions changing from open water with minimal ice cover to pancake ice. In Leg 2, the surface is more uniform, making a transition from pancake ice to pack ice.

The atmospheric conditions shows  $T_a$  increases ~1.9 K along the entire flight leg, from 249.5 to 251.4 K (Figure 28). The  $q_a$  increases 0.33 g kg<sup>-1</sup> from ~0.16 to 0.49 g kg<sup>-1</sup> (Figure 29).



Figure 30. Digital images from a camera on board the UAV during Legs 1 and 2 on 23 September 2009. a) depicts an image taken at the location of Profile 1, b) at Profile 2, and c) at Profile 3.

Wind speeds decrease  $\sim 8 \text{ m s}^{-1}$  from  $\sim 24$  to 15.9 m s<sup>-1</sup>. Both the SHF and LHF values estimated from the UAV methodology (Tables 6 and 7) decrease with increasing fetch over the polynya.

To understand whether the surface or atmospheric state is the primary forcing mechanism for heat exchange on this flight, the downstream changes in the components of the vertical temperature and moisture differences are compared ( $T_{sfc}$  and  $T_a$ , and  $q_{sfc}$  and  $q_a$ ). The vertical differences should decrease with increasing fetch as the atmosphere gets warmer due to heat transfer from the ocean and the  $T_{sfc}$  gets cooler due to increased sea ice concentrations (Bromwich and Kurtz 1984; Fiedler et al. 2010).

Figure 28 shows the vertical temperature difference to be 20 K at the beginning of Leg 1, falling to less than 5 K by the end of Leg 2. This 15 K change in the vertical temperature difference with increasing fetch is much more representative of the 14 K change in  $T_{sfc}$  along the flight leg than the ~2 K change in  $T_a$ . Similarly, the vertical moisture difference changes from 3 g kg<sup>-1</sup> at the beginning of Leg 1 to 1 g kg<sup>-1</sup> by the end of Leg 2 (Figure 29), which is more representative of the 2.0 g kg<sup>-1</sup> change in  $q_{sfc}$  than the ~0.3 g kg<sup>-1</sup> change in  $q_a$ . These results show that changes in the surface conditions are more influential in changing the vertical temperature and moisture differences than the atmospheric state.

To quantify whether the 14 K change in the vertical temperature difference or the 8 m s<sup>-1</sup> wind speed change is a primary factor controlling the observed decrease of ~300 W m<sup>-2</sup> in SHF between Legs 1 and 2 (or similarly the 2.0 g kg<sup>-1</sup> vertical moisture difference for the observed decrease of ~50 W m<sup>-2</sup> in LHF between Legs 1 and 2) (Table 6), sensitivity studies were conducted with the COARE algorithm to understand how changes in the vertical temperature and moisture difference and wind speed would impact heat fluxes. When the sensitivity study is run decreasing the vertical temperature difference by 14 K, the COARE SHF decreases by ~450 W m<sup>-2</sup>. Decreasing the wind speed 8 m s<sup>-1</sup> results in a decrease of the COARE SHF by ~200 W m<sup>-2</sup>. A 2.0 g kg<sup>-1</sup> decrease in vertical moisture difference results in a ~155 W m<sup>-2</sup> decrease in the LHF. Decreasing the wind speed by 8 m s<sup>-1</sup> results in a ~50 W m<sup>-2</sup> decrease in the LHF. While the magnitude of the SHF and LHF changes calculated as part of this sensitivity study do not match

the observed change in fluxes between Legs 1 and 2 (a more detailed comparison of the COARE and UAV estimated fluxes is given in Section 4) this sensitivity study indicates that the decrease in UAV observed fluxes between the two flight legs is dominated by the decreasing vertical temperature and moisture differences with the decrease in wind speed playing a secondary role. Based on this sensitivity analysis as well as the dependence of the vertical temperature and moisture differences on the surface conditions, it is determined that the changes in the surface state is the primary forcing mechanism for changes in heat exchange between the ocean and atmosphere on 23 September.

### b. 25 September

The primary forcing mechanism for heat exchange on the 25<sup>th</sup> (black lines and symbols on Figures 28 and 29) can be found similarly to the method on the 23<sup>rd</sup>. On the 25<sup>th</sup>, both Everest and MODIS T<sub>sfc</sub> values are available. The MODIS T<sub>sfc</sub> values (black  $\otimes$  symbol in Figure 28) show a decrease of 14 K along the flight leg, from 270 to 256 K. The Everest values (black filled circles) also show a 14 K decrease over the flight leg, with a decrease from 266 to 252 K. The MODIS q<sub>sfc</sub> decreases by 2.0 g kg<sup>-1</sup> from 3.0 to 1.0 g kg<sup>-1</sup>, and the Everest q<sub>sfc</sub> decreases 1.5 g kg<sup>-1</sup> from 2.3 to 0.8 g kg<sup>-1</sup> (Figure 29). Sea ice concentrations increase from 0.6 to 0.83 (Figures 28, 29, and 31). Examining the atmospheric values, the T<sub>a</sub> increased 1.2 K (from 250.8 to 252 K) over Legs 1 and 2, while q<sub>a</sub> shows an increase of ~0.36 g kg<sup>-1</sup> in both Legs 1 and 2 (Figures 28 and 29). Wind speeds along the flight leg decrease approximately 11 m s<sup>-1</sup>, from 28 to 17 m s<sup>-1</sup>.



Figure 31. MODIS Terra Visible 0.25 km satellite image from 25 September 2009 at 21:30 UTC. This image is valid approximately 9 hours and 30 minutes after the UAV flights were flown, due to the lack of a 0.25 image during the flight.

Figure 28 shows a 15 K decrease in the MODIS and Everest calculated vertical temperature differences along the flight leg, which more closely matches the  $T_{sfc}$  change downstream (14 K) than the  $T_a$  change (1.2 K). The MODIS and Everest vertical moisture differences decrease by ~1.9 g kg<sup>-1</sup> downstream, more closely matching the  $q_{sfc}$  differences (1.5-2.0 g kg<sup>-1</sup>) than  $q_a$  difference (0.36 g kg<sup>-1</sup>). Sensitivity studies conducted also show the vertical temperature difference change to have a greater impact on heat fluxes than the wind speed

change (with results mirroring the sensitivity study on the 23<sup>rd</sup>), further demonstrating that the surface state is the primary forcing mechanism on this flight day.

Table 6 shows a decrease in the SHF on the  $25^{\text{th}}$  with increasing fetch over the polynya (by ~370 W m<sup>-2</sup>), as expected, but an increase in the LHF of ~75 W m<sup>-2</sup>. The increase of the LHF along the flight leg is not completely clear, but it is suspected that a sensor lag in the relative humidity values within the profiles impacts the q<sub>a</sub> values along Leg 2, leading to a dry bias. Closer examination shows a large decrease (0.2 g kg<sup>-1</sup>) in the mixing ratio over a very small distance (~1 km) at Profile 2, located at ~20.5 km in Figure 29. The dry bias is manifest as the UAV relative humidity sensor adjusts too slowly as the UAV descends at profile 2 from the drier air aloft to the moister near surface air. As the UAV moves downstream from Profile 2, the sensor is "catching up" to the larger relative humidity values, and falsely showing a larger mixing ratio change downwind. Difficulties in estimating the time lag yield this as only speculation at this time.

Efforts to reduce the sensor lag on all other flight legs were made by only using matching profile data (for example, only using ascent data from the two profiles), but for this flight leg, only descending data in Profile 2 and ascending data in Profile 3 were available. Because it is anticipated that the sensor lag will vary for both ascents and descents, the dry bias might be enhanced. Therefore, the UAV LHF estimates in Leg 2 are likely over estimated values.

## c. 18 September

On the 18<sup>th</sup>, the atmospheric and surface conditions also behave as expected, and do not vary much over the course of the flight leg (blue lines and symbols on Figures 28 and 29). As

this flight leg is ~40 km downstream of the coast (the furthest of all legs), over sea ice concentrations greater than 0.8 (Figures 28 and 29), and just east of the polynya (Figure 32), it is expected that the surface and atmospheric conditions will be less variable than the upwind flight legs. Examining first the surface conditions, the MODIS and Everest  $T_{sfc}$  show a ~3 K and ~2 K decrease, respectively (blue symbols in Figure 28) over the flight leg. The MODIS  $q_{sfc}$  decreases by ~0.3 g kg<sup>-1</sup> from 1.3 g kg<sup>-1</sup> to 1.0 g kg<sup>-1</sup> and the Everest  $q_{sfc}$  decreases by ~0.2 g kg<sup>-1</sup> (Figure



Figure 32. MODIS Aqua Visible 0.25 km satellite image from 18 September 2009 at 5:35 UTC. This image is valid approximately 6 hours before the UAV flights were flown, due to the lack of a 0.25 image during the flight.

29). Sea ice concentrations increase from 0.81 to 0.83. For the atmosphere,  $T_a$  increases ~1 K along the flight leg (Figure 28), while  $q_a$  increases by ~0.3 g kg<sup>-1</sup> (Figure 29). Wind speeds also decrease over the flight leg, starting at approximately 20 m s<sup>-1</sup>, increasing quickly to 24 m s<sup>-1</sup>, and then dropping back to ~18 m s<sup>-1</sup>.

Comparing the surface and atmospheric components of the vertical differences reflects the similarity between the atmosphere and surface expected over this more downwind flight leg. The MODIS vertical temperature difference decreases 3 K along the flight leg, and the Everest 2 K (Figure 28). The vertical temperature differences again more closely follow  $T_{sfc}$  (3 and 2 K, respectively) than  $T_a$  (1 K) changes downwind. However, the  $T_{sfc}$  and  $T_a$  values are quite similar, showing that the atmosphere is becoming much more important than in previous flight legs considered.

Examining the vertical moisture differences is a bit more complex. The MODIS vertical moisture difference decreases 0.5 g kg<sup>-1</sup>, while the Everest data decreases 0.6 g kg<sup>-1</sup> over the flight leg. Because the MODIS and Everest  $q_{sfc}$  values decrease by 0.3 and 0.2 g kg<sup>-1</sup>, respectively, while the  $q_a$  value increases by 0.3 g kg<sup>-1</sup>, these values also do not show a clear dependence on either the atmospheric or the surface state. Again, the uniformity of the conditions on this downwind flight leg makes this a reasonable assessment.

#### d. Boundary Layer Structure

Another way to enhance our understanding of air-sea interactions in TNB is to examine changes in the ABL structure between flight days and relating this to changes in the surface fluxes. Understanding the changes that must have occurred in the atmospheric and/or surface

states, which translates to changes in the ABL structure, can improve our knowledge of large variability in energy exchange, and therefore, air-sea interactions. During the 60 hours between the flight on the 23<sup>rd</sup> (red lines and symbols on Figures 28 and 29) and the flight on the 25<sup>th</sup> (black lines and symbols on Figures 28 and 29), the UAV SHF decreases for both flight legs, while the LHF decreases in Leg 1 and increases in Leg 2. The SHF in Leg 1 decreases 87 W m<sup>-2</sup> (Table 6) and the LHF decreases 95 W m<sup>-2</sup> (Table 7) between the two flight days. In Leg 2, the SHF decreases ~173 W m<sup>-2</sup> and the LHF increases ~30 W m<sup>-2</sup> between the two flight days. It is expected that the LHF increase from the 23<sup>rd</sup> to the 25<sup>th</sup> is spurious due to the dry bias discussed in Section 3b, and is not discussed further here. Below we will examine differences in the ABL structure on the 23<sup>rd</sup> compared to the 25<sup>th</sup> to determine if the reduction in the heat fluxes described in this paragraph are consistent with the idealized model of ABL behavior.

As described in Section 1 as an air parcel moves off the continent and over TNB, it carries with it the characteristics of the ABL that developed over the plateau, which in an idealized case, consists of a strong temperature inversion (Heinemann 2008). The entire ABL consists of the inversion, which is within a few hundred meters of the surface (Heinemann 2008; Raddatz et al. 2011). As the air parcel descends through the mountains west of TNB, its characteristics will be modified through both adiabatic warming and mixing as mechanical turbulence increases.

Once this colder continental air moves over the warmer polynya surface, convection will ensue and a convective internal boundary layer (CIBL) will be formed (Chang and Braham 1991; Renfrew and King 2000; Heinemann 2008; Raddatz et al. 2011). Within the CIBL the atmosphere is well mixed, with a temperature somewhere between the cold katabatic layer and the warm surface over the polynya. The top of the CIBL,  $z_c$ , will increase with height as the air parcel moves over the polynya due to the entrainment of air from the free troposphere (Chang and Braham 1991; Hartmann et al. 1997; Renfrew and King 2000; Heinemann 2008). This layer will erode the original katabatic layer that existed on the plateau. Above  $z_c$ , the residual layer from the plateau remains. The CIBL is the new top of the ABL.

Once the air parcel has moved a sufficient distance downstream,  $z_c$  will be the same height as the top of the original katabatic layer from the polar plateau, and the entire residual katabatic layer is eroded. There will be a new inversion present at the top of the ABL that is no longer related to the original katabatic layer but simply exists due to the transition between the colder ABL air and the relatively warmer free atmosphere. This idealized case also does not account for advection, entrainment, or cloudy conditions.

This conceptual model of ABL modification can be used with the data collected over the polynya in 2009 to understand the evolution of the ABL. On 23 September, the ABL matches overall with the idealized case. Potential temperature profiles show a well-mixed layer extending from flight level to approximately 240 m in Profile 1 (Figure 33a). Above this layer is a stable layer that is likely the residual katabatic layer. Profile 2 also shows potential temperatures increasing with height above a well-mixed layer that extends to at least 480 m (Figure 33a). Profile 3 shows the height of the well-mixed layer to be at 540 m. The increase in the ABL top from 240 to 480 to 540 m downstream matches well with the idealized case with  $z_c$  increasing with increasing downwind distance.

On 25 September, the potential temperatures from Profile 1 show a stable layer existing from flight level throughout the depth of the profile (Figure 33b). In Profile 2, a slightly stable layer is shown to extend from flight level to 240 m. Above 240 m the atmosphere is generally stable with conditions nearly identical to those observed in Profile 1. Profile 3 shows a more



Figure 33. Potential temperature profiles from 23, 25, and 18 September.

well-defined mixed layer extending to 250 m (the height of the BL), with a stable layer extending above this level. It is likely that if a mixed layer exists at the location of Profile 1 it is below flight level, but that once the air parcel has moved 20 km downstream to Profile 2, the

well mixed layer is becoming more established. The stability of the ABL is weaker in Profile 2 than Profile 1, indicating that the residual katabatic layer is being eroded. In Profile 3, the atmosphere is exhibiting a well-defined CIBL that has increased in depth from Profiles 2 to 3.

On the  $23^{rd}$ ,  $z_c$  is higher (240-480-540 m; Figure 33a) than on the  $25^{th}$  (below flight level – 240-250 m, Figure 33b), suggesting more active mixing and entrainment on the  $23^{rd}$  compared to the  $25^{th}$ . The well-defined mixed layer extending to 540 m by the end of the flight leg on the  $23^{rd}$  also supports the idea of enhanced convective mixing on the  $23^{rd}$ . These observations of the ABL depth and structure are consistent with the larger SHF on the  $23^{rd}$  compared to the  $25^{th}$  (Tables 6 and 7).

Another example of comparing flight days leading to a furthering of our understanding of air-sea interactions in TNB is comparing the  $18^{th}$  data to that from Leg 1 on the  $23^{rd}$  or  $25^{th}$ . The vertical temperature and moisture differences and wind speeds given in Figures 28 and 29 indicate that the fluxes on the  $18^{th}$  should be smaller than both of the other flight days. This is found for the SHF, but the LHF on the  $18^{th}$  is larger than that on Leg 1 on the  $25^{th}$ .

While the presence of the stable layer is consistent with the surface state, upward heat fluxes estimated from the UAV are inconsistent with this ABL structure. However, it is possible that entrainment of warmer (and drier) air aloft or subsidence warms the ABL downstream, which then results in a positive heat flux being estimated from the UAV profile data. An examination of wind speed profiles (Figure 34) shows an increase in the wind speed between 350 and 457 m, with a sharp decrease in wind speed above 457 m. This increase in wind speed, and the corresponding well-mixed layer at this level, could indicate the mechanical mixing of higher and drier potential air from aloft. It is possible that this could explain the spuriously larger LHF

on the 18<sup>th</sup> as compared to Leg 1 on the 25<sup>th</sup>, although it is unclear why this is not also reflected in the SHF. Further analysis must be conducted to better understand this potential discrepancy.

## 4. Comparison to COARE flux estimates

The following section will compare the UAV fluxes to fluxes estimated from the COARE bulk algorithm (Figures 35-39). These results aid in our understanding of the role of surface and atmospheric conditions in September 2009 in TNB, as well as offer a comparison between fluxes estimated from observational data to those estimated with a bulk flux algorithm. Two COARE fluxes are used – those with MODIS (COARE-M) and Everest (COARE-E)  $T_{sfc}$  as inputs. The



Figure 34. Wind speed profiles for 18 September.



Figure 35. SHF and LHF estimates from COARE-M and UAV-E for Leg 1 on 23 September 2009. The x-axis is the distance along the flight leg. UAV-E estimates are values from Table 5 with the uncertainty incorporated.



Figure 36. SHF and LHF estimates from COARE-M and UAV-E for Leg 2 on 23 September 2009. The x-axis is the distance along the flight leg. UAV-E estimates are values from Table 5 with the uncertainty incorporated.



Figure 37. SHF and LHF estimates from COARE-M and UAV-E for Leg 1 on 25 September 2009. The x-axis is the distance along the flight leg. UAV-E estimates are values from Table 5 with the uncertainty incorporated.





Figure 38. SHF and LHF estimates from COARE-M and UAV-E for Leg 2 on 25 September 2009. The x-axis is the distance along the flight leg. UAV-E estimates are values from Table 5 with the uncertainty incorporated.





Figure 39. SHF and LHF estimates from COARE-M and UAV-E for 18 September 2009. The x-axis is the distance along the flight leg. UAV-E estimates are values from Table 5 with the uncertainty incorporated.

mean values of each flux are plotted with a thicker line, while the uncertainties in the UAV fluxes (found in Tables 6 and 7) and the COARE fluxes are plotted with a thinner line. The uncertainty in the COARE fluxes are estimated by varying the  $T_{sfc}$  data within the range of uncertainty described above (+/-1.3 K for MODIS and +/-2 K for the Everest). The method for determining the uncertainty in the UAV estimated fluxes is described in detail in KC2014.

Across all five flight legs, the UAV and COARE SHF overall are comparable, with the COARE and UAV SHFs within the range of uncertainty on both flight legs on the  $23^{rd}$  (Figures 35 and 36) and  $25^{th}$  (Figures 37 and 38) for at least one of the two COARE calculated SHF values. Whether the COARE-E or COARE-M SHFs are more comparable to the UAV SHF depends on the flight leg. As discussed in Section 2c, it is expected that the Everest data will provide a better estimate of  $T_{sfc}$  in conditions where the surface state is highly variable, such as in Leg 1 or possibly Leg 2 on the  $23^{rd}$  or 25th. In areas such as Leg 3 on the 18th that are expected to have more uniform  $T_{sfc}$  values, the MODIS estimates might be more accurate.

On the  $23^{rd}$  (Figures 35 and 36), only the COARE-M fluxes are available, and show differences of ~50 W m<sup>-2</sup> on Legs 1 and 2, which is an 11% and 24% difference, respectively, from the UAV SHF (Table 6). On the  $25^{th}$  (Figures 37 and 38), the COARE-E SHFs are more comparable than COARE-M to the UAV SHF in both flight legs. In Legs 1 and 2, the COARE-E SHFs differ from the UAV SHF by 75 W m<sup>-2</sup> (20%) and less than 5 W m<sup>-2</sup> (25%). In contrast the COARE-M to UAV SHF differences are 125 W m<sup>-2</sup> (33%) and 65 W m<sup>-2</sup> (325%), for Legs 1 and 2 respectively (Figures 37 and 38, Table 6). On the 18<sup>th</sup> (Figure 39), the COARE-M SHF is more comparable to the UAV SHF than COARE-E, where the COARE-M value differs by 90 W m<sup>-2</sup> (72%) and the COARE-E values differs by 245 W m<sup>-2</sup> (108%) (Table 6). This analysis reflects the discussion in Section 2c, showing that the UAV and COARE-E SHF provide similar

estimates in areas where  $T_{sfc}$  is expected to be more highly variable (Figures 27 and 30), than in areas where  $T_{sfc}$  values are more uniform (Figure 32).

Unlike the SHF, the COARE and UAV LHFs are never within the range of uncertainty of each estimate. Leg 1 on 25 September is the only leg where the COARE-E LHF is more comparable to the UAV LHF than the COARE-M value (60 W m<sup>-2</sup>, 107%, compared to 140 W m<sup>-2</sup>, 212%) (Figure 37 and Table 7). In Leg 2 on the 25<sup>th</sup>, the COARE-M LHF differs from the UAV LHF by 50 W m<sup>-2</sup> (41%) compared to COARE-E (90 W m-2, 73%) (Figure 38, Table 7). On the 18<sup>th</sup>, the COARE-M LHF differs by 55 W m<sup>-2</sup> (52%) compared to the COARE-E fluxes (85 W m<sup>-2</sup>, 81%) (Figure 39, Table 7). On the 23<sup>rd</sup>, where no COARE-E fluxes are available, the COARE-M LHF differs from the UAV LHF in Legs 1 and 2 by ~50 W m<sup>-2</sup> (33% and 45%) (Figures 35 and 36, Table 7). The LHFs might be less comparable than SHF for two reasons: (1) the UAV LHF might have issues due to challenges in achieving accurate relative humidity observations during flights due to a sensor lag issue (Knuth et al. 2013), and (2) the COARE flux is less reliable for the LHF than the SHF (Fairall et al. 2003).

On most flight days, the COARE-E values have a larger range of uncertainty compared to the COARE-M values, which both almost always have a larger range of uncertainty than the UAV estimates (Tables 6 and 7). The COARE-E SHFs have a range of uncertainty of 125 and 85 W m<sup>-2</sup> in Legs 1 and 2 on the 25<sup>th</sup>, compared to the COARE-M SHFs of 90 and 65 W m<sup>-2</sup> (Figures 37 and 38, Table 6). The UAV range of uncertainty is 27 W m<sup>-2</sup> in Leg 1 and 30 W m<sup>-2</sup> in Leg 2 (Table 6). For the LHF, the range of uncertainty in the COARE-E values is 45 and 30 W m<sup>-2</sup> in Legs 1 and 2, and in COARE-M is 45 and 20 W m<sup>-2</sup> (Table 7). This is larger than the 6 and 19 W m<sup>-2</sup> range of uncertainty in the UAV LHFs. On the 18<sup>th</sup> (Figure 39), the range of uncertainty is comparable between the two COARE fluxes (70 W m<sup>-2</sup> for the SHF and ~20 W m<sup>-2</sup>).

<sup>2</sup> for the LHF), which are both smaller than the range of uncertainty in the UAV fluxes (88 W m<sup>-2</sup> for the SHF and ~25 W m<sup>-2</sup> for the LHF) (Table 6). The UAV range of uncertainty on the 18<sup>th</sup> is larger due to issues with the pressure measurements on this day (KC2014). On the 23<sup>rd</sup>, only the COARE-M fluxes are available, which show a larger range of uncertainty than the UAV fluxes (Figures 35 and 36). The range of uncertainty in Legs 1 and 2 for the SHF is 100 and 65 W m<sup>-2</sup> for COARE-M, compared to 40 and 30 W m<sup>-2</sup> for the UAV estimates (Table 6). For the LHF, the range of uncertainty in Legs 1 and 2 for COARE-M is 45 and 20 W m<sup>-2</sup>, compared to 9 W m<sup>-2</sup> for the UAV estimates (Table 7).

There are two key points when considering the comparison of COARE to the UAV estimated fluxes. The first is that there is a large amount of uncertainty in COARE compared to the UAV fluxes due to the large amount of uncertainty in the  $T_{sfc}$ . This comparison has shown that varying the  $T_{sfc}$  by up to 2 K can provide uncertainties in the COARE fluxes of 125 W m<sup>-2</sup> in the SHF and 45 W m<sup>-2</sup> in the LHF (Figures 35-39). These fluxes are therefore much less robust than the UAV fluxes without having more accurate  $T_{sfc}$  observations. In KC2014, the uncertainty in the UAV estimates is, at most, 88 W m<sup>-2</sup> for the SHF and 24 W m<sup>-2</sup> for the LHF.

Second, the average wind speeds for the 2009 TNB flights are at the upper end of COARE's tested accuracy (Fairall et al. 2003) (Table 8). It is therefore possible the COARE estimates might differ from the UAV estimates for this reason. The COARE algorithm might also provide more comparable estimates by including data from this work to adjust input coefficients at higher wind speeds.

#### 5. Summary

In September 2009, UAVs were flown over a polynya located in TNB that is present due to strong, cold, and dry winds that advect sea ice away from the coast. The UAVs sampled the atmosphere at various locations downwind to measure the evolution of the ABL as it passed over the polynya. KC2014 used this data to estimate SHFs and LHFs based on changes in the ABL as the air mass passed over the polynya.

This study examined the atmospheric and surface conditions during three flights in September 2009 to understand their role in controlling temporal and spatial variations in heat fluxes. It was determined that surface conditions play a primary role during most of these flights, as greater variability in the surface state compared to the atmosphere was present on all flight days but one. On 18 September, variations in the atmosphere were shown to be comparably important to the surface conditions. The atmospheric variables, such as temperature and mixing ratio, were generally far less variable compared to the T<sub>stc</sub> and saturation mixing ratios (with the exception of the 18<sup>th</sup>). In areas with more open water present, fluxes were greater than in areas with greater sea ice concentrations, which is consistent with expectations. Additionally, the atmospheric boundary layer temperature structure was used to assess, from a qualitative standpoint, the heat flux input at the ocean-atmosphere interface. Days with a well formed mixed layer coincided with days that experienced higher heat fluxes from the surface

The UAV estimated fluxes, calculated using only mean atmospheric state data, from KC2014 were compared to those estimated using the COARE bulk flux algorithm (Fairall et al. 2003) with UAV observations of the atmospheric and surface conditions. Overall, the COARE algorithm showed larger uncertainty compared to the UAV estimates due to uncertainty in the

surface input data.  $T_{sfc}$  data from both the MODIS satellite and an Everest infrared thermometer on board the UAVs showed variations from the UAV SHF of more than 325% and 212% for the LHF. This study shows that accurate measurements of surface conditions are crucial in understanding heat exchange in TNB and that the method presented by KC2014 of using mean state atmospheric data over the depth of the boundary layer provides a more accurate assessment of surface turbulent fluxes than bulk flux algorithms for the 2009 TNB data.

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#### **Chapter 5: Conclusions**

Using data collected from a series of UAV flights over the TNB region of Antarctica, this dissertation explores air-sea interactions in the ABL during the late austral winter month of September 2009. TNB is a region of interest in the Antarctic, as cold, dry air originating from the continental interior is modified by a relatively warm and moist surface present due to open water. The open water occurs due to a recurring polynya that forms from strong offshore winds in the region. The modification of the continental air mass by the underlying warm, moist surface leads to great implications on the atmosphere, ocean, and cryosphere due to air-sea heat exchange, the formation of high-density shelf waters that become part of Antarctic Bottom Water, and due to large amounts of sea ice production. The UAV flights provided the first ever winter measurements of the ABL in TNB, and offered invaluable information regarding the atmospheric structure over the polynya. This study used these measurements to estimate heat exchange between the ocean and atmosphere, and to understand the primary forcing mechanisms of this heat exchange and the ABL response.

Chapter 2 provides a comprehensive study of the typical atmospheric conditions of TNB from a 15-year analysis. The purpose of this study was to put the observations from the field campaign into context to understand whether September 2009 was a typical or anomalous year. Wind and temperature data collected from Rita AWS, located just west of TNB, was examined from 1993-2007 to determine dominant weather conditions in the region. As well, an analysis of cyclone density, size, and shape in the Ross Sea from NOAA LAC (1 km) resolution satellite data was conducted, as these systems also have great impacts on the region. This study found that 55% of winds were west to northwesterly, with speeds less than 20 m s<sup>-1</sup>. Strong wind

events were categorized based on those that had wind speeds greater than 20 m s<sup>-1</sup> for at least 10 hours. 83% of the strong wind events were from directions consistent with downslope flow, with the majority of these events (68%) occurring during the winter months. Temperatures were influenced by wind fluctuations, with most (39%) between -15°C and -25°C. The cyclone climatology revealed most cyclones were less than 1000 km in size, with a high density in the southwestern Ross Sea (including TNB).

When comparing the analysis from 1993-2007 to AWS and satellite data in September 2009, an increased number of strong winds and an eastward shift in cyclone activity indicated that September 2009 was an anomalous year. Further analysis revealed a shift in the upper level trough in September 2009 due to changes in the Southern Hemisphere Rossby wave pattern. The increased intensity of the 500 mb trough enhanced downslope flow over TNB, and its eastward shift contributed to a corresponding shift in the cyclone activity. The results from this study showed that air-sea interactions during September 2009 might be uncharacteristic of typical years.

The atmospheric conditions in TNB, combined with the unique surface conditions of the polynya, lead to strong air-sea interactions in the region. The results from the contextual study showed that air-sea exchange during 2009 would be part of an anomalous weather pattern. To understand the response of the atmosphere to the energy input from the ocean, heat fluxes were estimated in the region. The atmospheric data collected by the UAVs were used to develop an innovative methodology to estimate heat exchange in TNB that did not require the use of bulk flux algorithms that can have large uncertainties. This method estimated the heat and moisture content within the ABL based on profile measurements that were collected from the UAV flights. The heat fluxes were estimated based on how much the heat or moisture content changed

between pairs of profiles collected in the downwind direction away from the coast. The underlying polynya was assumed to have the greatest impact on modifying atmospheric properties.

This study initially estimated fluxes that were "uncorrected", in that all heat content changes measured were assumed to be diabatic and due to surface turbulent fluxes, and that the UAV followed the background airflow perfectly. Subsequent modifications to these uncorrected fluxes were made to account for adiabatic (for SHF) and non-Lagrangian (SHF and LHF) processes, as well as corrections to account for the heat content of the air below the minimum UAV flight level. This method implemented a spatial technique for correcting the fluxes where two different, but scientifically appropriate, versions of this method were tested. These two versions provided flux estimates within 5 W  $m^{-2}$ , showing the robustness of this methodology. Examining the impact of instrument inaccuracy on the resulting SHF and LHF tested further uncertainties of the method, with variability in the results of less than 15 W m<sup>-2</sup> (in most cases). The estimates of the heat fluxes, including corrections accounting for adiabatic processes, non-Lagrangian processes, and the heat content below the UAV minimum flight level, showed SHFs ranging from 11 to 373 W m<sup>-2</sup>, while the LHF ranged from 19 to 121 W m<sup>-2</sup>. These estimates were from 18, 23, and 25 September 2009 and were chosen because their flight missions were designed to measure heat and moisture content changes. These estimates are generally lower than previous studies discussed in Chapter 1 that assumed largely open water conditions.

Once the heat fluxes were quantified over the region, an understanding of the primary forcing mechanisms causing temporal and spatial variability in the heat fluxes was necessary. A thorough examination of the surface and atmospheric states from UAV and satellite data was conducted in Chapter 4 to determine how variability in each leads to variability in the estimated

fluxes. The sensitivity of the heat flux estimates to the surface state was also determined using the TOGA-COARE bulk flux algorithm.

This study showed that the fluxes estimated for September 2009 varied more due to changes in the surface state than to changes in the atmosphere. This was primarily due to the larger variability in surface conditions compared to the atmosphere. Surface temperatures typically varied 15 K from the coast to the end of the downstream UAV flight path (~40 km), while over this same distance atmospheric temperatures typically only changed ~2-4 K. Similarly, surface saturation mixing ratios changed ~1.5 g kg<sup>-1</sup> along the flight path, with an atmospheric change of only 0.4 g kg<sup>-1</sup>. Fluxes were typically higher over areas with more open water, consistent with expectations.

Using the TOGA COARE bulk flux algorithm, the sensitivity of the fluxes to the underlying surface state was quantified. Two sets of surface temperatures were used as input data to the algorithm – one from MODIS satellite data, and one from Everest infrared thermometers mounted on board the planes. The uncertainty in the TOGA COARE fluxes was over 150 W m<sup>-2</sup> for the LHF and 400 W m<sup>-2</sup> for the SHF. Much of this variability was due to the uncertainty in the surface temperature measurements.

The atmospheric response to air-sea interactions over TNB was explored by examining the structure of temperature profiles over the region. It was shown that a convective boundary layer forms near the surface when the cold, dry continental air passes over the relatively warm, moist polynya. The convective layer erodes the continental layer, and increases with depth as the fetch increases.

The purpose of this dissertation is to examine air-sea interactions based on aircraft measurements over TNB. The results from this work have enhanced our knowledge of these interactions, and the ABL response, over this important area of the Earth's climate system. The estimates of SHF and LHF provided as part of this work represent the first fluxes calculated from in situ data over TNB during the winter months. Prior to this work, all estimates of fluxes during the winter months were estimated based on model simulation. These previous estimates are consistently larger than the fluxes estimated from the in situ UAV data. Additionally, the methodology used to estimate these fluxes has widespread applicability to other regions where surface data might not be present. Finally, the understanding of the forcing mechanisms of heat exchange in TNB allows a better understanding of how this heat can influence other areas of the atmosphere, such as cyclone development or precipitation formation. Carrasco et al. (2003) has shown the TNB region to be a region of strong mesocyclone activity, which can have implications on the local and regional atmosphere. The results from examining the response of the ABL during cold air outbreaks in TNB can be applied to other areas of the globe. Changes in the atmosphere can lead to changes in ocean currents, which can have implications on climate and weather across the globe.

There are several areas where the work completed as part of this dissertation can be expanded in the future. With the lessons learned from the data collected from the 2009 flights and the studies conducted in Chapters 3 and 4, a second set of flights to measure the three dimensional atmospheric boundary layer were conducted in September 2012. One of the largest improvements of the 2012 data from the 2009 data is that Eulerian measurements were collected along the flight path to also understand changes over time. The exact flight path over the polynya was flown twice approximately one hour apart to understand how quickly the boundary layer evolves.

These flights collected measurements over the region that will allow some of the limitations highlighted in Chapter 2 to be better understood. With the data collected during September 2012, a temporal methodology can also be used to estimate the fluxes. In Chapter 2, heat fluxes estimations are corrected to account for non-Lagrangian processes, wherein the difference between the UAV and the air parcel following the background flow as both move downstream from the coast are corrected by accounting for the difference in space. The UAV and air parcel travel at different speeds, and this correction accounts for their difference in space after a certain amount of time has passed. A variation on this method incorporating Eulerian techniques can be used in which the difference in time between when it takes the air parcel and the UAV to traverse a certain distance can also be used to correct the fluxes. This version was broadly tested as part of the work presented in Chapter 3, but a lack of accurate and available data prevented a full comparison to the spatial methodology presented as part of this work. It is anticipated that adding corrections to account for time as well as space will further prove the robustness of this methodology.

Additionally, data collected from the 2012 flights (and some data from 2009) will be used to estimate basic characteristics of the continental air passing over the polynya, such as the horizontal dimensions of the downslope wind jet. The ABL structure and heat flux forcing mechanisms during these flights will also be examined and compared to the 2009 data. The 2012 data will also be compared to the idealized structure outlined in Chapter 4 to determine whether the boundary layer follows this pattern. As well, the atmospheric and surface data will be compared to determine whether surface conditions are also the primary forcing mechanism in 2012. Preliminary work has been completed to compare the fluxes estimated from the UAV methodology and from TOGA-COARE to that from model data. Comparisons to the flux algorithm as part of the Antarctic Mesoscale Prediction System (AMPS) have showed mixed results, varying from the UAV fluxes by ~250 W m<sup>-2</sup>. Further work needs to be conducted to better evaluate the AMPS model forecasts. The flux algorithm code used to estimate the AMPS values will be compared to the UAV estimates and the COARE data in a similar manner as described in Chapter 4. As well, a comparison to reanalysis data will be conducted. Preliminary analysis shows the reanalysis data underestimates the UAV data by 400-500 W m<sup>-2</sup>. Further study on this will also be conducted in a manner similar to that of Chapter 4.

Finally, understanding what, if any connection exists between the heat fluxes estimated over TNB and synoptic patterns is underway. The work discussed in Chapter 2 showed that 100% of the strong wind events studied occurred when a synoptic system was present in the Ross Sea. Closer examination of the three September 2009 flights studied as part of Chapters 3 and 4, however, show that synoptic systems in the Ross Sea do not significantly impact wind, pressure, or temperature patterns in TNB, and that most patterns appear to be largely influenced by local factors. Variations in wind speed, pressure gradient, and temperature patterns will be studied in TNB and compared to the location of synoptic systems in the Ross Sea. As well, large-scale weather patterns on the polar plateau will be used to understand the impacts in TNB. The katabatic forcing term (Parish and Cassano 2001) will be quantified to determine the role of its impact on flow in TNB compared to synoptic forcing. The inversion strength on the polar plateau will also be studied using AWS or satellite data to understand the strength of the katabatic flow. Changes in these patterns can then be used to understand variability in the air-sea interactions over TNB.

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# Appendix: Unmanned Aircraft System Measurements of the Atmospheric Boundary Layer over Terra Nova Bay, Antarctica

**ABSTRACT.** In September 2009, a series of long-range unmanned aircraft system (UAS) flights collected basic atmospheric data over the Terra Nova Bay polynya in Antarctica. Air temperature, wind, pressure, relative humidity, radiation, skin temperature, GPS, and operational aircraft data were collected and quality controlled for scientific use. The data has been submitted to the United States Antarctic Program Data Coordination Center (USAP-DCC) for free access (doi: 10.1594/USAP/0739464).

# 1. Introduction

Within the Terra Nova Bay (TNB) region of Antarctica, strong air-sea interactions occur during the winter months due to the presence of a persistent latent heat polynya generated largely due to strong downslope winds originating from the interior of the continent (Kurtz and Bromwich 1983, Ciappa et al. 2012). These air-sea interactions are strongest during the winter months as the downslope and katabatic winds bring cold, dry air in contact with the relatively warmer and moister air overlying the polynya. Typical ice conditions in the region consist of open water immediately offshore, with areas of pancake and pack ice further away from the coast (Figure 40). However, ice conditions within TNB can vary greatly depending the strength of the prevailing winds, where calm winds can result in pack ice being present up to the coastal edge. Understanding the impact of the interaction between the cold, dry continental air with the open water of the polynya on atmospheric storms, energy transfer to and from the atmosphere, and the



Figure 40. One kilometer Aqua True Color satellite image from 24 September 2009. The Drygalski Ice Tongue, located in the center of the image, extends approximately 50 km from the coastal edge into the sea. North is at the top of the image. Data collected from the National Aeronautics and Space Administration (NASA)'s website.

changing properties of oceanic water are important for furthering our knowledge of both local and large-scale meteorological, glaciological, and oceanographic changes.

In September 2009, a series of unmanned aircraft system (UAS) flights were undertaken over the TNB polynya to measure changes in the atmospheric layer overlying the polynya (Cassano et al. 2010). The UAS used for these flights were built and designed by Aerosonde®, and collected atmospheric measurements (air temperature, skin temperature, u- and vcomponents of the wind, atmospheric pressure, relative humidity, and shortwave and longwave radiation), geo-reference data (latitude, longitude, altitude), and aircraft specific information (roll, pitch, yaw, aircraft height, and ground speed). A Canon SX110 IS digital camera was mounted on the UAS to collect images of the sea surface and ice state in the polynya. The data were recorded on board the aircraft using data loggers and memory cards and also telemetered back to the field team in real-time (except for digital photographs) during the flights using Iridium satellite communication (when over the horizon) or 900 MHz radio (when within line of sight).

The purpose of these flights was to document the downstream evolution of the boundary layer over Terra Nova Bay and to study air-sea interactions in the winter months over the Terra Nova Bay polynya (Figures 40 and 41). The meteorological data collected by the UAS are sufficient to allow us to document the downstream evolution of the temperature, humidity, winds, and the boundary layer over the polynya. The data collected are also sufficient to allow us to estimate the turbulent sensible heat, latent heat, and momentum fluxes as well as to estimate all of the terms in the horizontal momentum equation. The method for estimating these terms from the UAS data will be described in separate papers that are currently in preparation. Turbulent flux instrumentation was not carried during the flights due to weight restrictions and to keep the instrumentation simple and robust.

Various data loggers on the UAS collected the data with varying sampling rates. As part of the quality control process, the data was rigorously checked for false or spurious data as well as streamlined into cohesive sampling rates. The final dataset is provided to the general public in two different formats at a sampling rate of ten seconds. A description of the UAS and flights, the quality control process, and the publically available dataset is provided below.



Figure 41. Map of the entire UAS flight region, from Pegasus Runway to Terra Nova Bay.

# 2. UAS Description and Flights

During September 2009, the Aerosonde® UAS flew sixteen flights – eight flying north from the Pegasus ice runway near McMurdo Station to TNB (Figure 42), six flights flown near Pegasus (Figure 43), and two unsuccessful flight missions (Figure 44). The six flights near Pegasus were largely test flights used to ensure proper aircraft instrument operations, while the eight flights to TNB were science flights (Table 9).

The following describes the standard flight pattern for the science flights to TNB. During the flights to TNB, the aircraft followed the Transantarctic coast north to TNB, and telemetered data back to the field team at Pegasus Runway via Iridium satellite. Once the aircraft did a complete south to north transect of TNB near the coast, the mission scientist used the telemetered meteorological data to locate the position of strongest winds across TNB. Based on this, a revised set of waypoints and altitudes were then uploaded in real time to the Aerosonde's autopilot, and the UAS then flew parallel to the strongest winds (as determined from the telemetered data stream) across TNB (Figure 45). Finding the area of strongest winds was important for meeting the scientific goals of the project, which included estimating the largest air-sea fluxes over the polynya. As air-sea fluxes are (in part) controlled by wind strength, flights within the area of strongest winds provided the data to calculate the largest fluxes within TNB. This flight strategy also documented the downstream evolution of the fluxes when moving from stronger to weaker winds within the jet.

Along this flight path the UAS performed soundings of the atmosphere at several locations. Figure 46 shows examples of three specific humidity profiles taken along the flight path measured by a UAS on 23 September. Each of these profiles consisted of a spiral ascent, a



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d)

Figure 42a-d. Flight path for the eight science flights from Pegasus Runway to Terra Nova Bay. Gaps in the flight path indicate a lack of data. Profile locations are marked with a black dot. Flight altitude is indicated by the color scale shown in each panel.





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b)





c)

Figure 43a-c. Flight path for the six local flights near Pegasus Runway. Gaps in the flight path indicate a lack of data. Profile locations are marked with a black dot. Flight altitude is indicated by the color scale shown in each panel.



Figure 44. Flight paths for the two flights that did not reach Terra Nova Bay. Gaps in the flight path indicate a lack of data. Profile locations are marked with a black dot. Flight altitude is indicated by the color scale shown in each panel.

Science Flights		Local/Unsuccessful TNB Flights		
Start – End Times	UAS Number	Start – End Times	UAS Number	
Start: 14 Sept. 04:37:28	214	Start: 7 Sept. 00:14:38	217	
End: 14 Sept. 19:39:45		End: 7 Sept. 01:04:08		
Start: 18 Sept. 03:02:54	214	Start: 8 Sept. 21:02:20	217	
End: 18 Sept. 19:39:00		End: 9 Sept. 03:47:30		
Start: 21 Sept. 18:51:54	214	Start: 9 Sept. 00:56:57	214	
End: 22 Sept. 05:51:32		End: 9 Sept. 05:29:31		
Start: 23 Sept. 19:09:10	215	Start: 10 Sept. 05:58:57	214	
End: 24 Sept. 06:24:48		End: 10 Sept. 06:26:07		
Start: 23 Sept. 20:08:09	216	Start: 12 Sept. 04:02:24	215	
End: 24 Sept. 06:13:28		End: 12 Sept. 04:52:24		
Start: 25 Sept. 06:50:11	215	Start: 13 Sept. 04:38:49	214	
End: 25 Sept. 19:41:38		End: 13 Sept. 21:44:35		
Start: 26 Sept. 18:33:46	214	Start: 16 Sept. 03:20:19	215	
End: 27 Sept. 04:35:04		End: 16 Sept. 06:30:39		
Start: 26 Sept. 19:23:40	216	Start: 22 Sept. 04:52:04	216	
End: 27 Sept. 04:19:51		End: 22 Sept. 06:18:34		

# Table 9. Start and end times (in UTC) and the UAS aircraft number for science flights to TNB and local flights near Pegasus ice runway.

Spiral descent, or both. The top altitude of each spiral ascent or descent pattern varied by flight but was chosen to ensure that the entire depth of the boundary layer was sampled. The local flights near Pegasus Field had no regular and consistent pattern between flights (Figure 43).

The Aerosonde® UAS (Maslanik et al., 2002; Curry et al., 2004; Inoue and Curry, 2004;

Inoue et al. 2008) are designed and built in Australia for commercial use, and for these flights



Figure 45. Winds observed during the 23 September 2009 UAS flight over Terra Nova Bay.

Required a field team of four to fly. Extensive, previous use in the Arctic demonstrated the potential of the Aerosonde system for polar operations (Curry et al., 2004). The UAS used for the TNB missions in 2009 had a wingspan of 3 meters, a weight of 15 kg, and a payload capacity of 2-5 kg. The aircraft had a range of over 1000 km, and an endurance of 18 hours, and were flown between approximately 150 and 3000 m altitude during the flights. Typically, the downwind flight transects were flown at an approximately constant altitude between 150 and 250 m. Figures 42 to 44 also now indicate the flight altitude along the entire flight path for each UAS flight. The UAS were not flown at a lower altitude due to concerns about the presence of

tall icebergs in the area as well as inaccuracies in the GPS altitude data. The planes were manually pilot controlled at take-off and landing and operated in full autonomous mode during



23 September 2009 UAV #216 Specific Humidity Profiles

Figure 46. Specific humidity profiles from the 23 September 2009 (UAS #216) flight. Profile 1 was measured at approximately 1:35 UTC, Profile 2 was measured at approximately 1:52 UTC, and Profile 3 was measured at approximately 2:18 UTC on 24 September 2009.

Most of the flight, with real-time telemetry to the ground station and pilot in command. The pilots had the ability to take over controls or alter the flight plan at any time.

The instruments on board the UAS include both the instruments internal to the UAS as well as additional meteorological instruments added specifically for the 2009 flights. The internal UAS flight management system, including the autopilot, GPS navigational system, flight sensors, communications, and payload interfaces were Piccolo components developed by Cloud Cap Technologies, Inc. Data from the Piccolo system were available on all flights, and were telemetered back in real time. The meteorological instruments added to all of the UAS included pressure data (Vaisala PTB110), skin temperature (Everest Interscience, Inc. Everest 3800ZL), air temperature and relative humidity (Vaisala HMM213), and a nadir-viewing digital camera (Canon Powershot SX110 IS). The u and v components of the wind were calculated during aircraft maneuvers. The autopilot navigation system contains a Kalman filter, which estimates the wind vector continuously during the flight. Periodic 'wind finding' maneuvers are performed during the flight to improve the wind estimates which otherwise degrade in accuracy over time whilst flying a constant heading, as the aircraft heading is not directly observable without magnetometer input. It is important to note that the response time of the relative humidity sensor is likely slower than stated due to the cold temperatures in the region and this may present some issues when interpreting the data collected during the quick ascent/descent of the profiles (Table 10). The use of this data in the soundings should, as such, be used with caution.

One UAS was also outfitted with a Kipp and Zonen CNR2 radiometer to measure shortwave and longwave radiation. An additional GPS, a Canmore GT-730F, was also added to several UAS. Finally, a laser altimeter was used to measure sea ice freeboard and wave state. This laser altimeter, part of the CU LIDAR Profiling and Imaging System (CULPIS), was designed by the University of Colorado (Crocker et al. 2012). The CULPIS profilometer was designed specifically to collect fine-resolution surface elevation measurements from small UAS in polar environments (Crocker et al. 2012). The system consists of a single-beam near infrared LiDAR sensor that measures the distance from the aircraft to the ground surface at 400 Hz, a L1 GPS unit that detects the aircraft's altitude and location at 10 Hz, and an inertial measurement

Instrument	Accuracy	Response Time	Operating Temperature	
			Range	
Piccolo Autopilot			-40 to +80°C	
Vaisala PTB110	+/- 1.5 hPa (at +20°C)	500 ms (at +20°C)	-40 to +60°C	
Everest 3800ZL	+/- 0.5°C	0.25 s	-40 to +100°C	
Vaisala HMM213	+/- 2-3% (RH), +0.1°C	60 s (at +20°C)	-70 to +180°C	
	(Temperature) (at +20°C)			
Kipp and Zonen CNR2	10-20 µV/W/m <sup>2</sup>	<10 s	-40 to +80°C	
Canmore GT-730F	5 m	1s	-40 to +85°C	
CULPIS	1 m globally		-15 to 50°C	

Table 10. Instrument specifications for the equipment carried on board the UAS.

The extreme conditions during the end of the Antarctic winter led to aircraft issues on several flights. Aircraft #217, flown on 8 September, crashed on the sea ice near 76.62°S 165.89°E on its return flight to Pegasus Runway. It is suspected that a fuel pump failure caused the engine to fail and the plane to crash. On 10 September, aircraft #215 indicated a generator belt failure shortly after take off. The Aerosonde® pilots were able to make an emergency landing at Pegasus Runway without incident. On 13 September, aircraft #214 lost Iridium communications with the ground team at Pegasus Runway. After a half hour of lost communications, the aircraft returned to Pegasus where 900 MHz radio communications were

reestablished, and flew over the runway for over 12 hours until daylight permitted safe landing by personnel.

A variety of conditions were sampled during the sixteen flights to TNB or near Pegasus Runway. A summary of these conditions (maximum, minimum, and mean temperature, maximum and mean wind speed, mean wind direction, and maximum flight altitude) is listed in Table 11. Figure 47 shows the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis mean sea level pressure for the six days flights were flown to TNB.

# 3. Data Description

There were five data streams in which the data were collected from the UAS – four logger data streams, and one telemetered (Table 12). The telemetered data stream was sent back to the field team in real-time to monitor the aircraft and the weather (specifically temperature, wind speed and direction, and relative humidity) in which it was flying. The data sent back in real-time consisted of latitude, longitude, altitude, ground speed and direction, roll, pitch, and yaw data, air pressure, air temperature, relative humidity, u- and v-components of the wind, skin temperature, longwave and shortwave radiation, and other data relevant to monitor the aircraft (such as box temperature or instrument voltages). The latitude, longitude, altitude, ground speed and direction, roll, pitch, yaw, air pressure, and u- and v-components of the wind were sourced from the Piccolo avionics. The remaining data was typically the same as that logged onboard the aircraft, but telemetered at a different sampling rate (generally lower) than was logged. The

Date and UAS number	Temperature (°C)		Wind Speed (ms <sup>-1</sup> )		Wind Direction (°)	Altitude (m)	
	Max	Min	Mean	Max	Mean	Mean	Max
7, #217	-19	-26.4	-21.48	12.82	6.63	291.33	712.17
8, #217	-18.6	-34	-22.16	12.01	4.44	244.26	753.84
9, #214	-21.3	-35.1	-24.83	16.48	2.83	197.81	1355
10, #214	-28	-37	-28.85	8.77	3.93	176.43	587.03
12, #215	-13.3	-26.2	-14.91	18.35	6.13	223.87	577.95
13, #214	-4.3	-29.17	-10.95	14.84	4.87	210.49	654.95
14, #214	-12.97	-34.43	-19.07	28.29	8.19	176	1554.03
16, #215	-17.6	-33.1	-25.15	12.61	5.07	137.43	3054.16
18, #214	-15.94	-28.54	-20.86	25.65	10.39	202.7	1567.35
21, #214	-21.38	-29.76	-24.31	24.98	7.81	200.85	1560.43
22, #216	-24.36	-27.74	-26.27	9.19	4.65	85.84	985.12
23, #215	-21	-35.35	-24.23	26.56	6.63	266.03	2557.18
23, #216	-21.37	-36.62	-23.96	25.98	6.38	267.91	1559.15
25, #215	-15.53	-33.38	-19.08	27.97	6.06	232.08	2056.87
26, #214	-17.39	-34.28	-23.24	34.84	10.1	231.03	1833.5
26, #216	-17.7	-34.2	-23.79	39.89	8.85	214.5	2567.25

one exception is the u- and v-components of the wind, which were only telemetered and not logged by the on-board data logger.

Table 11. Maximum, minimum, and average values for various meteorological variables during each of the sixteen UAS flights. The first column lists the day of September 2009 for the start of each flight and the UAS number.



Figure 47. NCEP/NCAR reanalysis mean sea level pressure data for the days on which flights were flown to TNB. Image provided by the NOAA/ESRL Physical Sciences Division, Boulder Colorado (Kalnay et al. 1996).

Data Stream	Measurements			
	Latitude	Air Temperature		
	Longitude	Air Pressure		
	Altitude	Relative Humidity		
Telemetered (0.33-1 Hz)	Ground Speed	Longwave Radiation		
	Ground Direction	Shortwave Radiation		
	Roll	Skin Temperature		
	Pitch	u-Component of Wind		
	Yaw	v-Component of Wind		
GPS (1 Hz)	Latitude	Altitude		
	Longitude	Ground Speed		
ADC (10 Hz)	Longwave Radiation	Air Pressure		
	Shortwave Radiation	Skin Temperature		
RS232 (0.10 Hz)	Air Temperature	Relative Humidity		
CULPIS	Laser Altimeter			

Table 12. Data available in the four logger and one telemetered data streams.Sampling rates are also given.

The final four data streams were from various logger systems on the UAS, with the first of these from the Canmore GPS. The data within this stream consisted of latitude, longitude, altitude, and ground speed, and was only available on certain flights (Table 13). The second data stream was the logged analog-to-digital converter (ADC) data that consisted of longwave and shortwave radiation, static air pressure, and skin temperature. The third logged data stream was from the RS232 logger interface consisting of temperature and relative humidity data. The final data stream consisted of the laser altimeter data from the CULPIS instrument. Surface elevation measurements were derived from the LiDAR range data, the GPS position data, and the IMU

Date and UAS	RS232	GPS	ADC	Telemetered
number				
7 Sept., #217				Х
8 Sept., #217				Х
9 Sept., #214			Х	Х
10 Sept., #214				Х
12 Sept., #215				Х
13 Sept., #214	Х			Х
14 Sept., #214	Х	Х	Х	Х
16 Sept., #215				Х
18 Sept., #214	Х	Х	Х	Х
21 Sept., #214	Х	Х	Х	Х
22 Sept., #216	Х	Х	Х	Х
23 Sept., #215	Х		Х	Х
23 Sept., #216	Х	Х	X	Х
25 Sept., #215	Х		X	Х
26 Sept., #214	Х	Х	X	Х
26 Sept., #216		Х	X	Х

pointing data during post-processing, although these data have not undergone full post-processing.

Table 13. Available data for each of the sixteen September 2009 flights.

The data logged onboard and telemetered to the field team were collected at varying sampling rates. The telemetered data was sent back at a sampling rate of 1 Hz when within line of site using the 900 MHz radio link, and at 0.33 Hz when using Iridium connectivity. The logged data from the RS232 was collected at a rate of 0.10 Hz, while the ADC logged data was

collected at 10 Hz. The GPS data was logged at a rate of 1 Hz. When the data in the logger file is at a lower sampling rate than the rate at which the telemetered data is provided, the data in the telemetered file is repeated. When the data in the logger file is at a higher sampling rate than the telemetered data, not all of the logger data is included in the telemetered file.

Data was logged and telemetered during pre-flight operations as well as the actual flight of the aircraft. Using the ground speed and altitude data to determine when the UAS was airborne, the five data streams were concatenated to only include data during the actual flight time.

# 4. Quality Control

Quality control (QC) was performed on the data in four of the five data streams as well as synced for similar sampling rates. Spurious data points, unit changes, and other corrections needed to be made in each data set. The CULPIS data provided is the raw LIDAR, GPS, and IMU measurements, and post-processing and QC has not been performed on this data. A summary of the QC process for all other data is provided below.

#### a. RS232 Data (Temperature and Relative Humidity)

The data within the RS232 data stream was not time stamped when logged, and extra processing was needed to ensure proper accuracy when adding time markers to this data. To accomplish this, the RS232 logged data was compared to the telemetered temperature and relative humidity to determine a matching point between the two files. Because the RS232

logger had a step counter, once an initial matching point was determined, subsequent time markers could be added. This application of time markers could yield an inaccuracy of up to 1-2 seconds. This is because the telemetered data is occasionally repeated, and when searching for the initial time marker, it can be difficult to determine which value in the telemetered file is the true initial start time.

#### b. ADC Data (Air Pressure, Longwave and Shortwave Radiation, and Skin Temperature)

The data within the ADC logger file was also not time stamped, but also did have a step counter. However, there were no common data values between the telemetered and ADC logger data to set an initial time stamp, as was true with the RS232 data. In order to set an appropriate time stamp to the ADC logged data, a manual record of when the ADC logger was switched on was used to time stamp the initial value reported in the ADC logged data stream.

The unavailability of comparable data within the two data streams existed for several reasons. First, when the skin temperature and longwave and shortwave radiation voltages from their respective instruments were reported through the telemetered and logger process, the data was compared to two different reference values and calibrated, and thus the final reported value was slightly different for each. Secondly, there were two types of pressure data – the air pressure provided by the Piccolo avionics package, and the air pressure provided from the Vaisala meteorological sensor. Both values were telemetered, but the Vaisala pressure that was telemetered needed to undergo a conversion process, and thus a direct comparison between the two could not be made.

The Everest skin temperature was also calibrated and corrected via post processing steps due to issues discovered during flight operations. It was determined that when the infrared probe was in a cold environment and sensing a relatively warmer environment (such as when flying above open ocean) the thermometer would report an incorrect value (i.e., instrument housing temperature affected the estimated radiometric temperature). This problem was identified through ground calibration tests done on site. The post-processing corrections were carried out by determining the sensor readings that corresponded to a known low temperature target and a high temperature target. The high temperature target was represented by open water areas during flight, as determined using the Canon aerial photographs. These were assumed to be at a sea-water freezing point of -1.8°C. Low temperatures given by near-coincident MODIS satellite data. A linear fit between voltages associated with the low and high temperature targets was then used to convert all voltages to temperatures. This process was done for each flight individually. The estimated error associated with this ad-hoc calibration is about 2 degrees.

Longwave and shortwave radiation data was only available on the flights on 23 September #216 and 26 September #216. This data has not been quality controlled or corrected for aircraft orientation. It is suggested that users carefully evaluate this data prior to use.

# c. GPS Data (Latitude, Longitude, Altitude, Ground Speed)

The Canmore GPS altitude data was reported as height above the WGS84 ellipsoid rather than altitude above the surface. Consequently, the ellipsoid height needed to be converted to topographic altitude by correcting with the EGM96 datum geoid height, which was found at each latitude and longitude location along the flight path. This correction does not provide the exact height above the surface due to the variable influence of ocean dynamic topography, tides, and atmospheric pressure loading, but it does provide the aircraft mean sea level (MSL) altitude.

#### d. Telemetered Data

Very little processing was required of the telemetered data. The altitude reported as part of the Piccolo system was also the ellipsoid height, and also needed to be corrected by the geoid height to obtain the true altitude above mean sea level. As well, the u- and v-components of the wind provided from the Piccolo were converted into wind speed and direction. The telemetered data was repeated at times such that several different time stamps might have the same data over a given time span. In these cases, the first repeated data and time stamp was kept.

Additionally, there are a few instances where one time stamp might report two different values of data – for example, a single time stamp might have two different temperature values. This seemingly repeating time data is actually not repeating data at all, but rather an issue that occurred when the data was telemetered back in real time. Sometimes there would be delays in the transmission of telemetered data that would cause several data packets to be sent at once. Typically when there is repeating data there is a data gap that occurs adjacent to the data. However, because the correct time of the repeated data cannot be accurately determined, the data in the file has been left as is.

# 5. Data Availability

The data from the September 2009 Antarctica flights has been submitted to the United States Antarctic Program Data Coordination Center (USAP-DCC) (doi: 10.1594/USAP/0739464). The available data is in text and csv formats, and is provided at tensecond intervals, representing the lowest sampling rate of the logged data, with an accuracy to within five seconds (Table 14). The data from the flights provided in these files are: date and time, latitude, longitude, altitude, temperature, relative humidity, pressure, wind speed, wind direction, u and v components of the wind, longwave and shortwave radiation, skin temperature, ground speed and direction, roll, pitch, and yaw, and laser altimeter.

As described in Section 3, there are several instruments that might provide duplicate data (for example, on flights when the Piccolo and Canmore GPS are used). As such, there is a hierarchy to the type of data that is provided in the final dataset. Typically, data provided by instruments that are not part of the Piccolo system, and were added specifically as part of the 2009 flights, are used whenever available. This impacts latitude, longitude, altitude, pressure, and ground speed data. If the Canmore GPS data is available, the latitude, longitude, and altitude from that device will be used over the Piccolo GPS data. The only exception to this is the ground speed data. Because the Canmore GPS did not provide ground direction data, the ground speed and direction data reported is always from the Piccolo system.

Because the Piccolo air pressure was not used during flight operations and therefore was not properly calibrated, the Vaisala air pressure was used whenever available. In instances where the Piccolo air pressure needed to be used, the data was rigorously analyzed to ensure

Wind Direction <deg></deg>	0	0	231.50	236.14
Wind Speed <ms<sup>-1&gt;</ms<sup>	0	0	7.61	1.93
Pressure <pa></pa>	100378.09	100354.58	100255.65	99866.85
RH <%>	70.67	70.41	70.67	73.05
Temp <c></c>	-33.89	-35.61	-36.31	-30.6
Altitude <m></m>	4.71	5.70	3.54	33.88
Lon <deg></deg>	166.49	166.49	166.49	166.49
Lat <deg></deg>	-77.95	-77.95	-77.95	-77.95
Second	9.54	19.54	29.54	39.54
Minute	∞	∞	∞	∞
Hour	20	20	20	20
Day	23	23	23	23

Table 14. A sample of the data available in the USAP-DCC repository. Data has been truncated in this table to save space.

validity. The air pressure at flight take-off and landing was compared to a local automatic weather station at the Pegasus ice runway to determine both accuracy and that any differences between the two pressures at take-off were the same when the aircraft landed. For all flights where the Piccolo air pressure was used, it was deemed to be accurate.

# 5. Summary

In September 2009, a series of long-range unmanned aircraft system flights were made to Terra Nova Bay, Antarctica to study the downstream evolution of the atmospheric boundary layer over a latent heat polynya during winter conditions. Eight flights were flown on science missions to Terra Nova Bay, with an additional six flights flown near the Pegasus ice runway (the origin of the UAS flights) as test flights to ensure proper aircraft operations. Two additional flights were attempted to TNB, but were unsuccessful. The data was quality controlled and processed, and has been submitted to the USAP-DCC data repository for public use. The data available at the USAP-DCC consists of date and time, latitude, longitude, altitude, temperature, relative humidity, pressure, wind speed, wind direction, u and v components of the wind, longwave and shortwave radiation, skin temperature, ground speed and direction, and roll, pitch, and yaw provided at ten second intervals for all sixteen flights. The laser altimeter data for two of the flights is also provided.
## 7. Acknowledgments

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