What drives the low-level winds over the Ross Ice Shelf, Antarctica?

by

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A thesis submitted to the
Faculty of the Graduate School of the
University of Colorado in partial fulfillment
of the requirement for the degree of
Doctor of Philosophy
Department of Atmospheric and Oceanic Sciences
2012
This thesis entitled:

**What drives the low-level winds over the Ross Ice Shelf, Antarctica?**

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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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What drives the low-level winds over the Ross Ice Shelf, Antarctica?

Thesis directed by Associate Professor John J. Cassano

Due to the limited observations in the Antarctic and the challenges in validating numerical weather prediction model output, a new method for evaluating the performance of numerical weather prediction models is presented. Typical model evaluation techniques evaluate models using a case study of a specific event or over a large period of time (i.e. days, months, years, etc.). The new evaluation technique uses the method of self-organizing maps to conduct a weather pattern based model evaluation of the Antarctic Mesoscale Prediction System (AMPS). AMPS is used in the subsequent analysis of the low-level winds over the Ross Ice Shelf.

The dynamics of the low-level wind field over the Ross Ice Shelf are investigated using a case study of a high wind event off the coast of the Prince Olav Mountains and a low-level wind climatology over the Ross Ice Shelf. Both of these analyses use automatic weather station observations and output from AMPS.

The region to the northwest of the Prince Olav Mountains contains some of the fastest mean wind speeds over the Ross Ice Shelf. The case study presented in this dissertation diagnoses the atmospheric dynamics associated with the strong winds in this region, concluding the area of maximum wind speed is a barrier wind corner jet.
The low-level wind climatology presented in this dissertation is based on output from the 15 km Weather Research and Forecasting model run within AMPS and is the first Ross Ice Shelf wind climatology presented at this resolution. The wind climatology provides information on the types of winds present over the Ross Ice Shelf and quantifies the frequency and seasonality of these patterns. Subsequently, a climatology of the Ross Ice Shelf airstream identifies the variability in the position and strength of the Ross Ice Shelf airstream and quantifies the frequency and seasonality of these patterns. The atmospheric dynamics associated with the barrier wind component of the Ross Ice Shelf airstream are analyzed.
Acknowledgements

This research was supported by National Science Foundation grants ATM-0404790, ANT-0636811, ANT-0838834, and ANT-0943952 and teaching assistantships through the Department of Atmospheric and Oceanic Sciences at the University of Colorado at Boulder.

I would like to thank John Cassano for allowing me to fulfill my dream of studying the Antarctic. He has taught me how to conduct fieldwork in Antarctica, how to objectively analyze scientific data and how to effectively write a paper. Most importantly, he patiently, guided and supported me through this dissertation.

I am thankful to my committee members Andrew Monaghan, David Noone, Cora Randall and Mark Serreze for their guidance and feedback throughout the research process.

I would like to recognize Matthew Lazzara, Linda Keller, Jonathan Thom, George Weidner and Lee Wellhouse at the Antarctic Meteorological Research Center (AMRC) at the University of Wisconsin—Madison for providing extensive quality controlled automatic weather station observations, as well as, Jordan Powers and Kevin Manning at the National Center for Atmospheric Research for providing the Antarctic Mesoscale Prediction System (AMPS) archive.

I am grateful to the members of the Cassano group Elizabeth Cassano, Alice DuVivier, Matthew Higgins, Mimi Hughes, Shelley Knuth, David Porter, Keah Schuenemann, and Mark
Seefeldt for providing programming support, manuscript proof reading, and hours of discussion on atmospheric dynamics.

Many thanks to my husband, Thomas Nigro, my parents, Bruce and Phyllis Richards, and my sister, Stefanie Richards, for many years of support and encouragement leading up to this dissertation.
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Chapter 1: Introduction

The low-level winds over the Ross Ice Shelf (RIS), Antarctica are greatly influenced by the steep topography surrounding the RIS. Figure 1 shows a map of the RIS region, which is bordered by the Transantarctic Mountains to the west, the West Antarctic Ice Sheet to the east and the Ross Sea to the north. The average elevation of the RIS is approximately 50 m and the peaks of the Transantarctic Mountains reach over 4500 m. This sharp contrast in topography provides forcing for mesoscale wind features such as katabatic winds, barrier winds, and the Ross Ice Shelf airstream (RAS). The combination of these features with the synoptic winds from cyclones that traverse the Ross Sea, provide a complex low-level wind field over the RIS.

Fig. 1. Map of the Ross Ice Shelf region.
Katabatic winds occur when negatively buoyant air flows downward over sloping terrain (Parish 1988). In the region of the RIS, katabatic winds flow down the glacial valleys of the Transantarctic Mountains and the Siple Coast confluence zone (Figure 1), draining the cold, continental air from the interior of the continent onto the RIS (Parish and Bromwich 1987). Katabatic winds most frequently occur during the austral winter when the lack of sunlight allows a strong surface inversion to develop over the sloping terrain of the Transantarctic Mountains and the Siple Coast confluence zone, providing forcing for the katabatic winds in this region (Parish and Cassano 2003).

Katabatic winds are an important component to the Southern Hemisphere tropospheric circulation. Parish and Bromwich (1998) analyzed a case study of a 20 mb drop in pressure over the Antarctic continent during a 4-day period in the winter of 1988. The study showed the drop in pressure was mainly driven by katabatic drainage of mass off the Antarctic continent. This event transported mass to lower latitudes, altering the pressure field from the Antarctic continent into the subtropics of the Southern Hemisphere. Katabatic events, such as this, create an upper level return flow onto the Antarctic continent, satisfying the physics of mass continuity (Parish 1992). More specifically, the katabatic drainage of mass off the continent causes low-level divergence and sinking motion over the interior of the continent. The sinking motion induces convergence aloft, resulting in the upper level return flow onto the continent.

Barrier winds, another common wind feature over the RIS, form when stably stratified flow is directed towards a barrier, or mountain. If the Froude number of the
approaching flow is less than one, the barrier blocks the flow and mass convergence occurs (O’Connor et al. 1994; Buzzi et al. 1997). The mass convergence increases the pressure at the base of the barrier, creating a pressure gradient force (PGF) that is directed perpendicular and away from the barrier. The PGF induces a wind that becomes approximately geostrophic and flows parallel to the barrier. This barrier parallel flow is called a barrier wind. In the region of the RIS, barrier winds form along the base of the Transantarctic Mountains flowing from the southeast towards the northwest (O’Connor et al. 1994; Parish et al. 2006; Seefeldt et al. 2007; Steinhoff et al. 2009).

A combination of katabatic winds, barrier winds and synoptic forcing create the RAS (Parish et al. 2006). The RAS is a dominant stream of air originating in the Siple Coast confluence zone (Parish and Bromwich 1986), flowing over the western to central RIS to the north over the Ross Sea (Seefeldt and Cassano 2012). The RAS drains cold, continental air from the interior of the continent, transporting it across the RIS to the Ross Sea. In fact, in the Parish and Bromwich (1998) case study where a katabatic event caused a 20 mb drop in pressure over the Antarctic continent, it was shown that approximately one-third of the mass drained from the continent passed through the Siple Coast confluence zone and was transported to more northerly latitudes by the RAS. This type of transport makes the RAS an important component of the Southern Hemisphere climate system.

The research presented in this dissertation was conducted to better understand the basic atmospheric dynamics that drive the low-level winds over the RIS. The low-level winds over the RIS are forced by a combination of the steep topography surrounding the
RIS and synoptic forcing from cyclones that traverse the Ross Sea. A similar interaction of synoptic and topographic forcing occurs in many regions of the world. For example, the barrier winds along the Transantarctic Mountains are similar to the barrier parallel winds along the Sierra Nevada Mountains, the Appalachian Mountains, and the mountains along southeastern coast of Greenland. Therefore, the atmospheric dynamics diagnosed in this dissertation have application to other regions of the world and, in general, further the basic scientific understanding of the atmosphere.

The low-level winds over the RIS also create severe weather in the region of the RIS. For example, a severe windstorm passed through McMurdo station on 15-16 May 2004. McMurdo station, which is located on Ross Island in the region of the RIS, is the largest of the United States Antarctic Program bases and a hub for many logistical activities. The windstorm produced winds up to 71.5 m s⁻¹ at McMurdo, causing structural damage to the buildings (Powers 2007; Steinhoff et al. 2008). Therefore, for the safety of people working in the Antarctic it is important to be able to accurately forecast severe weather events in the region of the RIS. This dissertation presents information on the dynamics that cause strong winds over the RIS, quantifies the frequency and seasonality of these strong winds and identifies the weather patterns that typically evolve into these strong wind events. This type of information will hopefully better the forecasts of strong wind events over the RIS in the future.

The overall intent of this research is to answer the question, “What drives the low-level winds over the RIS?” To answer this question, a combination of observations and
numerical weather prediction (NWP) model output is used. This is a common research approach in the Antarctic, where observations are spatially sparse. Due to the lack of observations in the Antarctic region, it is typical to use the available observations to validate the performance of a model in the region of interest, providing some level of confidence in using the model output for research purposes. Typical model evaluation techniques evaluate models using a case study for a specific event or over a large period of time (i.e. days, months, seasons, years, etc.).

Through the use of NWP model output, it has been observed that NWP models perform better under some weather patterns than others. Therefore, the ability to analyze the performance of an NWP model as a function of weather pattern could be a useful tool when validating model output for use in Antarctic research. The typical model evaluation techniques listed above (case studies and over a large period of time) do not directly address the evaluation of model performance as a function of weather pattern. A series of case studies could be used to determine the model performance as a function of weather pattern, although this would be time consuming. For this reason, Chapter 2 of this dissertation investigates a new approach for evaluating the performance of NWP models as a function of weather pattern. This new approach uses the method of self-organizing maps (SOM) (Kohonen 2001) to conduct a weather pattern based model evaluation of the Antarctic Mesoscale Prediction System (AMPS). The hypothesis is that this technique will be able to determine the model errors as a function of weather pattern, providing information on the errors associated with each weather regime in the region. The contents of Chapter 2 were published in *Weather and Forecasting* in an article titled “A Weather
Pattern Based Approach to Evaluate the Antarctic Mesoscale Prediction System (AMPS) Forecasts: Comparison to Automatic Weather Stations Observations” (Nigro et al. 2011).

The analysis of the low-level wind field over the RIS starts in Chapter 3, where a case study of a high wind event off the coast of the Prince Olav Mountains is presented. The region to the northwest of the Prince Olav Mountains contains some of the fastest mean wind speeds over the RIS (Parish et al. 2006). Parish et al. (2006) used the MM5 model run within AMPS to estimate mean annual wind speeds of approximately 20 m s\(^{-1}\) (at an approximate height of 300 m) in this region. The winds in this region have previously been studied (Seefeldt and Cassano (2008); Steinhoff et al. (2009)). Seefeldt and Cassano (2008) hypothesized the winds were similar to the tip jets found off the southern coast of Greenland (Moore and Renfrew 2005; Renfrew et al. 2009; Outten et al. 2009; Outten et al. 2010) and Steinhoff et al. (2009) referred to these winds as a “knob flow”, or a corner wind (Dickey 1961; Kozo and Robe 1986; Olafsson 2000; Olafsson and Agustsson 2007; Lefevre et al. 2010). Although, neither study was able to explicitly state the atmospheric dynamics that drive the winds to the northwest of the Prince Olav Mountains. Each study concludes additional research is necessary to diagnose the atmospheric dynamics associated with the strong winds in this region. Therefore Chapter 3 analyzes a case study of a high wind event in the region of the Prince Olav Mountains to determine the dynamics that drive the winds in this region. The contents of Chapter 3 were published in *Monthly Weather Review* in an article titled “Case Study of a Barrier Wind Corner Jet Off the Coast of the Prince Olav Mountains, Antarctica” (Nigro et al. 2012).
The analysis of the low-level wind field over the RIS continues in Chapter 4, where a low-level wind climatology over the RIS is presented. The low-level wind climatology is based on output from the 15 km Weather Research and Forecasting (WRF) model run within AMPS and is the first RIS wind climatology presented at this resolution. Seefeldt and Cassano (2012) presented a low-level wind climatology over the RIS using output from the 30 km MM5 model run within AMPS. As will be shown in Chapter 4, the accuracy of the wind forecasts over the RIS greatly improved with the change from the AMPS-MM5 30 km dataset to the AMPS-WRF 15 km dataset. Given the better representation of the low-level wind field over the RIS with the AMPS-WRF 15 km dataset, the wind climatology presented here is a likely improvement over the previously presented wind climatology using the AMPS-MM5 30 km dataset. The wind climatology presented in Chapter 4 provides information on the types of wind patterns present over the RIS, as well as, the frequency and seasonality of these patterns.

Subsequently, a RAS climatology is investigated to better understand the different RAS patterns that occur over the RIS. Through the analysis of wind patterns over the RIS, it has been observed that the location of the RAS ranges from strong winds adjacent to the Transantarctic Mountains to strong winds located through the center of the RIS. The RAS climatology will determine the types of RAS patterns that occur over the RIS (showing the variability within the RAS), as well as, the frequency and seasonality of these patterns. The results of RAS climatology will be analyzed to understand the atmospheric dynamics that drive the RAS. It is hypothesized that the variability within the RAS patterns is a result of differing atmospheric dynamics. For instance, RAS patterns with strong winds adjacent to
the Transantarctic Mountains likely have a barrier wind component to the forcing while RAS patterns with strong winds through the center of the RIS likely have little to no barrier wind component to the forcing. In Chapter 4, the atmospheric dynamics associated with the barrier wind component of the RAS in these patterns are analyzed and presented. These findings are in preparation to be submitted to *Monthly Weather Review* in an article titled, “Analysis of high winds over the Ross Ice Shelf, Antarctica Part 1: Barrier winds along the Transantarctic Mountains.”
Chapter 2: A Weather-Pattern-Based Approach to Evaluate the Antarctic Mesoscale Prediction System (AMPS) Forecasts: Comparison to Automatic Weather Station Observations

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This article appears in the April 2011 issue of Weather and Forecasting Volume 26, Number 2 published by American Meteorological Society and available online through AMS Journals. http://dx.doi.org/10.1175/2010WAF2222444.1

Abstract: Typical model evaluation strategies evaluate models over large periods of time (months, seasons, years, etc.) or for single case studies such as severe storms or other events of interest. The weather-pattern-based model evaluation technique described in this
paper uses self-organizing maps to create a synoptic climatology of the weather patterns present over a region of interest, the Ross Ice Shelf for this analysis. Using the synoptic climatology, the performance of the model, the Weather Research and Forecasting Model run within the Antarctic Mesoscale Prediction System, is evaluated for each of the objectively identified weather patterns. The evaluation process involves classifying each model forecast as matching one of the weather patterns from the climatology. Subsequently, statistics such as model bias, root-mean-square error, and correlation are calculated for each weather pattern. This allows for the determination of model errors as a function of weather pattern and can highlight if certain errors occur under some weather regimes and not others. The results presented in this paper highlight the potential benefits of this new weather-pattern based model evaluation technique.

1. Introduction

Typical model evaluation strategies evaluate models over large periods of time (months, seasons, years, etc.) or for single case studies such as severe storms or other events of interest. The technique described below, which is based on the use of self-organizing maps (SOMs; Kohonen 2001), involves objectively identifying the synoptic weather patterns that influence the region of interest, classifying each model forecast time as matching one of these weather patterns, and calculating model validation statistics for each weather pattern. This allows for the determination of model errors as a function of
weather pattern and can highlight if certain errors occur under some weather regimes and not others.

As mentioned above, there are currently two basic methods that are used to evaluate numerical weather prediction models. These methods will be reviewed with relevant Antarctic examples cited. The first method commonly calculates statistics [e.g., bias, root-mean square error (RMSE), and correlation] for comparisons of model forecasts and observations of numerous variables (e.g., temperature, pressure, wind speed, and wind direction) for a given location. There are many variations that can be made to this technique. For instance, the analysis can be categorized temporally by seasons, months, times of day, etc., or spatially by region, elevation, proximity to the coast, etc. Guo et al. (2003) used this method to evaluate the polar-modified version of the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5; Grell et al. 1995) over Antarctica. The research involved analyzing several variables, such as surface pressure, temperature, wind speed, and wind direction, on annual, monthly, daily, and hourly time scales. The study concluded that this version of MM5 was able to successfully predict the surface pressure, temperature, wind direction, and water vapor mixing ratio on all time scales; and that improvements to the model were necessary in order to accurately predict some of the other atmospheric variables, such as wind speed. Similarly, Bromwich et al. (2005) analyzed the performance of the polar-modified version of the MM5 model used in the Antarctic Mesoscale Prediction System (AMPS). The analysis looked at the spatial variability, seasonal variability, and forecast hour variability associated with the model performance. The study found that the surface temperature
predictions are most accurate during the winter months, the surface pressure is well represented by the model, and the surface winds are strongly dependent on the topography of the region. The type of analysis shown in these two examples provides important information on how the model performs on large scales, both spatially and temporally, but does not identify model errors that occur under specific synoptic patterns.

The second method that is commonly employed to evaluate numerical weather prediction models makes use of case studies. Case studies usually involve an isolated weather event where some knowledge of the atmospheric conditions is known and the ability of a model to simulate the event is assessed. For instance, Powers (2007) evaluated the ability of the Advanced Research Weather Research and Forecasting Model (ARW-WRF) running in AMPS to simulate a windstorm that impacted McMurdo Station, Antarctica, in May 2004. The analysis used Moderate Resolution Imaging Spectroradiometer (MODIS) infrared satellite imagery and surface observations as a comparison for the model output. The model in AMPS was run with different data types assimilated, and it was determined that the best model representation of the windstorm used the assimilation of filtered MODIS data as an input to the model. Another example of the case study method of model evaluation is presented in Bromwich et al. (2003). In this study, surface observations, ship observations, and satellite images were used in order to evaluate the polar-modified version of the MM5 model in AMPS under conditions of mesoscale cyclogenesis during 13–17 January 2001. Each of the case studies mentioned here were able to provide evidence of successes and weaknesses within the model for a very specific time and place.
The weather-pattern-based method of model evaluation, as presented in this paper, combines advantages of both methods discussed above. The weather-pattern based method allows for analysis over long periods of time, while providing information about model errors under a variety of synoptic situations. Due to the fact that models do not perform with equal accuracy under various synoptic weather patterns, this can be an important level of detail for all model users to understand.

To conduct a weather-pattern-based model evaluation, the prominent weather patterns of the region of interest must first be identified. The SOM technique will be used for this process because of its ability to objectively group large amounts of data into categories that can easily be analyzed. Researchers have used this technique to relate synoptic patterns to a variety of atmospheric features such as precipitation and temperature. For instance, the SOM technique was used to analyze changes in net precipitation in the Arctic (Cassano et al. 2007) and the Antarctic (Uotila et al. 2007) by using various model scenarios from the Intergovernmental Panel on Climate Change (IPCC). Additional examples include the use of the SOM technique to analyze the winds over the Ross Ice Shelf (RIS) from a synoptic climatology perspective (Seefeldt and Cassano 2008) and to analyze the changes in atmospheric circulation, temperature, and precipitation caused by a reduction in Arctic sea ice (Higgins and Cassano 2009).

The SOM-based analysis presented here has been completed as a preliminary study to understand the benefits of using SOMs as a tool for evaluating operational numerical
weather prediction forecasts. The analysis for this paper was conducted over the Ross Ice Shelf, Antarctica, using forecasts from AMPS.

2. Data and methods

2.1 Model data

Forecasts from the ARW-WRF running within AMPS, a numerical weather prediction system tailored for use in the Antarctic (Powers et al. 2003), are evaluated in this paper. AMPS was originally developed to provide high-resolution, polar-specific forecasts over Antarctica in support of U.S. Antarctic Program (USAP) operations. AMPS is currently used as an experimental real-time numerical weather prediction model by USAP and other Antarctic programs.

AMPS has been based on two different numerical weather prediction models since its origin. The original version of AMPS was based on the polar-modified version of MM5 (Bromwich et al. 2001; Cassano et al. 2001). The more recent version of AMPS uses a polar-modified version of the WRF (Hines and Bromwich 2008) and was introduced in 2006. During the time period used for this study (the operational field seasons, October–February, of 2005–06, 2006–07, and 2007–08), the two models in AMPS (MM5 and WRF) were run simultaneously. Therefore, the training of the SOM will use both the MM5 and WRF output,
while the analysis presented in this paper is conducted on the WRF forecasts from AMPS only.

The WRF in AMPS is run with a set of six two-way interactively nested domains with horizontal grid spacing of 60, 20, 6.7, 6.7, 2.2, and 6.7 km for the time period of this study (see Fig. 2). The analysis for this study is conducted on the AMPS output for domain 2 (20 km), but could easily be applied to the other available domains. AMPS is initialized twice daily, with simulations on domain 2 run for 72 h with the output archived every 3 h. For more information on the physics packages used in the operational version of AMPS refer to Powers (2007) and the AMPS Web site (http://www.mmm.ucar.edu/rt/wrf/amps/).

![Fig. 2. Map of the AMPS domains. Domains 1, 2, 3, 4, and 6 are labeled in the upper-right corner of the respective outlined boxes. Domain 5 is located within the outline of domain 3 and is represented by an unlabeled box. (Figure courtesy of NCAR/Mesoscale and Microscale Meteorology Division; information online at http://www.mmm.ucar.edu/rt/wrf/amps/information/configuration/maps.html).]
2.2 Weather pattern identification

To use the weather-pattern-based method of model evaluation, a set of common weather patterns that occur in the region of interest must first be identified. For the identification process, the self-organizing maps technique was used to objectively determine the weather patterns that influence the RIS region of Antarctica. The SOM training process uses a neural network algorithm that includes an unsupervised, iterative learning process to identify general patterns in a dataset. The training process organizes similar data records into a user-specified number of groupings. Kohonen (2001) provides a thorough explanation of the SOM algorithm and training process. For the evaluation of AMPS, the SOM technique was used to create a synoptic climatology of the weather patterns that occur in the Ross Sea sector of Antarctica during the operational field season (1 October–15 February). To do so, the SOM was trained with the 20-km-grid MM5 forecasts of sea level pressure from the AMPS archives for the 2005–06, 2006–07, and 2007–08 USAP operational field seasons and the 20-km-grid WRF forecasts of the sea level pressure from the AMPS archives for the 2006–07 and 2007–08 USAP operational field seasons. Fig. 3a shows the section of the 20-km-grid output that was used for the training of the SOM. Specifically, the sea level pressure anomaly fields were used for the training process. The use of the sea level pressure anomalies is preferred for this analysis, because atmospheric circulation is dependent on pressure gradient, not the magnitude of the sea level pressure. To calculate the anomaly field for each forecast, the sea level pressure at each model grid point was retrieved from the model output for the forecast. The sea level
pressure values for elevations greater than 500 m were removed due to the difficulty in calculating accurate sea level pressure over the high terrain of Antarctica. From this dataset, the domain-averaged sea level pressure was calculated, which was then subtracted from the sea level pressure at each grid point, resulting in a field of sea level pressure anomalies. The resultant sea level pressure anomaly fields were used to train the SOM.

Fig. 3. (a) Analysis region for synoptic climatology (black box) and (b) location of AWS sites used for model evaluation.
During the training process, the user must specify the number of weather patterns that the SOM should identify. For the model evaluation application presented here, the SOM technique was used to identify a number of different patterns, but it was found that using 20 weather patterns, or a 5 X 4 grid size, was sufficient to capture the range of synoptic conditions that influence the RIS region while maintaining a sufficient number of data points associated with each weather pattern for calculating statistics. These 20 weather patterns, when shown together as in Fig. 4, are referred to as the master SOM and each of the weather patterns identified on the master SOM is referred to as a node. Reusch et al. (2005) provides a good explanation for determining the number of patterns the SOM should identify. The authors define the quantization error as the difference between the model forecast and the SOM-identified pattern that it is mapped to, or the pattern that it most closely resembles. The paper concludes that the quantization error decreases with an increase in grid size. Essentially, the larger the grid size, the closer the forecast matches one of the weather patterns that make up the SOM. With an increase in the number of weather patterns, the number of forecasts that map to each weather pattern decreases, and thus the robustness of the statistics calculated for each weather pattern decreases. To calculate meaningful statistics on the accuracy of the model within AMPS for a given weather pattern, a large enough number of forecasts with valid observations must be mapped to each weather pattern. When choosing the SOM size for this paper, it was determined that either a 5 X 4 or a 6 X 4 grid size would be sufficient to capture the variability of weather patterns present over the RIS, based on the authors previous experience analyzing weather patterns over this region for studies of the low-level winds (Seefeldt and Cassano 2008) and in
working with USAP operational weather forecasters. Ultimately, the 5 X 4 grid size was chosen in order to maximize the number of forecasts and observations mapped to each weather pattern to allow for calculation of meaningful statistics.

Fig. 4. Master SOM of sea level pressure anomalies over the RIS.

As mentioned above, Fig. 4 depicts each of the weather patterns identified by the SOM training, showing the sea level pressure anomalies for each weather pattern. The nodes on the SOM are referenced by their column number (one is the leftmost column) and by their row number (one is the top row), such that node [1,1] indicates the node in the
top-left corner of the SOM and node [5,4] indicates the node in the bottom-right corner of the SOM. In this plot of the sea level pressure anomalies, the blue colors indicate areas of lower pressure and the red colors indicate areas of higher pressure. Therefore, it can be seen that node [1,1] has a strong cyclone in the northeastern corner of the domain, over the Ross Sea. Similarly, node [5,4] has a very weak cyclone in the southeastern corner of the domain, over the southern part of the RIS. One advantage of the SOM methodology is that the nodes on the master SOM are arranged such that similar patterns are adjacent, while the least similar patterns are located in opposite corners of the plot (Fig. 4). A general analysis of the plot reveals that by moving across the top row the cyclone strength decreases and the cyclone center shifts toward the western Ross Sea. Other general observations include the following: the strongest low pressure systems are found on the left side of the SOM, cyclones centered in the northern part of the domain are found on the left and top sides of the SOM, southerly cyclones are represented by patterns in the lower-right portion of the SOM, and the weakest cyclones are found in the center-right portion of the SOM. This broad grouping of the weather patterns shown on the SOM will be useful when analyzing the performance of AMPS for the different weather regimes.

For the SOM evaluation of the WRF running in AMPS, the 20-km WRF forecasts from the operational field seasons of 2006–07 and 2007–08 were used. This process involves matching each model forecast to the weather pattern on the master SOM that most closely resembles it. Specifically, the squared differences between the sea level pressure anomalies in the forecast and the sea level pressure anomalies of each SOM node are calculated. The node that results in the smallest squared difference is the weather pattern that the forecast
is matched to. This process is known as “mapping” the forecasts to the master SOM and is repeated for each model forecast. Once the mapping is complete, a set of model forecasts are associated with each node and can be used in conjunction with a set of observations to evaluate the performance of the model for each weather pattern.

2.3 Observational data

To evaluate the model, the forecasts must be compared to a set of observations. In this instance, surface observations from the University of Wisconsin automatic weather stations (AWSs) were used. The AWSs take measurements of temperature, pressure, wind speed, and wind direction at an interval of 10 min. These observations are then processed by the Antarctic Meteorological Research Center of the University of Wisconsin where the data are quality controlled and a time series of 3-hourly observations is created. The quality control is a manual process where erroneous observations are removed from the dataset. The process of creating the 3-hourly observations involves extracting the observation that is made closest to the 3-hourly time, ±40 min (Keller et al. 2010). This process is used due to the scarcity of observations in Antarctica and the high occurrence of missing data points from AWS observations. The ±40-min window is the standard used by the AWS research group at the University of Wisconsin when creating the quality controlled 3-hourly AWS data, and as such is the dataset used for the model evaluation presented in this paper.
Prior to comparison, the model surface pressure was adjusted to the elevation of the AWS by using the hypsometric equation. This is necessary due to the fact that the model resolution spatially smooths the topography and the model grid point elevation may not match the actual elevation at the AWS sites. For the majority of the AWS sites, the differences between the actual elevation and the model grid point elevation were small, resulting in pressure adjustments on the order of a few tenths of a millibar, although a few AWS sites had much larger differences, with a maximum elevation difference of 117 m at the Cape Bird AWS. The pressure adjustment associated with this elevation difference was approximately 14 mb.

Given that the analysis presented in this paper is intended primarily as a demonstration of the utility of the weather-pattern-based model evaluation technique, a variety of AWS sites from across the RIS will be used. The locations of these sites are shown in Fig. 3.

2.4 Model evaluation

As mentioned above, after the mapping of the model forecasts a set of model forecasts exists for each weather pattern shown in Fig. 4. Subsequently, the observations that match the forecast valid times (±40 min) can be obtained, resulting in a set of model forecasts and AWS observations for each weather pattern. These datasets can further be grouped into the following forecast categories: 0–9, 12–21, 24–33, 36–45, 48–57, and 60–
69 h in order to evaluate how the model performs as a function of forecast hour. Due to the fact that the model is initialized twice daily, extracting 12-h segments (four 3-h segments) from each model run is sufficient to create a continuous time series. From these datasets, the average temperature, pressure, wind speed, and wind direction of both the model output and the AWS observations can be calculated for various locations within the domain. Additionally, statistics such as the model bias, the model root-mean-square error, and the model correlation can also be calculated from the paired model and observation time series. The results can be displayed in what will be referred to as a verification plot. The verification plot will show each of the statistics mentioned above as a function of the SOM nodes (weather patterns). In these plots, the associated node is labeled along the x axis. Additionally, statistics have been calculated for an “ALL” node. The ALL node represents the statistics calculated for all 20 weather patterns. This is representative of the traditional method of model evaluation discussed in the introduction where the statistic of interest would be calculated for all weather patterns occurring over the period of interest. An example of a verification plot is shown in Fig. 5 and displays the average wind speed, the wind speed bias, and the wind speed RMSE at Lettau.
Fig. 5. (a) Verification plot of average pressure for Lettau AWS location calculated for the AWS observations (black dotted line) and the following forecast categories: 0–9 (black solid line), 12–21 (red solid line), 24–33 (orange solid line), 36–45 (green solid line), 48–57 (light blue solid line), and 60–69 h (dark blue solid line). (b) Verification plot of pressure bias calculated for the Lettau AWS location for the same forecast categories (same lines and colors) presented in (a). (c) Verification plot of pressure RMSE calculated for the Lettau AWS location for the same forecast categories (same lines and color) presented in (a) and (b).
3. Results

The main goal of this paper is to demonstrate the utility of the weather-pattern-based model evaluation technique. Figure 5a shows the verification plot for the average pressure at the Lettau AWS (LET in Fig. 3b). In the verification plot, the average pressure for the 0–9-, 12–21-, 24–33-, 36–45-, 48–57-, and 60–69-h model forecast categories and the average pressure for the AWS observations that match the 0–9-h valid times are shown. Viewing the information plotted in Fig. 5a, from left to right, shows the pressure averaged over all of the weather patterns and then averaged for each individual weather pattern (node) shown in Fig. 4, progressing from the top-left corner of Fig. 4, down the leftmost column, and then proceeding down through each remaining column. The vertical lines in Fig. 5a indicate weather patterns at the top of each column in Fig. 4. The pressures plotted in Fig. 5a can be related to the master SOM in Fig. 4. For instance, the results that relate to the first column of the master SOM (nodes [1,1], [1,2], [1,3], and [1,4] in Fig. 5a) show that the AMPS-forecasted pressure at Lettau increases when moving from the weather pattern in node [1,1] to the weather pattern in node [1,2]. Subsequently, the AMPS-forecasted pressure decreases when moving to the weather patterns in nodes [1,3] and [1,4]. A careful analysis of Fig. 4 will reveal the same results, as the cyclone is positioned farther north and east in node [1,2] than in node [1,1] resulting in a higher pressure at Lettau for the weather pattern in node [1,2]. Subsequently, in nodes [1,3] and [1,4] the cyclone shifts farther south causing the pressure at Lettau to decrease for both of these weather patterns. A similar
analysis can be conducted for the remaining columns. By doing so, it will be seen that Lettau experiences the highest sea level pressure for the weather patterns identified in the top row of the master SOM and the lowest sea level pressure for the weather patterns in the bottom row of the master SOM. Lettau is located toward the south-central portion of the RIS and, therefore, these results are consistent with the positioning of the cyclones for the different weather patterns as discussed when Fig. 4 was introduced. The cyclones centered in the northern part of the domain are found in the top row of the SOM, causing the pressure at the more southern Lettau site to be higher in these instances. Additionally, the southerly cyclones are represented by the weather patterns in the bottom row of the master SOM, which causes lower sea level pressure at Lettau for those weather patterns.

As mentioned above, Fig. 5a shows the verification plot for the average pressure at the Lettau AWS. It can be seen that the average pressure at Lettau ranges from approximately 968 to 987 hPa depending on the weather pattern present and the model forecast hour. This indicates that the pressure at Lettau is quite variable and strongly correlated to the weather pattern present over the RIS (as seen by the sawtooth pattern in the average pressure verification plot). There are other AWS sites on the southern ice shelf that display similar pressure results (plots not shown). These sites include Elaine, Gill, Marilyn, and Schwerdtfeger. To better understand how the model performs for different weather patterns, the model forecasts are compared to the AWS observations. In Figs. 5b and 5c, respectively, the model bias and the model RMSE verification plots for the pressure at Lettau are shown. It can be seen that even though the pressure at Lettau is variable and dependent on the weather pattern (Fig. 5a), the model bias and RMSE for the shorter
forecast hours are fairly consistent over the different weather patterns identified by the master SOM (Figs. 5b and 5c). This is shown in the verification plots by the lack of variation in the pressure bias and RMSE from left to right across Figs. 5b and 5c for the shorter forecast hours, which indicates small variation in the model error from one weather pattern to the next. Therefore, for the short-term forecasts, the model predicts the pressure at Lettau with a consistent bias. As a result, the bias value indicated in Fig. 5b could be used as an approximation to adjust the forecasted model pressure at this location as a post processing correction. As the model forecast times increase to 45–57 and 60–69-h, the model bias and RMSE are no longer consistent between weather patterns. This is shown in the results for the 60–69-h model bias and RMSE (the dark blue line in Figs. 5b and 5c), where the bias ranges from approximately -4 to -9 hPa and the RMSE ranges from approximately 5 to 9.5 hPa. This indicates that for the longer forecast hours, the model predicts the pressure more accurately in some weather patterns than others. In this instance, a standard bias cannot be applied to adjust the forecasted model pressure. Instead, the bias would need to be applied as a function of the synoptic weather pattern.

A similar analysis can be conducted for the wind speed at Pegasus North. Figure 6 shows the verification plot for the average wind speed and the model bias at the Pegasus North AWS. It can be seen that AMPS predicts the strongest wind speeds at Pegasus North for the weather patterns identified at the top of the three left-most columns in the master SOM (Fig. 4), with wind speed decreasing sharply when moving down each column. Referring back to the sea level pressure anomalies plot of the master SOM (Fig. 4), this is consistent with the strong pressure gradient that exists over Pegasus North for nodes [1,1],
[2,1], and [3,1] due to the location of the cyclone in the northern portion of the domain for these weather patterns with a weakening pressure gradient found by moving down each of these columns. Conversely, the average AWS wind speeds at Pegasus North are not as strong as those simulated by AMPS and do not show the strong variability when moving from the top to the bottom of the SOM column. Analysis of the wind speed bias verification plot (Fig. 6b) at Pegasus North indicates that the model results are not consistent over the identified weather patterns (Fig. 6b), as seen by the variability in the wind speed bias between different weather patterns. The plot shows that on average the model greatly overpredicts the wind speed for the weather patterns in the first row of the SOM and that the model bias for the remaining weather patterns is quite variable, but generally small.
Fig. 6. (a) Verification plot of average wind speed for the Pegasus North AWS location calculated for the AWS observations (black dotted line) and the following forecast categories: 0-9 (black solid line), 12-21 (red solid line), 24-33 (orange solid line), 36-45 (green solid line), 48-57 (light blue solid line), and 60-69 h (dark blue solid line). (b) Verification plot of wind speed bias for the Pegasus North AWS location calculated for the same forecast categories (same lines and colors) presented in (a).

A situation where the model bias is variable over the identified weather patterns provides an excellent example of the advantage to using the weather-pattern-based method of model evaluation. For instance, one of the traditional methods for analyzing forecasting models involved calculating an average of the forecasted wind speeds and an average of the observed wind speeds at Pegasus North. Subsequently, the bias would have
been calculated as the difference between these two averages. The ALL node that is labeled in the verification plots is the equivalent of the traditional method described above. The ALL node represents the statistic calculated over all of the 20 weather patterns. The ALL model bias in Fig. 6a reveals that a traditional method of model evaluation would have indicated that on average the model over-predicts the wind speed at Pegasus North by about 1.5 m s\(^{-1}\). Using the weather-pattern-based method for model evaluation reveals that depending on the weather pattern the model can overpredict the wind speed by almost 7.5 m s\(^{-1}\) or underpredict the wind speed by 1.5 m s\(^{-1}\). In this instance, the use of the weather-pattern based method of model evaluation provides information to the end user that can be valuable in analyzing the forecasted wind speed at Pegasus North.

The weather-pattern-based method of model evaluation can also be used to determine if model errors associated with a given weather pattern are spatially consistent. Figure 7 shows the wind speed average and bias verification plots for Ferrell AWS (FER in Fig. 3b). It can be seen in the average wind speed verification plot that the model-forecasted wind speed is connected to the synoptic weather pattern in a fashion similar to that of the model-forecasted wind speed at Pegasus North. The strongest wind speeds at Ferrell are also forecasted for the synoptic patterns found in the top-left corner of the master SOM (as seen by the sawtooth pattern in the average wind speed verification plot).
Ferrell is located approximately 100 km to the northeast of Pegasus North. Pegasus North is positioned to the south of Ross Island in a region of complex terrain while Ferrell is situated east of Ross Island in an area away from significant topographic influences. Taking a look at the Ferrell wind speed bias verification plot (Fig. 7b), it can be seen that unlike at Pegasus North, the model forecasts the wind speeds at Ferrell with a fairly consistent bias over each of the weather patterns. Therefore, the model is correctly
adjusting the forecasted wind speed at Ferrell as the synoptic weather patterns change. The weather-pattern-dependent bias at Pegasus North is thus likely due to a poorly handled representation of the interaction of the synoptic flow with the local complex topography in AMPS at this coarse resolution.

Expanding on the spatial analysis of model performance, the Willie Field AWS (WFD in Fig. 3) is located about 16 km to the northeast of Pegasus North (see Fig. 3b). Even though Willie Field is situated somewhat in between Pegasus North and Ferrell, the average wind speed verification plot shows a much less defined sawtooth pattern for the forecasted wind speed over the different weather patterns (Fig. 8a). Similar to Ferrell and Pegasus North, the strongest wind speed at Willie Field is forecasted for the weather patterns in the top-left corner of the master SOM, but the magnitude of these wind speeds is much less than at Pegasus North or at Ferrell. For instance, the average forecasted wind speed for node [1,1] for the 12–21-h forecast category at Pegasus North is 10.8 m s\(^{-1}\), at Ferrell it is 11.8 m s\(^{-1}\), and for Willie Field it is 5.6 m s\(^{-1}\). The lower wind speeds at Willie Field can be attributed to the topographic influence of Ross Island within the model. Monaghan et al. (2004) analyzed the average wind speed and direction from the 3.3-km grid WRF output from AMPS in the area of Ross Island. Figure 3 in Monaghan et al. (2004) shows that the Willie Field AWS is located in an area of climatologically low wind speeds due to the persistent southerly winds in the area and the blocking effect of Ross Island. Figure 8b shows the wind speed bias for the Willie Field AWS and indicates that the model predicts the wind speed at this site with decent accuracy. The model bias is relatively close to 0 m s\(^{-1}\), when compared to other AWS locations, for each forecast hour category and for
all of the 20 identified weather patterns. Conversely, the Pegasus North verification plot in Fig. 6b shows that the model has a strong positive wind speed bias at this location for nodes [1,1], [1,2], [2,1], [2,2], [3,1], and [3,2], which corresponds with southerly flow at Pegasus North. Furthermore, comparing the AWS observations by weather pattern for Pegasus North (Fig. 6a) and Willie Field (Fig. 8a), it can be seen that the observed wind speeds at the two locations are quite similar, indicating that for a given weather pattern Pegasus North experiences wind speeds similar to those at Willie Field. Figure 3 of Monaghan et al. (2004) shows that the Pegasus North AWS is located, according to the model, in an area of slightly stronger average wind speeds than those predicted at Willie Field. Therefore, for weather patterns with southerly flow (nodes [1,1], [1,2], [2,1], [2,2], [3,1], and [3,2]), it appears that the model is under-representing the blocking effects of Ross Island at the location of the Pegasus North AWS. This type of information provided by the weather-pattern-based method of model evaluation can be useful for model developers when making improvements to the model.
Fig. 8. As in Fig. 6, but for the Willie Field AWS location.

The wind direction at the Vito AWS is another example of where the weather-pattern-based model evaluation technique is beneficial in evaluating the performance of the model for different weather patterns. Figure 9b shows the average wind direction verification plot for the Vito AWS (VTO in Fig. 3b). Analyzing the average model forecasted wind direction from Fig. 9b, it can be seen that on average AMPS predicts a fairly consistent wind direction of approximately 190° for each weather pattern. The AWS observations
(shown in Fig. 9b by the black, dotted line) indicate that the average wind direction varies more than the modeled wind direction and that the wind direction varies clearly as a function of weather pattern. For instance, by inspecting the bottom row of the node-averaged sea level pressure fields of the master SOM (nodes [1,4], [2,4], [3,4], and [4,4] in Fig. 4) it can be seen that by moving across the row the cyclone shifts from the northeast corner to the southeast corner of the domain. This would locate the cyclone to the east of the Vito AWS in node [1, 4], causing a southwesterly wind to be expected at Vito for this weather pattern. Moving through the other weather patterns associated with the bottom row of the SOM, the cyclone progressively shifts to the southeast of the Vito AWS location.
Therefore, the winds at Vito would be expected to become more westerly and eventually northwesterly as this progression is made. The AWS observations (shown by the black, dotted line) in the wind direction verification plot (Fig. 9b) indicate that the wind direction associated with the nodes from the bottom row of the SOM does in fact shift from a southwesterly direction in the bottom-left corner of the master SOM (node [1, 4]) to a
more westerly direction in the bottom center of the master SOM (node [3,4]) and eventually a northwesterly direction in the bottom-right corner of the master SOM (node [5,4]). The AMPS forecasts do no indicate this progression of wind direction as the synoptic patterns changes, but it should be noted that in nodes [4,4] and [5,4] where northwesterly wind would be expected, and where the strong wind direction bias exists, the pressure gradient is extremely weak. This weak pressure gradient results in lower wind speeds (see Fig. 9a) and thus less predictable wind directions. This, in part helps explain the large wind direction biases, but it appears that the model is still not adequately representing the impacts of the varying direction of the pressure gradient for these patterns. In this example, the use of the weather-pattern-based method of model evaluation provides us with the ability to easily identify the weather patterns that the model will forecast the wind direction with good accuracy and the weather patterns for which the model will struggle with the accuracy of the wind direction forecasts.

The weather-pattern-based model evaluation technique can also be used to find areas of interesting phenomenon for further research. For instance, at the Cape Bird AWS (CBD in Fig. 3b), the SOM analysis can be used to investigate potential connections between model errors for different weather parameters (e.g., pressure, wind speed, etc.). Figure 10 shows the average pressure and average wind speed verification plots for Cape Bird AWS. The average pressure plot indicates that the model overpredicts the pressure at Cape Bird for each of the identified weather patterns (the AWS observations are shown by the black, dotted lines and the forecasts are shown by solid lines). The plot also indicates that the forecast pressure decreases for the longer forecast hours. The average wind speed plot
indicates that the model predicts the wind speed fairly well for the shorter forecast hours, but greatly overpredicts the wind speed for the longer forecast hours. For example, looking at node [1, 1], the forecast pressure drops from 987.5 mb during the 0–9-h forecast to 979.8 mb during the 60–69-h forecast with the AWS observations averaging at 977.6 mb. Doing a similar analysis on the wind speed, shows that the forecast wind speed increases from 6.4 m s⁻¹ during the 0–9-h forecast to 13.9 m s⁻¹ during the 60–69-h forecast with the AWS observations averaging at 4.6 m s⁻¹.
Fig. 10. (a) Verification plot of average pressure for the Cape Bird AWS location calculated for the AWS observations (black dotted line) and the following forecast categories: 0-9 (black solid line), 12-21 (red solid line), 24-33 (orange solid line), 36-45 (green solid line), 48-57 (light blue solid line), and 60-69 (dark blue solid line). (b) Verification plot of average wind speed for the Cape Bird AWS location calculated for the AWS observations (black dotted line) and the same forecast categories (same lines and colors) as presented in (a).

A possible explanation for the change in the model output over forecast time could be a connection between the decrease in model pressure with time and an increase in the model wind speed with time. For instance, Cape Bird is located on the leeward side of Ross Island for weather patterns with southerly flow (nodes in the first three columns of the master SOM). Powers et al. (2003) discusses the tendency for von Karman vortices to spin
up in the lee of Ross Island under conditions of strong southerly flow. Therefore, the presence of these vortices as the model is given time to spin up could potentially explain the strong decrease in model pressure and strong increase in model wind speed with increased forecast time, especially for weather patterns with southerly flow. Although further research would need to be conducted in order to determine the cause of the model errors associated with this scenario. Therefore, this is an example where the SOM evaluation method does not provide all of the information necessary to understand why the model performs better in some instances than others, but can indicate areas where further research could help us to better understand the model results and the atmospheric dynamics in the region.

4. Conclusions

In this study the model in AMPS was evaluated using a weather-pattern-based method of model evaluation. The purpose of the study was to determine the benefits of using a weather-pattern-based method of model evaluation. The analysis involved using SOMs to identify 20 weather patterns that represent the range of weather patterns that occur over the RIS. Subsequently, each model forecast was associated with one of the 20 identified weather patterns in order to evaluate the model performance for each weather pattern. AWS observations of pressure, wind speed, and wind direction were used to validate the model forecasts and statistics, such as average wind speed and model wind speed bias, were calculated for various locations on the RIS. The results showed instances
where model performance was a function of the weather pattern, confirming the weather-pattern-based method of model evaluation as a useful evaluation technique. Furthermore, it was also shown that the model performance as a function of weather pattern is dependent on the atmospheric variable (pressure, temperature, wind speed, or wind direction) and on the location of the area being analyzed. The model dependency on weather pattern for wind speed in one location did not imply model dependency on weather pattern in a different location. Therefore, the weather-pattern-based method of model evaluation provides specific results for each individual atmospheric variable of interest and for each individual location of interest. It was also determined that the weather-pattern-based method of model evaluation has the ability to identify large model biases that may go undetected when using other methods of model evaluation. In conclusion, the weather-pattern-based method of model evaluation presented here has the ability to identify model errors as a function of weather pattern and therefore has the ability to provide model developers and users with additional guidance regarding model performance.

5. Acknowledgments

NSF Grants ANT-0636811 and ATM-0404790 supported this work. The AMPS data were retrieved from the National Center for Atmospheric Research’s Computational and Information Systems Laboratory. The AWS data were retrieved from the Automatic Weather Station Program and Antarctic Meteorological Research Center (Matthew Lazzara,
Chapter 3: Case Study of a Barrier Wind Corner Jet off the Coast of the Prince Olav Mountains, Antarctica

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This article appears in the July 2012 issue of Monthly Weather Review Volume 140, Number 7 published by American Meteorological Society and available online through AMS Journals. http://dx.doi.org/10.1175/MWR-D-11-00261.1

Abstract: The Ross Ice Shelf airstream (RAS) is a barrier parallel flow along the base of the Transantarctic Mountains. Previous research has hypothesized that a combination of katabatic flow, barrier winds, and mesoscale and synoptic-scale cyclones drive the RAS.
Within the RAS, an area of maximum wind speed is located to the northwest of the protruding Prince Olav Mountains. In this region, the Sabrina automatic weather station (AWS) observed a September 2009 high wind event with wind speeds in excess of 20 m s\(^{-1}\) for nearly 35 h. The following case study uses in situ AWS observations and output from the Antarctic Mesoscale Prediction System to demonstrate that the strong wind speeds during this event were caused by a combination of various forcing mechanisms, including katabatic winds, barrier winds, a surface mesocyclone over the Ross Ice Shelf, an upper-level ridge over the southern tip of the Ross Ice Shelf, and topographic influences from the Prince Olav Mountains. These forcing mechanisms induced a barrier wind corner jet to the northwest of the Prince Olav Mountains, explaining the maximum wind speeds observed in this region. The RAS wind speeds were strong enough to induce two additional barrier wind corner jets to the northwest of the Prince Olav Mountains, resulting in a triple barrier wind corner jet along the base of the Transantarctic Mountains.

1. Introduction

The Ross Ice Shelf (RIS) airstream (RAS), a persistent wind over the Ross Ice Shelf, Antarctica, originates in the Siple Coast confluence zone (Parish and Bromwich 1986), flows parallel to the base of the Transantarctic Mountains and eventually toward the north over the Ross Sea (Fig. 11). The general path of the RAS is constrained by the topography of the Transantarctic Mountains and is therefore fairly consistent throughout the year (Parish et al. 2006). Parish et al. (2006) identified three areas of maximum wind speed within the
RAS. These maxima are located off the coast of the Prince Olav Mountains, south of Ross Island, and off the coast of Cape Adare (Fig. 11). The following case study focuses on the formation of the wind speed maximum off the coast of the Prince Olav Mountains.

![Map of the RIS region](image)

Fig. 11. Map of the RIS region. The location of the Sabrina AWS is labeled as SAB and the location of the Eric AWS is labeled ERC. The black dots indicate the location of the cross section presented in Fig. 19. The letters A and B indicate the orientation of the cross section.

The katabatic winds in the region of the RIS contribute to the formation of the RAS (Parish et al. 2006). Katabatic winds are a gravity-forced phenomenon where negatively buoyant air flows downward over sloping terrain. In the area of the RIS, these winds flow down the glacial valleys of the Transantarctic Mountains and through the large Siple Coast confluence zone located at the southern tip of the RIS (Parish and Bromwich 1987). However, because of the seasonal cycle of the katabatic winds and the consistent presence of the RAS throughout the year, something other than katabatic winds must be the
dominant forcing associated with the RAS. Parish et al. (2006) concludes that the barrier winds along the Transantarctic Mountains, forced by the semipermanent low pressure system located in the eastern Ross Sea, primarily drive the RAS.

Several studies have looked at the process of barrier winds forming along the Transantarctic Mountains over the RIS (O'Connor et al. 1994; Parish et al. 2006; Seefeldt et al. 2007; Steinhoff et al. 2009). Barrier winds occur when flow in a stable atmosphere is directed toward the base of a barrier: in this case, the Transantarctic Mountains. If the atmosphere is stably stratified and the Froude number of the approaching flow is less than one, the barrier blocks the flow and mass convergence occurs (O'Connor et al. 1994; Buzzi et al. 1997). This alters the sea level pressure (SLP) field creating a local high pressure region at the base of the mountains. The pressure decreases away from the mountains, causing a pressure gradient force (PGF) directed perpendicular to and away from the mountains. The winds induced by this PGF approach geostrophic balance over time and induce a wind that is parallel to the base of the mountains. In this case, the winds blow toward the northwest and run along the base of the Transantarctic Mountains in this region (Parish et al. 2006).

The RAS and its components are important to study because of the impact of the RAS on the atmospheric circulation of the Southern Hemisphere. Specifically, the RAS transports energy and momentum from the Antarctic continent toward the north. For example, Parish and Bromwich (1998) analyzed a case study involving a large drop in the pressure over the Antarctic continent. The mass transport associated with this drop in
pressure was driven by katabatic drainage of mass off of the continent. Approximately one-third of the mass drained from the continent passed through the Siple Coast confluence zone.

The RAS also affects the local atmosphere over the RIS and in the region of the Ross Sea. During RAS events, when high wind speeds occur along the base of the Transantarctic Mountains, the signature of the winds can be seen up to 1000 km away from the source regions. In fact, the strength of the RAS influences the size of the RIS polynya that is located off the northern edge of the ice shelf. The size of the polynya has implications for sensible and latent heat flux exchange with the Ross Sea in this region (Bromwich et al. 1992).

During RAS events, a wind speed maximum forms along the coast of the Prince Olav Mountains (a subset of the Queen Maud Mountains). Parish et al. (2006) used output from the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) run within the Antarctic Mesoscale Prediction System (AMPS) to estimate mean annual wind speeds of approximately 20 m s\(^{-1}\) (at an approximate height of 300 m) in this region. The Prince Olav Mountains, as shown in Fig. 11, protrude onto the RIS and into the path of the RAS. Previous studies (Seefeldt and Cassano 2008; Steinhoff et al. 2009) hypothesize that the topographic influences associated with the protrusion provide the forcing for the development of the wind speed maximum in this region.
Steinhoff et al. (2009) referred to this area of maximum wind speed as a “knob flow,” or a corner wind (Dickey 1961; Kozo and Robe 1986; Olafsson 2000; Olafsson and Agustsson 2007; Lefevre et al. 2010). A corner wind is an asymmetric flow around an obstacle, or barrier. In the Northern (Southern) Hemisphere, the majority of the flow passes on the left (right) side of the barrier (when looking downstream with the flow), whereas a minimal amount of the flow passes on the right (left) side of the barrier. This imbalance is caused by the blocking of stably stratified air by the barrier, the creation of a terrain-induced region of high pressure, and the subsequent presence of a barrier wind on the upwind side of the barrier. Because of the balance between the PGF and the Coriolis force (CF), the barrier wind will flow around the left (right) side of the obstacle in the Northern (Southern) Hemisphere. Additionally, once the barrier wind reaches the end of the barrier, the flow accelerates as it transitions from the area of terrain-induced high pressure to the background pressure field of the region. During this transition, the PGF aligns with the direction of flow and increases the magnitude of the wind speed in this region, causing the wind speeds on the left (right) side of the obstacle in the Northern (Southern) Hemisphere to be stronger than the wind speeds on the right (left) side of the obstacle. Barstad and Gronas (2005) refer to the strong corner winds on the left side of the obstacle (in the Northern Hemisphere) as a left-side jet.

Seefeldt and Cassano (2008) referred to the area of maximum wind speed off the coast of the Prince Olav Mountains as the Queen Maud Mountains tip jet. Tip jets have been studied off the coast of Greenland. The Greenland easterly tip jet (Moore and Renfrew 2005; Renfrew et al. 2009; Outten et al. 2009) is driven by the same forcing mechanisms as
the Northern Hemisphere left-side jet (Barstad and Gronas 2005). The Greenland easterly tip jet occurs when a low pressure system is located to either the south or southeast of Greenland. The positioning of the low pressure system directs low-level flow toward the southeastern coast of Greenland (Fig. 12). In an atmosphere that is stably stratified, the high terrain of the southern tip blocks the flow from traversing up and over the mountain range. This creates a buildup of mass along the windward coast of the barrier and induces a barrier wind. In this case, the barrier wind flows toward the southwest (left of the original upstream wind direction), parallel to the coast. As the barrier wind reaches the end of the barrier, the tip of Greenland, the terrain-induced PGF perpendicular to the mountains no longer exists (Moore and Renfrew 2005). In fact, at the end of the barrier, the PGF aligns with the direction of flow (Fig. 12). This is due to the transition from a region of high pressure induced by the terrain to a region of pressure free from topographic influences. The alignment of the PGF with the direction of the barrier wind at this point increases the magnitude of the wind speed and creates an area of maximum wind speed downstream of the barrier (Outten et al. 2009). Additionally, at this location, the PGF no longer balances the CF and the winds are turned toward the right (in the Northern Hemisphere). The combination of the acceleration of the winds by the PGF and the turning of the winds by the CF creates the easterly tip jet downstream of the tip of Greenland (Moore and Renfrew 2005) and is consistent with the dynamics of corner wind discussed above.
This study analyzes a high wind event off the coast of the Prince Olav Mountains, where a corner wind forms on the right side (in the Southern Hemisphere) of the obstacle. This jet forms when the barrier wind reaches the end of the barrier and is accelerated through the transition from the terrain-induced high pressure to the background pressure field of the region. The dynamics of this jet are the same dynamics associated with the Northern Hemisphere left-side jet and the Greenland easterly tip jet. Because the term “left-side jet” is specific to the Northern Hemisphere and the term “easterly tip jet” is specific to Greenland or other regions where there is a clearly defined tip in the topography, the term “barrier wind corner jet” (BWCJ) will be used to refer to the jet that forms when a barrier wind reaches the end of the barrier and the flow adjusts from the area of terrain-induced high pressure to the large-scale pressure field of the region. This term is inclusive of either left-side (Northern Hemisphere) or right-side (Southern Hemisphere) corner jets.
and indicates the critical role of the barrier wind in defining whether the jet is left or right sided.

2. Data and methods

2.1 Model

Because of the scarcity of observations in Antarctica, this case study uses a combination of observations and numerical weather prediction (NWP) forecasts from AMPS. The National Center for Atmospheric Research, in collaboration with The Ohio State University, originally created AMPS as an experimental real-time NWP system for use in the Antarctic by the United States Antarctic Program and other national Antarctic research programs (Powers et al. 2003).

AMPS is based on the polar-modified version of the Weather Research and Forecasting Model (WRF; Hines and Bromwich 2008; Bromwich et al. 2009; Hines et al. 2011), which has been optimized for use in the Antarctic. The major changes made to the model include implementation of a scheme to treat fractional sea ice, improved treatment of heat transfer through ice and snow surfaces, a revised surface energy balance calculation, and selection of model physics options that are appropriate for polar applications.
AMPS is set up with a set of six two-way nested domains. The domains are numbered 1–6 with horizontal grid spacings of 45, 15, 5, 5, 1.67, and 5 km, respectively (Fig. 2). The vertical grid includes 44 eta levels. The first-guess initialization and the lateral boundary conditions are taken twice daily by the National Centers for Environmental Prediction 0.5° Global Forecast System model output. The system assimilates observations using the WRF three-dimensional variational data assimilation capability. Table 1 lists the physical parameterizations used in AMPS. Additional information on AMPS can be found in Powers (2007) and on the AMPS website (http://www.mmm.ucar.edu/rt/wrf/amps/).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Scheme</th>
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</thead>
<tbody>
<tr>
<td>Longwave radiation</td>
<td>Rapid Radiative Transfer Model (RRTM) radiation</td>
</tr>
<tr>
<td>Shortwave radiation</td>
<td>Goddard shortwave radiation</td>
</tr>
<tr>
<td>Boundary layer</td>
<td>Mellor–Yamada–Janjic (Eta)</td>
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<td></td>
<td>turbulent kinetic energy (TKE)</td>
</tr>
<tr>
<td>Surface layer</td>
<td>Monin–Obukhov (Janjic Eta)</td>
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<tr>
<td>Land surface option</td>
<td>Unified Noah land surface model</td>
</tr>
<tr>
<td>Microphysics</td>
<td>WRF Single-Moment 5-Class (WSM5)</td>
</tr>
<tr>
<td>Cumulus parameterization</td>
<td>Kain–Fritsch (new Eta)</td>
</tr>
<tr>
<td>Sea ice</td>
<td>Fractional sea ice</td>
</tr>
<tr>
<td>Upper boundary condition</td>
<td>Model top at 10 mb</td>
</tr>
</tbody>
</table>

Table 1. A list of the physical parameterizations in AMPS and the upper boundary condition.

The following case study uses the AMPS output from domain 2 (15-km grid spacing) for the analysis. Domain 2 covers the Antarctic continent and provides forecasts of both the local and regional atmosphere for this case study. AMPS domain 2 is run for a total of 120 h with the output archived every 3 h. The case study uses the 12–21-h forecasts from the AMPS output. The 0–9-h forecasts are not used so as to provide ample spinup time for the model to adjust from the initial conditions. This follows the approach used by Bromwich et
al. (2005) and Seefeldt and Cassano (2008). Because of the 3-hourly output archive, the 12–21-h forecasts provide a continuous time series for the duration of the case study. The case study analyzes spatial plots of variables, such as pressure, wind speed, and temperature, and calculations of dynamic variables, such as the Froude number and the individual terms of the horizontal momentum equations.

The Froude number is a term used to describe how an atmospheric flow will interact with an obstacle and is a ratio of the kinetic energy of a parcel to the potential energy to be overcome by crossing a barrier of a specific height. The Froude number is dependent on the speed of the flow, the height of the obstacle, and the stability of the flow. If the Froude number is less than one, the flow lacks sufficient kinetic energy to pass over the barrier and is blocked by the obstacle. When this occurs, atmospheric mass accumulates on the upwind side of the barrier. Conversely, if the Froude number is greater than one, the flow has enough kinetic energy to pass over the obstacle (O’Connor et al. 1994; Buzzi et al. 1997).

The Froude number is defined by the following equation:

\[ Fr = U \times (gH \frac{\Delta \Theta}{\Theta})^{-0.5} \]  

(1)

where \( U \) is the speed of the flow directed toward the obstacle, \( g \) is gravity, \( H \) is the vertical distance the flow must travel to pass over the obstacle (the difference between the height
of interest in the flow and the top of the barrier), and $\Delta \Theta / \Theta$ is the static stability parameter (O'Connor et al. 1994). The $gH \Delta \Theta / \Theta$ is proportional to the potential energy required to lift a parcel that is $\Delta \Theta$ cooler than the environment through a height of $H$ and $U$ is proportional to the square root of the parcel kinetic energy. Because the parcel does not have a constant $\Delta \Theta$ relative to the environment between its initial height and the height of the barrier, we use

$$Fr = U \times \left( \frac{g}{\Theta'} \sum_{z} (\Theta' - \Theta) \Delta z \right)^{-0.5}$$

(2)

where, for each grid point in the cross section, $U$ is the wind speed directed toward the mountains, $g$ is the acceleration due to gravity, and $\Theta'$ is the potential temperature of the air parcel. The summation is then evaluated for each model eta level from the initial height of the air parcel to the height of the mountain (which is taken to be 2852 m). Within the summation, ($\Theta' - \Theta$) is the difference between the potential temperature of the air parcel and the environmental potential temperature of the layer (the average potential temperature of the layer is used) and $\Delta z$ is the vertical thickness of the layer.

In the following case study, the Froude number is calculated for a vertical cross section that is perpendicular to the mountain range (Fig. 11 shows the location of the cross section) to determine what portion of the flow is blocked by the Transantarctic Mountains. Within the cross section, the Froude number is calculated for each model grid point that
has a component of the wind that is directed toward the mountains and allows for the identification of air parcels that are blocked (Fr < 1) or can pass over the barrier (Fr > 1).

The case study also evaluates the individual terms of the horizontal momentum equations. These terms are not included as AMPS output variables and therefore must be calculated from the AMPS output. The calculations are based on the fundamental horizontal momentum equations given in Holton (2004). The terms of interest for this case study include advection, pressure gradient (PG), Coriolis, acceleration, and a residual term.

The terms of the horizontal momentum equations are calculated for each model grid point over the RIS, providing spatial information on the atmospheric forcing over the RIS during the case study. Grid points with an elevation greater than 500 m are not included in the calculations to avoid errors associated with calculating horizontal derivatives in the vicinity of steeply sloping terrain (i.e., the Transantarctic Mountains). The terms containing partial derivatives, such as the advection term and the PG term, are solved using a two-point centered finite differencing calculation. This calculation uses the four adjacent model grid points to evaluate the derivative. Additionally, for the PG term, all of the pressure values used in the calculation are interpolated to a common elevation using the hypsometric equation to provide better accuracy in the calculation of the derivative. The acceleration term is calculated using finite differencing over time.
### 2.2 Observations

Throughout the case study, AMPS forecasts are compared to observations to evaluate the AMPS forecasts and provide confidence for using AMPS to diagnose the dynamics of this event. Observations from the Sabrina automatic weather station (AWS) and the Eric AWS (Fig. 11) are used to verify the pressure, temperature, wind speed, and wind direction forecasts from AMPS. Each AWS takes measurements of temperature, pressure, wind speed, and wind direction at an interval of 10 min. These data are transmitted from the AWS in real time to satellite using the ARGOS system. The data are retrieved, quality controlled, and archived by the Antarctic Meteorological Research Center (AMRC) at the University of Wisconsin—Madison. The quality control procedure is a semiautomatic process that involves removing erroneous data points from the time series. Subsequently, AMRC creates 10-min, 1-hourly, and 3-hourly datasets of the observations. The following case study uses the 1-hourly and 3-hourly datasets, which are created by extracting the observation that is closest to the respective 1-hourly or 3-hourly time within a ±40-min time limit (Keller et al. 2010). The 3-hourly dataset provides an observational dataset with the same time resolution as the 15-km AMPS output used in this case study.

To provide the most accurate comparison between the AMPS forecasts and the AWS observations, adjustments are made to the model output prior to the evaluation. Specifically, the AMPS surface pressure is adjusted to the elevation of the AWS using the hypsometric equation (Holton 2004), and the AMPS 10-m wind speed is adjusted to the height of the AWS using the logarithmic wind profile equation (Holton 2004). The wind
calculation uses a roughness length of 0.0001 m, the same value used within AMPS, and neglects the stability correction term. The AMPS 10-m winds are adjusted to a height of 2.88 m for the Sabrina AWS and 1.57 m for the Eric AWS. These adjustments allow for a reasonable comparison between the model output and the AWS observations.

The case study also uses infrared satellite imagery to verify the presence of cyclonic systems and frontal zones in the AMPS forecasts. Specifically, the Antarctic satellite composite imagery retrieved from the AMRC archive at the University of Wisconsin—Madison (Lazzara et al. 2011) is used for the model verification. The composites are generated using data from multiple swaths of polar orbiting and geostationary satellites. Because of the limited satellite coverage in the polar regions, the composites provide temporal and spatial coverage that is not available from individual data sources. For more information on the creation of the composite imagery, see Lazzara et al. (2011).

3. Case study

The Sabrina AWS was installed off the coast of the Prince Olav Mountains in February 2009 to observe and validate the AMPS-modeled wind speed maximum in this region of the RAS (Parish et al. 2006; Seefeldt and Cassano 2008). The 1-hourly wind speed observations from February 2009 through April 2011 indicate that high wind events occur in this region on a fairly regular basis. These observations indicate that 29 high wind events, classified as winds in excess of 15 m s$^{-1}$ for at least 12 h with a maximum of one
observation below the 15 m s\(^{-1}\) threshold, occurred during this time period. The average maximum wind speed for these high wind events was 22 ms\(^{-1}\), and the average duration was 23 h. The maximum wind speed for these events was 28.2 m s\(^{-1}\), and the maximum duration of an individual event was 49 h. A category for extreme high wind events was also analyzed. The criteria for this category included winds in excess of 20 m s\(^{-1}\) for at least 12 h with a maximum of one observation below the 20 m s\(^{-1}\) threshold. During this time period, seven extreme high wind events occurred. The average maximum wind speed for these events was 25.7 m s\(^{-1}\), and the average duration was 22.5 h. The maximum wind speed during these events was 28.2 m s\(^{-1}\), and the maximum duration of an individual event was 35 h. The analysis also determined that the extreme high wind events only occurred during the non-summer months (March–September).

The following case study analyzes a high wind event from September 2009. The Sabrina AWS wind speed observations for this time period (shown by the solid line in Fig. 13c) indicate the high wind event began on 0000 UTC 6 September 2009. During the event, wind speeds were in excess of 20 m s\(^{-1}\) for approximately 35 h with the maximum wind speed reaching 27.4 m s\(^{-1}\). Therefore, this event fits the category of an extreme high wind event based on the previous analysis. The dashed line in Fig. 13c shows the AMPS 10-m wind speed forecasts for the same time period, indicating that AMPS predicted both the timing and magnitude of this high wind event with good accuracy. Additionally, Figs. 13a,b,d show the pressure, temperature, and wind direction comparisons, respectively. These observations also indicate that AMPS modeled the high wind with good accuracy.
Given the good agreement between the observations and the AMPS output, AMPS forecasts are used to analyze the dynamics of this event.

Fig. 13. Sabrina AWS observations and AMPS forecasts of (a) pressure, (b) temperature, (c) wind speed, and (d) wind direction for a portion of September 2009. In (a)-(c), the AWS observations are shown by solid lines and the AMPS forecasts are shown by dashed lines. In (d), the AWS observations are shown using circles and the AMPS forecasts are shown using stars.
To understand the development and progression of the high wind event observed at Sabrina, the case study will analyze the state of the atmosphere both prior to and during the high wind event. Specifically, the study will evaluate the development of the RAS event, a precursor to the high winds observed at Sabrina. As discussed in the introduction, the main components providing the forcing for the RAS event include the katabatic drainage through the Transantarctic Mountains and the Siple Coast confluence zone and the barrier winds along the base of the Transantarctic Mountains. Once the development of the RAS event has been explained, the case study will investigate the high wind event observed at Sabrina. Specifically, the topographic influences involved in the formation of the area of maximum wind speed to the northwest of the Prince Olav Mountains will be studied.

3.1 0300 UTC 5 September 2009: 21 h prior to wind event

Figures 14a,b show the AMPS forecasts of SLP and 10-m winds from 0300 UTC September 2009, 21 h prior to the onset of the high wind event. At this time, the SLP field showed a weak PG on the western RIS. On the eastern RIS the SLP contours were fairly straight with a PGF directed toward the east. The 10-m winds over the RIS were fairly light and southerly, while katabatic drainage flowed down the glacial valleys of the Transantarctic Mountains (the warm signatures associated with the katabatic drainage can be seen in Fig. 14d). The wind speeds at the Sabrina AWS were light and there was no evidence of the high wind event in the region of the Prince Olav Mountains at this time.
For this same time period, the AMPS-forecast SLP (Fig. 14a) indicates that a synoptic-scale low pressure system was located off the coast of West Antarctica. Figure 14c shows the infrared satellite image for 0400 UTC 5 September 2009, the closest satellite image available for comparison to the AMPS forecasts. The satellite image shows the cloud
signature associated with the low pressure system off the coast of West Antarctica, validating the AMPS forecast for the presence of the cyclone in this region. The AMPS forecasts and satellite images prior to this time period were analyzed to understand the origin of this cyclone. The AMPS forecasts and the infrared satellite images (not shown) indicate that the synoptic-scale low pressure system traversed from the west, over the Ross Sea, and into the region of the West Antarctic coast, which is a fairly typical cyclone track for this region (Lamb and Britton 1955; Jones and Simmonds 1993).

The movement of the cyclonic system into the region of the eastern Ross Sea caused the advection of warm maritime air onto the West Antarctic Plateau. Prior to the advection of warm air into this region, the AMPS-forecast 2-m temperatures over the plateau ranged from approximately -58° to -40°C. Because of the positioning of the cyclone, the warm maritime air overran the cold air over the plateau, creating a warm front. As the warm front moved across the plateau, the AMPS-forecast 2-m temperatures increased and reached up to -14°C in some regions of West Antarctica. The warmer temperatures and the strong temperature gradient associated with the warm front can be seen in the plot of the AMPS-forecast 2-m temperatures (Fig. 14d). Additionally, the enhanced cloud signature associated with the cyclone and frontal band over West Antarctica can be seen in the satellite infrared image (Fig. 14c). This warm front is important to the case study because it will provide the baroclinic forcing for the development of a mesocyclone over the RIS. This mesocyclone will strengthen the RAS event by enhancing both the katabatic drainage onto the RIS and the barrier winds along the Transantarctic Mountains.
3.2 1200 UTC 5 September 2009: 12 h prior to wind event

The AMPS-forecast SLP plot (Fig. 15a) for 1200 UTC 5 September 2009, 12 h prior to the onset of the high wind event, shows that the pressure over the RIS has decreased during the previous 9 h. Analysis of the SLP plots during these 9 h (not shown) indicate that the synoptic cyclone off the coast of West Antarctica strengthened, causing the decrease in pressure over the ice shelf. This pressure drop (shown by the dashed line in Fig. 13a) is validated by the Sabrina AWS pressure observations (shown by the solid line in Fig. 13a), which indicate that a large drop in pressure, approximately 34 mb, occurred between 4 and 6 September 2009. The decrease in pressure over the RIS during this time period increased the PG between the RIS and the East and West Antarctic Plateaus, synoptically enhancing the katabatic drainage through the glacial valleys of the Transantarctic Mountains (Fig. 15b). This also induced a downslope flow from West Antarctica into the region of the Siple Coast confluence zone (Fig. 15b). The enhanced katabatic drainage onto the RIS indicates the initiation of the RAS event for this case study.

During this time period, the synoptic cyclone remained off the coast of West Antarctica and in a favorable position to continue warm air advection onto the West Antarctic Plateau. Analysis of the 2-m temperature forecasts (not shown) and the infrared satellite imagery (not shown) indicates that the warm front continued to traverse over the West Antarctic Plateau toward the RIS during this time period.
Fig. 15. AMPS forecast of (a) SLP and (b) 10-m winds for 1200 UTC 5 Sep 2009. The black dots in the AMPS plots indicate the location of the Sabrina and Eric AWS locations.

### 3.3 2100 UTC 5 September 2009: Barrier wind development

The onset of the barrier wind began on 2100 UTC 5 September 2009, 3 h prior to the high wind event. The barrier wind was induced by an area of high pressure, which formed to the southeast of the Prince Olav Mountains (region A in Fig. 16a). The region of high pressure to the southeast of the Prince Olav Mountains was formed by the presence of an upper-level ridge over the southern tip of the RIS and winds directed toward and blocked by the Prince Olav Mountains. During the time period of the case study, an upper-level ridge moved over the southern tip of the RIS and contributed to the increase in the SLP to the southeast of the Prince Olav Mountains. Figures 17a–d show the evolution of the 500-mb geopotential heights from 0300 UTC 5 September to 0300 UTC 6 September 2009. Figure 17a shows an area of upper-level low geopotential heights located over the southern
tip of the RIS. Moving forward in time, Figs. 17b,c show an upper-level ridge over the southern tip of the RIS. The movement of the ridge into this region contributed to the increase in SLP to the southeast of the Prince Olav Mountains. Additionally, Fig. 17d indicates that for the next time period to be analyzed the upper-level ridge continues to move into the region of the southern RIS and provides further forcing for an increase in SLP to the southeast of the Prince Olav Mountains.

Fig. 16. As in Fig. 15, but 2100 UTC 5 Sep 2009.
Fig. 17. AMPS forecast of 500-mb geopotential heights for (a) 0300 UTC 5 Sep, (b) 1200 UTC 5 Sep, (c) 2100 UTC 5 Sep, and (d) 0300 UTC 6 Sep 2009. The black dots in the AMPS plots indicate the location of the Sabrina and Eric AWS locations.

The movement of the upper-level wave pattern, as shown in Figs. 17a–d, also contributed to the increase in SLP to the southeast of the Prince Olav Mountains by directing flow from the northeast toward the Trans-antarctic Mountains. The plots of the AMPS-forecast 700-mb geopotential heights and the 700-mb winds for 2100 UTC 5 September 2009 (Figs. 18a,b) show the 700-mb geopotential heights aligned to direct the 700-mb winds from the northeast toward the Transantarctic Mountains in the region of the
Siple Coast confluence zone. During this time period, the flow was stably stratified and the Transantarctic Mountains acted as a barrier, preventing the flow from traversing over the mountain range. This caused mass convergence at the base of the mountains and subsequently contributed to the development of the surface high pressure to the southeast of the Prince Olav Mountains. The surface high pressure induced a pressure gradient perpendicular and away from the mountains, creating a barrier wind in this region.

Figure 19 shows plots of the cross section parallel winds (Fig. 19a), perpendicular winds (Fig. 19b), Froude number (Fig. 19c), and potential temperature (Fig. 19d) in this region (Fig. 11 shows the location of the cross section) for this time period. In Fig. 19a, the solid shaded contours indicate the portion of the winds directed toward the mountains and the crosshatch contours indicate the portion of the winds directed away from the mountains. The plot indicates that the low-level flow was directed away from the mountains,
indicative of katabatic drainage from the Transantarctic Mountains. Conversely, the flow aloft, above approximately 1500 m, was directed toward the mountains. Figure 19b shows the cross section perpendicular winds, which are all into the page. The initial barrier jet is seen adjacent to the mountains, and it is apparent that the flow was accelerated around the Prince Olav Mountains rather than over the mountains. Additionally, the core of the jet was roughly centered on a height of approximately 850 m. Figure 19d shows the cross-sectional potential temperature, indicating a stable inversion layer over the RIS (distances >160 km) and weaker stratification indicative of enhanced mixing associated with the strong barrier winds seen in Fig. 19b in the region adjacent to the mountains. Figure 19c shows the results of the Froude number calculation. The Froude number was calculated only for points with flow directed toward the mountains. The results show a region between 1000 and 2400 m where the calculated Froude number was less than one, thus indicating that the flow in this region was blocked by the barrier (O’Connor et al. 1994; Buzzi et al. 1997). This blocking of the flow by the Transantarctic Mountains resulted in mass convergence and contributed to the formation of the surface high pressure in this region. Additionally, Fig. 19 illustrates the advantage to using model output for the case study analysis. The upper-level flow directed toward the mountains in this region would not have been identified using surface observations (the only in situ observations available in this region), because the surface flow was dominated by katabatic flow away from the mountains.
Fig. 19. (a) AMPS-calculated parallel wind speeds, (b) AMPS-calculated perpendicular wind speeds, (c) AMPS-calculated Froude number, and (d) AMPS-forecast potential temperature along the cross section (shown by the dotted line in Fig. 11) for 2100 UTC 5 Sep 2009. The letters A and B indicate the orientation of the cross section. In (a), positive winds (crosshatch contours) are directed away from the mountains (to the right) and negative winds (solid shaded contours) are directed towards the mountains (to the left). In (b) all winds are directed into the page.

The combination of the upper-level ridge and the flow directed toward the mountains caused the area of high pressure to the southeast of the Prince Olav Mountains that can be seen in the SLP plot shown in Fig. 16a. This surface high pressure induced a PGF perpendicular to and away from the Transantarctic Mountains, providing the forcing for
the barrier wind in this region. The start of the barrier wind can be seen in the 10-m wind plot in Fig. 16b and the further development of the barrier wind will be discussed in the next section.

3.4 0300 UTC 6 September 2009: Barrier wind strengthening

The AMPS-forecast SLP and 10-m wind plots (Figs. 20a,b) for 0300 UTC 6 September 2009 show a mesoscale low pressure system over the eastern RIS, near Siple Coast. The mesocyclone developed when the warm front from West Antarctica moved into the region of Siple Coast resulting in increased baroclinicity over the RIS. The infrared satellite imagery (not shown) confirmed the existence of this mesocyclone and the accuracy of the AMPS forecast for this time period. The mesocyclone over the RIS enhanced the PG between the RIS and the East and West Antarctic Plateaus, strengthening the katabatic drainage onto the RIS and the RAS event (Fig. 20b). The mesocyclone also enhanced the PG in the southeastern portion of the RIS. In fact, the region of high pressure to the southeast of the Prince Olav Mountains continued to increase over the previous 6 h, due to the further movement of the upper-level ridge into this region (Fig. 17d) and the persistent 700-mb flow directed toward the Transantarctic Mountains (not shown). Therefore, the strengthening of the high pressure region to the southeast of the Prince Olav Mountains combined with the decrease in pressure off of Siple Coast, due to the development of the mesocyclone, increased the PG in this region. The stronger PG enhanced the barrier wind along the base of the Prince Olav Mountains, as seen by the increased wind speeds off the
coast of the Prince Olav Mountains in Fig. 20b. Additionally, it can be seen that these strong winds have reached the location of the Sabrina AWS (Fig. 20b), initiating the start of the high wind event at Sabrina (Fig. 13c).

Fig. 20. As in Fig. 15, but for 0300 UTC 6 Sep 2009.

3.5 2100 UTC 6 September 2009: Barrier wind corner jets

By 2100 UTC 6 September 2009, the high wind event at the Sabrina AWS reached its peak (Fig. 13c). The SLP plot for this time period (Fig. 21a) indicates that the mesocyclone traversed over the RIS toward the northwest and is located over the northwestern RIS. The area of high pressure is still evident to the southeast of the Prince Olav Mountains.
Additionally, two areas of high pressure formed to the southeast of the Queen Alexander and to the southeast of the Churchill Mountains (Fig. 11) and can be seen in the plot of SLP at this time. The 10-m wind plot for this time period (Fig. 21b) shows a classic RAS event with strong winds flowing parallel to the base of the Transantarctic Mountains (Parish et al. 2006). Additionally, three areas of maximum wind speed can be seen within the RAS. These areas are located to the northwest of the Prince Olav, Queen Alexander, and Churchill Mountains. The area of maximum wind speed to the northwest of the Prince Olav Mountains is the focus of this case study; however, because of a connection between the formation of the three areas of maximum wind speed, all three jets will be analyzed.

Fig. 21. As in Fig. 15, but for 2100 UTC 6 Sep 2009.
The similarity in the patterns of the three areas of high pressure seen to the southeast of the mountain ranges and the three areas of maximum wind speed seen to the northwest of the mountain ranges (Figs. 21a,b) implies a possible connection between the two phenomena. To analyze this in more detail, the forces associated with the individual terms of the horizontal momentum equations were calculated for this region. Specifically, the model output was used to calculate the PG, Coriolis, advection, Eulerian acceleration, and residual terms for each grid point in the model domain. The PG and Coriolis results are shown spatially in Figs. 22c,d with the corresponding SLP and 10-m winds shown in Figs. 22a,b. Figure 23 shows the cross-jet and along-jet forces for the cross section depicted in Figs. 22a,b.

The PGF plot shown in Fig. 22c indicates that, to the southeast of the Prince Olav Mountains, a PGF was directed perpendicular and away from the mountains. In this same region, the CF (Fig. 22d) was directed perpendicular and toward the mountains. Although the magnitude of the PGF was larger than the magnitude of the CF (due to the effects of friction near the surface), the two forces approximately balanced each other and created a near-geostrophic wind parallel to the base of the mountains. This type of forcing is consistent with the presence of a barrier wind in this region, as discussed in the introduction. Additionally, the 10-m wind plot for this time period (Fig. 22b) shows the presence of the barrier wind off the coast of the Prince Olav Mountains.
Fig. 22. (a) AMPS forecast of SLP, (b) AMPS forecast of 10-m winds, (c) AMPS-calculated PGF and (d) AMPS-calculated CF for 2100 UTC 6 Sep 2009. In (a) and (b), the black arrows illustrate the pressure gradient and CFs and the numbered grid points depict the location of the cross section for Fig. 14. In (c) and (d), the black dots indicate the location of the Sabrina and Eric AWS locations.
Fig. 23. Individual forces of the horizontal momentum equations (a) across the jet (perpendicular to the flow) and (b) along the jet (parallel to the flow) for the cross section shown in Figs. 13a,b.

The SLP plot for 2100 UTC 6 September 2009 (Fig. 22a) indicates the area of high pressure was confined to the region to the southeast of the Prince Olav Mountains. In fact, at the point of maximum protrusion of the topography onto the RIS the PGF weakened and aligned with the direction of the flow (Figs. 22b,c). The change in the PGF was due to the transition from a region of high pressure induced by the terrain to a region of pressure free from topographic influences. The arrows in the plot of SLP (Fig. 22a) help to depict this transition. At the transition, the alignment of the PGF with the direction of the barrier wind increased the magnitude of the wind speed and created an area of maximum wind speed, a BWCJ, to the northwest of the Prince Olav Mountains. This is consistent with the mechanisms shown by Moore and Renfrew (2005) and Outten et al. (2009) for easterly tip jets near Greenland. The BWCJ can be seen in the plot of the 10-m winds in Fig. 22b. At this
same location, the CF was no longer balanced by the PGF and the flow was turned to the left (in the Southern Hemisphere), toward the base of the Transantarctic Mountains.

The line plots in Figs. 23a,b show the individual terms of the horizontal momentum equations along the cross section shown in Figs. 22a,b. The cross section aligns with the strongest 10-m winds within the jet (Fig. 22b). The 10-m winds were used for this analysis because the forcing for the barrier wind was strongest at the surface, the katabatic drainage that fed into the jet was strongest at the surface, and the observations used for model validation were from surface based AWS. At each of the model grid points, the x-y coordinate system was rotated such that the x axis aligned with the wind vector at that grid point and the y axis was oriented to the left of the wind vector. This allowed us to analyze the forces acting perpendicular (Fig. 23a) and parallel (Fig. 23b) to the jet and is similar to the method used by Outten et al. (2009). For the along-jet force balance (Fig. 23b) forces that are positive are directed in the direction of the wind vector, and for the cross-jet force balance (Fig. 23a) forces that are positive are directed to the left of the wind vector (toward the mountains).

In the analysis of the cross-jet forces (Fig. 23a), the PGF was negative (directed away from the mountains) and the CF was positive (toward the mountains) in the region from grid point 1 through grid point 8. At some grid points (2, 3, and 6) the CF roughly balanced the PGF, showing an approximate barrier wind at these locations (grid points 4 and 7 will be discussed below). At grid point 8, which is the approximate location of the area of maximum protrusion by the Prince Olav Mountains, the PGF became positive (toward the
mountains). This indicates the termination of the barrier wind in this region and is consistent with the barrier wind reaching the end of the barrier and transitioning to a different flow regime. The advection term was negative along the entire cross section but was generally small over the first seven grid points. The negative advection term indicates advection of flow with a smaller component toward the barrier [i.e., the flow turned toward the barrier (counterclockwise turning) in the along-jet direction]. The acceleration across the jet was negligible. The residual term was small from grid point 1 through grid point 8 and then increased in magnitude and was negative throughout the remainder of the jet.

In the analysis of the along-jet forces (Fig. 23b), the PG term was small for grid points 1, 2, 5, and 6. This weak along-jet PGF is consistent with the presence of a barrier wind where the pressure gradient force is dominantly perpendicular to the wind. Conversely, the along-jet PGF was large for grid points 4, 7, 8, and 9. At these grid points, the isobars (Fig. 22a) were oriented such that the pressure decreased in the along-jet direction. This pressure distribution was due to the jet passing out of the terrain-induced high pressure region producing a PGF that roughly aligned with the direction of flow (Fig. 22a) and accelerated the jet (Fig. 22b). In these regions of acceleration, the PGF was the only positive force in the along-jet analysis. Therefore, the alignment of the PGF with the direction of flow was the only mechanism that could increase the wind speeds within the jet.

In the along-jet analysis, the advection term was negative along the majority of the cross section, indicating that weaker upstream winds were being advected into the core of
the jet. At a few grid points, the advection term was near zero but slightly positive, indicating a weak along-jet gradient of wind speed. As required by the chosen coordinate system, the CF was zero. The residual term was negative (except for grid point 6) and primarily reflects the role of friction slowing the near surface winds. The larger residual at grid points 7, 8, 9, and 10 were likely due to the increase in friction as the wind speeds increased in this region of the jet. The acceleration along the jet was small compared to the other forces.

As mentioned, two additional areas of maximum wind speed were identified in the plot of the 10-m winds (Fig. 22b). These are related to the Prince Olav Mountain BWCJ because the flow from the Prince Olav Mountain BWCJ provided the forcing for the Queen Alexander Mountain BWCJ, which subsequently provided the forcing for the Churchill Mountain BWCJ. As described in the analysis of the Prince Olav Mountain BWCJ, the unbalanced CF to the northwest of the Prince Olav Mountains turned the flow to the left, toward the Transantarctic Mountains (Figs. 22b–d). The turning of the winds in this region directed the low-level flow toward the southeastern base of the Queen Alexander Mountains (Fig. 22b). Because of a stably stratified atmosphere, the protruding mountain range blocked the flow and caused a buildup of mass to the southeast of the Queen Alexander Mountains (Fig. 22a). As shown in Figs. 22a,c, this created an area of high pressure and a PGF directed perpendicular to and away from the mountains in this region. The PGF induced a barrier wind along the base of the Queen Alexander Mountains, similar to the formation of the barrier wind along the base of the Prince Olav Mountains. As the flow from the barrier wind passed the area of maximum topographic protrusion, the PGF weakened
and aligned with the direction of flow (Fig. 22c). The alignment of the PGF with the flow increased the magnitude of the wind speed and created a second BWCJ to the northwest of the Queen Alexander Mountains (Fig. 22b). Subsequently, to the northwest of the Queen Alexander Mountains the unbalanced CF turned the flow to the left and directed it toward the base of the Churchill Mountains. Following the same logic for the formation of the previous two BWCJs, it can be seen that a third BWCJ formed to the northwest of the Churchill Mountains.

The Eric AWS is located in the region to the northwest of the Churchill Mountains (Fig. 11) and can be used to validate the presence of the BWCJ in this region. The Eric AWS wind speed observations (shown by the solid line in Fig. 24) indicate that a high wind event started at 2100 UTC 6 September 2009, about 21 h after the start of the Sabrina AWS high wind event. This high wind event coincided with the presence of a BWCJ in the region at this time. The lag in the start of the high wind event at the Eric AWS from the start of the high wind event at the Sabrina AWS illustrates the sequential formation of each BWCJ along the base of the Trans-antarctic Mountains. Additionally, the AMPS wind speed forecasts (shown by the dashed line in Fig. 24) indicate that AMPS predicted both the timing and magnitude of this high wind event with good accuracy, further validating the use of AMPS forecasts for this case study.
4. Conclusions

This case study analyzed a September 2009 high wind event observed by the Sabrina AWS to better understand the area of maximum wind speed located to the northwest of the Prince Olav Mountains during RAS events. The results of the study indicate that the forcing associated with a BWCJ in the region to the northwest of the Prince Olav Mountains created the area of maximum wind speed. Additionally, the case study revealed the formation of two subsequent BWCJs along the Transantarctic Mountains, resulting in a triple BWCJ in this region.
The atmospheric conditions prior to the formation of the triple BWCJ included katabatic winds, barrier winds, a mesoscale surface low over the RIS, an upper-level ridge over the southern tip of the RIS, and topographic influences from the Transantarctic Mountains. The upper-level ridge and 700-mb flow directed toward the Transantarctic Mountains contributed to the formation of a high pressure region to the southeast of the Prince Olav Mountains. This high pressure region is the key component to the formation of the Prince Olav Mountain BWCJ. The presence of this high pressure region induced a barrier wind along the base of the Prince Olav Mountains. As this barrier wind reached the end of the barrier, the PGF became aligned with the direction of the flow and increased the magnitude of the wind speeds in this region. Subsequently, at this location the unbalanced CF turned the flow to the left and directed the flow toward the southeast side of the Queen Alexander Mountains. The mountains blocked the flow and created an area of high pressure to the southeast of the Queen Alexander Mountains. The area of high pressure induced a second barrier wind. When this barrier wind reached the end of the barrier, a second BWCJ formed to the northwest of the Queen Alexander Mountains. Subsequently, the same processes created a third BWCJ to the northwest of the Churchill Mountains.

The case study also highlighted the advantage to using a combination of observations and model forecasts to analyze the atmospheric dynamics in a region where observations are quite limited. Specifically, the AWS observations and satellite images were used to validate the accuracy of the model forecasts. This evaluation indicated that AMPS did a good job at predicting the timing and magnitude of the BWCJs at the Sabrina AWS and the Eric AWS, the large drop in pressure observed by the Sabrina AWS, the presence of a
synoptic-scale cyclone off the coast of West Antarctica, and the presence of a mesocyclone over the RIS. With this knowledge, the model forecasts were used to conduct a three-dimensional analysis of the atmosphere during the case study. This provided an additional level of detail that would not have been available through observations alone. Specifically, the 700-mb winds directed toward the Prince Olav Mountains, which provided forcing for the mass convergence and subsequent high pressure region to the southeast of the Prince Olav Mountains, would not have been detected through surface observations and satellite imagery. This was a key component to the development of the BWCJ and the conclusions from the case study would have been difficult to determine without the three-dimensional analysis of the model output.

Future work by the authors will analyze the different wind patterns, including BWCJs, over the Ross Ice Shelf. This research will investigate the frequency, seasonality, and forcing for each wind pattern. The results of the future study will provide climatologic information for the BWCJ and help to put the results presented in this case study of a high wind event into the broader context of the near-surface wind climate over the RIS.

5. Acknowledgments

The AMPS data were retrieved courtesy of the National Center for Atmospheric Research Computational and Information Systems Laboratory. This research is based upon
work supported by the National Science Foundation Office of Polar Programs under Grants ANT-0636811, ANT-0943952, ANT-0636873, ANT-0838834, and ATM-0404790.
Chapter 4: Analysis of high winds over the Ross Ice Shelf, Antarctica: Barrier winds along the Transantarctic Mountains

Abstract: The steep topography surrounding the Ross Ice Shelf, Antarctica greatly influences the wind patterns in the region of the Ross Ice Shelf. The topography provides forcing for features such as katabatic winds, barrier winds and barrier wind corner jets. The combination of topographic forcing and synoptic forcing from cyclones that traverse the Ross Sea, create a region of strong, but varying winds. This paper presents a surface wind climatology over the Ross Ice Shelf using output from the Weather Research and Forecasting model run within the Antarctic Mesoscale Prediction System. The dataset has 15 km grid spacing and is the first Ross Ice Shelf wind climatology presented at this resolution. The wind climatology shows the Ross Ice Shelf airstream, a dominant stream of air flowing northward from the interior of the continent over the western and/or central Ross Ice Shelf to the Ross Sea, is present over the Ross Ice Shelf approximately 34% of the time. A refined climatology of only the Ross Ice Shelf airstream patterns is also presented. This climatology indicates that the Ross Ice Shelf airstream varies in both its strength and position over the Ross Ice Shelf. These variations exist because Ross Ice Shelf airstream patterns can be driven by a variety of forcing mechanisms. The atmospheric dynamics associated with the barrier wind component of the Ross Ice Shelf airstream are analyzed in this paper.
1. Introduction

The steep topography that surrounds the Ross Ice Shelf (RIS), Antarctica greatly influences the wind patterns in that region (Fig. 25). The topography provides forcing for features such as the Ross Ice Shelf airstream (RAS), katabatic winds, barrier winds and barrier wind corner jets (BWCJ). The combination of these features with the synoptic forcing from cyclones that traverse the Ross Sea creates a region of strong, but varying winds.

Fig. 25. Map of the Ross Ice Shelf region. The shaded grid points indicate the model grid points used to train the SOM. The black dots indicate the AWS used in the statistical analysis of the AMPS output.
The RAS, the most common wind regime over the RIS, is a dominant stream of air flowing from the Siple Coast confluence zone, over the western to central RIS, to the north over the Ross Sea (Parish et al. 2006; Seefeldt and Cassano 2012). Satellite imagery shows the RAS propagates northward up to 1000 km away from the Siple Coast confluence zone (Bromwich et al. 1992; Bromwich et al. 1993). The RAS is driven by a combination of katabatic winds, barrier winds and forcing from the semi-permanent synoptic cyclone located off the coast of West Antarctica (Parish et al. 2006; Seefeldt and Cassano 2012). Parish et al. (2006) and Seefeldt and Cassano (2012) used the polar modified 30 km MM5 model run within the Antarctic Mesoscale Prediction System (AMPS) to study the forcing of the RAS. These studies concluded that the semi-permanent synoptic cyclone off the coast of West Antarctica provides the dominant forcing for the RAS.

Katabatic winds, another common wind in the region of the RIS, occur when negatively buoyant air flows downward over sloping terrain (Parish 1988). In general, katabatic winds are a key component of the Southern Hemisphere circulation. These winds drain the cold, continental air from the interior of the Antarctic continent, through confluence zones, to the edges of the continent (Parish and Bromwich 1987). This causes sinking motion over the interior of the continent and convergence aloft (Parish and Bromwich 1991; Parish 1992; Parish et al. 1994; Parish and Bromwich 1998). In the region of the RIS, katabatic winds flow through the glacial valleys of the Transantarctic Mountains and the Siple Coast confluence zone, providing forcing for the RAS.
Barrier winds commonly occur along the Transantarctic Mountains in the region of the RIS. Barrier winds form when stably stratified flow is directed towards a topographic barrier. Depending on the stability of the flow and the height of the barrier, this flow can be blocked. If the flow is blocked, mass convergence occurs and the pressure at the base of the barrier increases. The increase in pressure creates a pressure gradient force (PGF) directed away from the barrier. The wind induced by the PGF reaches approximate geostrophic balance, resulting in a barrier parallel flow at the base of the barrier (O’Conner et al. 1994; Parish et al. 2006; Seefeldt et al. 2007; Steinhoff et al. 2009). The barrier winds along the Transantarctic Mountains provide forcing for the RAS when the RAS is located adjacent to the Transantarctic Mountains.

BWCJs are commonly located to the northwest of the Prince Olav Mountains (Fig. 25). BWCJs form when a barrier wind reaches the end of a barrier. At this point, the flow transitions from a region of terrain induced high pressure to the background pressure field of the region and the PGF aligns with the direction of the flow, increasing the magnitude of the winds in this region. This results in an area of maximum wind speed downstream of the barrier (Nigro et al. 2012). BWCJs have similar forcing to Northern Hemisphere left-sided corner winds (Dickey 1961; Kozo and Robe 1986; Olafsson 2000; Olafsson and Agustsson 2007; Lefevre et al. 2010) and easterly tip jets (Moore and Renfrew 2005; Renfrew et al. 2009; Outten et al. 2009; Outten et al. 2010). Nigro et al. (2012) used the polar modified 15 km Weather Research and Forecasting (WRF) model run within AMPS and automatic weather station (AWS) observations to study a BWCJ located to the northwest of the Prince
Olav Mountains and more specific details on the formation of BWCJs in the region of the Prince Olav Mountains can be found in that paper.

This paper presents a wind climatology over the RIS using output from the polar modified 15 km WRF model run within AMPS. The previous low-level wind climatology presented by Seefeldt and Cassano (2012) used output from the polar modified 30 km MM5 model run within AMPS for the analysis, so the results presented here are the first RIS wind climatology presented at this higher resolution. The wind climatology presented below provides information on the types of wind patterns present over the RIS, the frequency of these patterns and the seasonality of these patterns.

Subsequently, a climatology of RAS patterns, the most dominant pattern in the low-level wind field over the RIS, is presented. The RAS climatology provides a better understanding of the variability in the strength and position of the RAS, as well as, information on the frequency and seasonality of the various RAS patterns. This paper further investigates the atmospheric dynamics associated with the barrier wind component of the RAS in the identified patterns. The dynamics of the other components of the RAS will be presented in a future paper.

2. Data
2.1 Model

Output from the 15 km WRF model run within AMPS is used to create the low-level wind climatology over the RIS. The polar modified version of WRF is currently run within AMPS and has been optimized for use in the polar regions (Hines and Bromwich 2008; Bromwich et al. 2009; Hines et al. 2011). The major changes made to the WRF model include implementation of a scheme to treat fractional sea ice, improved treatment of heat transfer through ice and snow surfaces, a revised surface energy balance calculation, and a selection of model physics options that are appropriate for polar applications.

AMPS is run with a set of six two-way nested domains. The grid spacing of the 6 domains is 45 km, 15 km, 5 km (3 domains), and 1.67 km and the vertical resolution includes 44 eta levels. AMPS is run twice daily by the National Center for Atmospheric Research (NCAR). The first guess initialization and boundary conditions are taken from The National Center for Environmental Prediction 0.5-degree Global Forecast System model output. Observations are assimilated into AMPS using WRF’s three-dimensional variational data assimilation. The twice-daily AMPS output is available through NCAR. For the RIS wind climatology, output from the AMPS 15 km domain two, which covers the Antarctic continent, was analyzed. AMPS domain two is run for a duration of 120 hours and the output is archived every three hours. The 12-21 hour forecasts are used to create the wind climatology. The 0-9 hour forecasts are not used as the model is still adjusting from the coarse resolution initial conditions during this time. The use of the 12-21 hour forecasts is consistent with Bromwich et al. (2005) and Seefeldt and Cassano (2008). For more
information on AMPS see Powers et al. (2012) and the AMPS website at http://www.mmm.ucar.edu/rt/wrf/amps/.

In addition to being used to create the wind climatology, the model data is also used to calculate the horizontal PGF over the RIS and the Rossby radius of deformation for a barrier wind along the Transantarctic Mountains.

The following analysis uses the PGF over the RIS to evaluate the presence of a barrier wind along the Transantarctic Mountains. During the formation of a barrier wind, the flow directed towards the barrier is blocked and mass convergence occurs. This increases the pressure adjacent to the mountains and creates a strong PGF located adjacent to the base of the mountains. Therefore, the presence of the strongest PGF located adjacent to the barrier is a common characteristic of a barrier wind. The PGF is calculated from the AMPS output using the following equations (Holton 2004).

\[
pg_x = -\frac{1}{\rho} \frac{\partial p}{\partial x} \\
pg_y = -\frac{1}{\rho} \frac{\partial p}{\partial y}
\]

(3)

where \( \rho \) is density of the atmosphere and \( \frac{\partial p}{\partial x} \) and \( \frac{\partial p}{\partial y} \) are the gradient of pressure in the grid relative zonal (x-direction) and meridional (y-direction) directions. The partial derivative is solved using a two-point centered finite difference. Additionally, prior to the calculation, each pressure value used in the calculation is interpolated to the elevation of
the grid point of interest using the hypsometric equation and the model surface
temperature, providing a more accurate calculation of the partial derivative.

The Rossby radius of deformation is used in this study to determine the
approximate width of a barrier wind along the Transantarctic Mountains. King and Turner
(1997) provide a good example of calculating the Rossby radius of deformation for a
barrier wind along the Antarctic Peninsula. The Rossby radius of deformation is the
horizontal length over which the atmosphere adjusts due to the blocked flow by the barrier.
For instance, a barrier wind forms when flow is directed towards a barrier. If the flow is
blocked, the pressure field is altered, providing the forcing for a barrier wind. The Rossby
radius of deformation indicates the distance away from the barrier in which the pressure
field is altered, providing an approximate width of the barrier wind. The Rossby radius of
deformation is calculated using the following equation from King and Turner (1997):

\[
R_R = \frac{1}{|f|} (gH \frac{\Delta \Theta}{\Theta})^{0.5}
\]  

(4)

where \( f \) is the Coriolis parameter, \( g \) is the acceleration due to gravity, \( H \) is the height of the
barrier, \( \Delta \theta \) is the stability of the layer of blocked flow and \( \theta \) is the average potential
temperature of the layer of blocked flow.

\[2.2\] Observations
Observations of wind speed and wind direction from University of Wisconsin AWS located over the RIS are compared to the AMPS output. This type of comparison provides information on how well the AMPS output represents the low-level wind field over the RIS. Each AWS records the temperature, pressure, wind speed and wind direction every 10 minutes. The data is then transmitted in real-time via satellite using the ARGOS system. The Antarctic Meteorological Research Center (AMRC) at the University of Wisconsin-Madison retrieves the AWS observations and uses a semi-automatic process to quality control the dataset. The observations are then processed into a 3-hourly time series by extracting the nearest observation to the 3-hourly time within a plus or minus 40-minute time limit (Keller et al. 2010, Lazzara et al. 2012). The resulting 3-hourly time series of AWS observations is at the same temporal resolution as the 15 km AMPS output and therefore can be used as a direct comparison between the two datasets.

The AMPS 10 m winds are adjusted to the height of each AWS using the logarithmic wind profile equation (Holton 2004). This calculation uses a roughness length of 0.0001 m, which is the same roughness length used within AMPS, and neglects the stability correction term since the calculation is conducted over a short distance.

3. Methods

The method of self-organizing maps (SOM) is used to create a wind climatology of the 10 m winds over the RIS. The SOM training method uses a neural network algorithm
and an unsupervised, iterative learning process to identify a user specified number of patterns within a dataset. See Kohonen (2001) for a thorough explanation of the SOM training method and algorithm. For this paper, the SOM was trained using a two-year dataset (October 2008 through September 2010) of the u and v-components of the 15 km AMPS-WRF 10 m winds over the RIS. The extent of the dataset was limited to the model grid points over the RIS (see the shaded grid points in Fig. 25). In contrast to the SOM analysis conducted by Seefeldt and Cassano (2012), this domain was chosen such that it excludes the strong winds that surround the RIS (i.e. katabatic winds in the Transantarctic Mountains and synoptic scale cyclonic winds in the Ross Sea). By excluding the surrounding winds, the SOM results provide a more detailed analysis of the mesoscale wind features over the RIS.

The SOM method requires the user to specify the number of patterns to be identified. See Reusch et al. (2005) for a thorough explanation on determining the number of patterns that should be used in a SOM analysis. For the wind climatology presented here, a variety of different grid sizes were run through the SOM (i.e. 4x3, 5x4, 6x4, 6x5, 7x5). The results were analyzed to determine how many patterns were necessary to capture each of the dominant wind patterns over the RIS. The results indicated a 6x4 grid size, or 24 patterns, was sufficient for identifying the dominant wind patterns over the RIS. In the smaller grid sizes, some of the wind patterns that are present in the 6x4 grid were missing. In the larger grid sizes some of the patterns in the 6x4 grid were duplicated, or the differences between the adjacent patterns were difficult to identify. A similar process was used to determine the
grid size for the RAS SOM climatology and a 5x4 grid was determined sufficient to capture each of the RAS patterns over the RIS.

Fig. 26a shows the 10 m wind patterns identified by the SOM. Each pattern is referred to as a node and is referenced by its column number (zero is the leftmost column) and row number (zero is the top row). For example, the pattern in the upper-left corner is referred to as node [0,0] and the pattern in the bottom-right corner is referred to as node [5,3] (each node on the SOM, Fig. 26a, is labeled in its upper, right-hand corner). The patterns are organized such that similar patterns are located adjacent to each other and non-similar patterns are located further away from each other.
Fig. 26. Node averaged 10 m winds over the Ross Ice Shelf.
Once the dominant 10 m wind patterns of the dataset have been identified by the SOM, each of the 10 m wind forecasts used to train the SOM is matched to the node it most closely resembles. This process is called “mapping”. The process uses a least squared differencing method to quantitatively determine the node that an individual forecast is
most similar to. Specifically, the squared differences between the 10 m winds in the forecast and the 10 m winds in the SOM node are calculated. The node that results in the smallest squared difference is the pattern the 10 m wind forecast maps to. Once each forecast is mapped to a node, a plot showing the average of the 10 m wind forecasts that map to each node can be made. This plot is called the node averaged 10 m wind plot (Fig. 26a). The SLP anomalies corresponding to the 10 m wind forecasts mapped to each node are averaged together to create the node averaged SLP anomalies plot (Fig. 26b). This process can be used to create node averaged plots for any available or calculated model variable.

4. Results

4.1 Wind Climatology

Previous studies have used the AMPS-MM5 30 km output to study the low-level winds over the RIS (Parish et al. 2006; Seefeldt et al. 2012), while this study uses the AMPS-WRF 15 km output to study the low-level winds over the RIS. Figure 27a shows the average lowest model level winds for the AMPS-MM5 30 km output (approximately 13 m above ground level (AGL) over the RIS) from January 2001 through December 2005 and Fig. 27b shows the average lowest model level winds for the AMPS-WRF 15 km output (approximately 12 m AGL over the RIS) from October 2008 through September 2010. The average winds from these datasets both show southerly winds, or northerly transport, over
the RIS. However, the average winds from these datasets differ in the magnitude of the wind speed and the location of the corridor of strongest winds flowing from the Siple Coast confluence zone, northward to the Ross Sea. The winds in the AMPS-MM5 30 km dataset are, in general, stronger than the winds in the AMPS-WRF 15 km dataset; and the corridor of strongest winds in the AMPS-MM5 30 km dataset are located adjacent to the Transantarctic Mountains while the corridor of strongest winds in the AMPS-WRF 15 km dataset are located through the center of the RIS.
Due to the differences in the average low-level winds in the two datasets (Figs. 27a and 27b), a statistical analysis of the AMPS-MM5 30 km (January 2001 through December
2005) and AMPS-WRF 15 km (October 2008 through September 2010) datasets was conducted. Each dataset was compared to wind speed and wind direction observations from 12 RIS AWS (shown in Fig. 25). The statistical analysis was conducted for the u and v components of the wind, the wind speed and the wind direction. The bias, root mean square error (RMSE) and correlation were calculated for each category, except for wind direction where only the bias was calculated. Table 2 shows the results of the statistical analysis. For the u and v components of the wind and the wind speed, the RMSE was lower and the correlation was higher for the AMPS-WRF 15 km dataset. Additionally, the wind direction bias decreased from -35° in the AMPS-MM5 30 km dataset to -6° in the AMPS-WRF 15 km dataset. These results indicate that the AMPS-WRF 15 km output better represents the observed low-level wind field over the RIS. Given the more realistic representations of the RIS winds in the AMPS-WRF 15 km dataset, the results presented here are likely an improvement over prior RAS studies conducted with the AMPS-MM5 30 km data.

<table>
<thead>
<tr>
<th>Wind</th>
<th>Stats</th>
<th>AMPS-MM5 30 km</th>
<th>AMPS-WRF 15 km</th>
</tr>
</thead>
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<tr>
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<td>Correlation</td>
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<tr>
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<td>RMSE</td>
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<td></td>
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<tr>
<td>Wind Direction</td>
<td>Bias</td>
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<td>-5.7 deg</td>
</tr>
</tbody>
</table>

Table 2: Results of the statistical analysis between 12 Ross Ice Shelf AWS and the AMPS-MM5 30 km and AMPS-WRF 15 km datasets.
The SOM node averaged 10 m winds, in Fig. 26a, shows the typical low-level wind patterns that occur over the RIS region. The frequency and seasonality of each wind pattern in Fig. 26a are shown in Figs. 28a and 28b, respectively. The frequency plot indicates the percentage of time each wind pattern occurs during the two-year time span. The shading in this plot is scaled by the frequency, with darker shading indicating a greater frequency of occurrence. The seasonality plot indicates the percentage of time each pattern occurs during the summer (December and January), fall (February, March and April), winter (May, June, July and August), and spring (September, October and November). These seasonal definitions better align with the temperature changes over Antarctica than the typical four, three-month seasons and have been used in previous research (Seefeldt and Cassano 2008, Seefeldt and Cassano 2012). The wind climatology patterns are subjectively grouped into the following categories, “RAS” (outlined in red in Fig. 26), “northward transport over the eastern RIS” (outlined in orange in Fig. 26), “katabatic winds” (outlined in blue in Fig. 26), “mesocyclones over the RIS” (outlined in purple in Fig. 26), and “weak winds” (outlined in green in Fig. 26). The dominant wind feature present within each pattern drives the identification of these categories. Other wind features may also be present within a pattern, such as katabatic winds within “RAS” patterns, although the pattern is categorized with respect to its dominant wind feature. Each of these categories will be described below, as well as, the frequency (Fig. 28a) and seasonality (Fig. 28b) of each category.
Figure 29. (a) Node frequency and (b) seasonal frequency for each wind pattern in Fig. 26a. The seasonality plot indicates the percentage of time each pattern occurs during the summer (DJ), fall (FMA), winter (MJA), and spring (SON).
The “RAS” patterns (nodes [0,0], [0,1], [0,2], [0,3], [1,0], [1,1], [1,2], [2,0], and [3,0]) show the RAS, or northward transport of air over the western to central RIS, as defined by Seefeldt and Cassano (2012). These patterns have strong winds over the RIS that transport air from the southern tip of the RIS towards the north, over the western to central RIS, to the Ross Sea. Within these patterns, the location of the strongest winds over the RIS varies. Nodes [2,0] and [3,0] show the fastest winds located through the center of the RIS (Fig. 26a). These patterns have a synoptic cyclone located off the coast of West Antarctica and a strong pressure gradient (PG) over the RIS with SLP contours that are oriented meridionally (Fig. 26b). Node [0,0] shows the strongest winds dominating the RIS. This pattern has a synoptic cyclone located over the eastern Ross Sea and a strong PG over the RIS. Nodes [0,2] and [0,3] show the strongest winds over the RIS located adjacent to the Transantarctic Mountains. These patterns have a weak synoptic cyclone in the Ross Sea, with a trough of low pressure extending from the primary low and penetrating the RIS. The remaining nodes, [0,1], [1,0], [1,1] and [1,2]), have a combination of the features discussed in this paragraph. The “RAS” patterns have a combined frequency of 34% (Fig. 28a). The patterns are predominantly non-summer patterns, although RAS patterns can occur during the summer (Fig. 28b).

The “northward transport over the eastern RIS” patterns (nodes [4,0] and [5,0]) have strong winds over the northern and eastern portions of the RIS (Fig. 26a). The node averaged SLP anomalies for these patterns have a synoptic cyclone located off the coast of West Antarctica (Fig. 26b). The combined frequency of the “northward transport over the
The “katabatic wind” patterns (nodes [2,1], [3,1], [4,1] and [4,2]) show katabatic drainage onto the RIS. These patterns have weaker winds over the RIS, with areas of stronger winds at the base of Byrd Glacier and the Siple Coast confluence zone (Fig. 26a). The “katabatic wind” patterns can be broken into “strong katabatic” patterns (nodes [2,1] and [3,1]) and “weak katabatic” patterns (nodes [4,1] and [4,2]). For these patterns the synoptic cyclone off the coast of West Antarctica is shifted to the east, almost out of the SOM domain, resulting in a weaker PG over the RIS. The SLP contours over the RIS are generally oriented meridionally (Fig. 26b). “Strong katabatic” winds occur 8% of the time and are non-summer patterns with a slight maximum in the frequency during the winter months. “Weak katabatic” winds occur 8% of the time and occur in all seasons (Fig. 28).

The “mesocyclones over the RIS” patterns (nodes [1,3], [2,2], [2,3], [3,2], and [3,3]) have a mesocyclone over the RIS, which can be seen in the node averaged SLP anomalies (Fig. 26b). Additionally, cyclonic flow over the RIS is shown in the node averaged 10 m wind for these patterns (Fig. 26a). Mesocyclone patterns occur over the RIS approximately 20% of the time (Fig. 28a). Mesocyclone patterns occur throughout the year, with a slight maximum in the frequency during winter for nodes [2,2], [3,2] and [3,3] (Fig. 28b).

The “weak winds” patterns (nodes [4,3], [5,1], [5,2], and [5,3]) show light, variable winds over the RIS (Fig. 26a). The SLP anomalies for these patterns show high pressure
and a weak PG over the RIS. Additionally, there is little to no influence from a synoptic
cyclone in the Ross Sea (Fig. 26b). Weak winds occur approximately 21% of the time over
the RIS (Fig. 28a). These patterns are predominantly summer patterns (Fig. 28b).

4.2 RAS Climatology

The SOM can be trained using a subset of the original dataset, providing a more
detailed analysis of the patterns chosen for the subset. In order to better understand the
RAS patterns in Fig. 26a, the SOM was retrained using only the weather patterns identified
as RAS patterns (nodes [0,0], [0,1], [0,2], [0,3], [1,0], [1,1], [1,2], [2,0] and [3,0]). The results
for the subset SOM are shown in Figs. 29 and 30, with the node averaged 10 m winds in Fig.
29a, the node averaged SLP anomalies in Fig. 29b, the annual frequency in Fig. 30a and the
seasonal frequencies in Fig. 30b. The node averaged 10 m winds show patterns ranging
from the strongest winds dominating the majority of the RIS (column zero) to the strongest
winds located adjacent to the Transantarctic Mountains (column four), with the patterns in
the center of the SOM showing a progression between the patterns on the left side of the
SOM to the patterns on the right side of the SOM. In general, the RAS wind patterns show
variability in the strength and position of the RAS.
Fig. 30. Node averaged 10 m winds over the Ross Ice Shelf for the subset SOM of RAS patterns. The triangle indicates the location of the Prince Olav Mountains.
Fig. 31. Node averaged SLP anomalies over the Ross Ice Shelf for the subset SOM of RAS patterns. The triangle indicates the location of the Prince Olav Mountains.
Fig. 32. (a) Node frequency and (b) seasonal frequency for each RAS wind pattern in Figure 29a. The seasonality plot indicates the percentage of time each pattern occurs during the summer (DJ), fall (FMA), winter (MJJA), and spring (SON).
4.3 Barrier Winds

As mentioned in the introduction, this paper will focus on the atmospheric dynamics associated with the barrier wind component of the RAS. The results presented in this section are preliminary results from this research. This section starts with a discussion of the identification of the RAS patterns with barrier winds and BWCJs, followed by a discussion on what causes the forcing for the barrier wind component of the RAS in these patterns.

4.3.1 Barrier Wind Identification

It is difficult to diagnose a barrier wind from a single point in time. Barrier winds form when stable flow is directed towards a barrier. Therefore, a time evolution of model output or observations is ideally used to show evidence of flow directed towards a barrier and then the development of a barrier wind. Although, the likely barrier wind patterns in Figure 29a can be identified using a set of characteristics that are commonly present in barrier winds. These common characteristics include: high pressure adjacent to the barrier (Parish et al. 2006), a strong PGF adjacent to the barrier (Parish et al. 2006), strong winds parallel and adjacent to the barrier (Parish et al. 2006), a barrier parallel thermal wind (as is described below), and strong winds within a Rossby radius of deformation of the barrier (King and Turner 1997).
Barrier winds commonly have high pressure located adjacent to the barrier. This occurs because barrier winds form when stably stratified flow is blocked by a barrier (i.e. the Transantarctic Mountains), causing mass convergence and an increase in pressure at the base of the barrier. Figure 29b of the node averaged SLP anomalies for the RAS patterns, indicates each of the identified RAS patterns has high pressure located adjacent to the mountains.

During the formation of a barrier wind, the region of high pressure located adjacent to the mountains creates a PGF directed perpendicular and away from the mountains. The magnitude of the PGF tends to be strongest adjacent to the mountains and decreases with distance away from the mountains (as the influence from the blocked flow decreases). Figure 31a shows the node averaged PGF for each pattern. The patterns in columns 0, 1 and 2 (with the exception of nodes [0,3] and [1,3]) have the strongest PGF through the center of the RIS with an area of weaker PGF adjacent to the mountains (Fig. 31a), indicating these patterns are non-barrier wind patterns. Nodes [0,3] and [1,3] show a strong PGF in the northwestern RIS which could indicate blocking of the flow around the complex topography of the Ross Island region (Seefeldt et al. 2003), but is not indicative of a typical barrier wind along the Transantarctic Mountains. On the other hand, the strongest PGF in columns 3 and 4 are located adjacent to some portion of the Transantarctic Mountains. The strongest PGF in nodes [4,2], [4,3] and [3,3] is found adjacent to the Transantarctic Mountains extending from the southern tip of the RIS to the region near Byrd Glacier and
the strongest PGF in nodes [3,0], [3,1], [3,2], [4,0] and [4,1] is adjacent to the Transantarctic Mountains in the southern tip of the RIS.

Fig. 33. Node averaged pressure gradient force over the Ross Ice Shelf for the subset SOM of RAS patterns. The triangle indicates the location of the Prince Olav Mountains.
The high pressure and strong PGF located adjacent to the mountains induces a wind, which becomes approximately geostrophic and flows parallel to the barrier. This barrier parallel flow is the barrier wind. Similar to the PGF, the magnitude of the barrier wind tends to be strongest adjacent to the mountains and decreases with distance away from the mountains. Figure 29a shows the node averaged 10 m winds for the RAS patterns. The nodes in columns 0, 1 and 2 were shown to be non-barrier wind patterns using the PGF plots and therefore will not be discussed in this section. The nodes in columns 3 and 4 of
Fig. 29a each have regions where the strongest winds flow parallel to the barrier and are located adjacent to the Transantarctic Mountains. In nodes [4,2], [4,3] and [3,3] these winds extend from the southern RIS to the region near Byrd Glacier. In nodes [3,0], [3,1], [3,2], [4,0], and [4,1] the winds are adjacent and parallel to the Transantarctic Mountains in the southern tip of the RIS and pull away from the Transantarctic Mountains in the northwestern RIS, indicating barrier winds confined to the southern RIS.

The thermal wind can also be used to analyze the presence of a barrier wind in the RAS patterns. The thermal wind indicates the change, in magnitude and direction, of the geostrophic winds between two layers of the atmosphere (i.e. the thermal wind between 900 mb and 700 mb is the difference between the geostrophic wind at 900 mb and the geostrophic wind at 700 mb). This is useful for diagnosing a barrier wind because a barrier wind is a low-level geostrophic feature, which is confined to the region below the height of the mountains. Therefore, a barrier wind, if present, will be evident in a thermal wind analysis between a layer near the top of the mountains that does not contain a barrier wind and a layer near the base of the mountains that contains a barrier wind. Figure 31b shows the thermal wind analysis between 900 mb and 700 mb for each node.

Nodes [4,2], [4,3], and [3,3] show thermal winds that are barrier parallel and adjacent to the Transantarctic Mountains extending from the southern RIS to the region near Byrd Glacier (Fig. 31b). This is consistent with the 10 m winds and PGF for these nodes, which both indicate a barrier wind extending from the southern RIS to the region near Byrd Glacier. Conversely, for node [3,2] the thermal winds are parallel and adjacent to
the Transantarctic Mountains from the southern RIS to Byrd Glacier, but the 10 m winds and PGF for this node indicate that a barrier wind is confined to the southern RIS. Therefore, the barrier wind in node [3,2] is confined to the southern RIS. Lastly, nodes [3,0], [3,1], [4,0], and [4,1] show a thermal wind (Fig. 31b) that is barrier parallel and adjacent to the mountains only in the southern RIS. This is consistent with the 10 m winds and PGF for these nodes, which indicate a barrier wind confined to the southern RIS. In summary, the evaluation of the barrier wind characteristics shows that nodes [4,2], [4,3], and [3,3] are consistent with barrier winds that extend from the southern RIS to the region near Byrd Glacier and the remaining nodes in columns 3 and 4 are consistent with barrier winds that are confined to the southern RIS.

An additional barrier wind characteristic is the Rossby radius of deformation. As previously described, a barrier wind will have a width that is less than or equal to the Rossby radius of deformation (King and Turner 1997). The Rossby radius of deformation is approximately 340 km in the region of the southern RIS (near the Prince Olav Mountains) and approximately 180 km in the region of the northwestern RIS (near Byrd Glacier), indicating the Rossby radius of deformation changes with the varying heights of the Transantarctic Mountains. These values were calculated using the average stability and potential temperature for various locations within each of the RAS patterns identified in Fig. 29. Using the length scale provided in Fig. 25 (located in the bottom, left) as a reference, the width of the previously identified barrier winds in columns 3 and 4 of Fig. 29a are less than 340 km in the southern RIS and less than 180 km in the region of Byrd Glacier and are therefore within the Rossby radius of deformation for this region. This provides additional
confirmation for the presence of barrier winds in these nodes. Conversely, the strong 10 m winds in columns 0 and 1 of Fig. 29a are wider than the Rossby radius of deformation for these regions, further indicating these nodes are non-barrier wind patterns.

In addition to identifying barrier winds, the SOM analysis can be used to identify BWCJs. Nigro et al. (2012) showed that BWCJs form off the coast of the Prince Olav Mountains when a southern RIS barrier wind reaches the end of the barrier, or the area of maximum protrusion by the Prince Olav Mountains. At this point, the flow transitions from an area of terrain induced high pressure to the background pressure field of the region and the PGF aligns with the direction of the flow. The alignment of the PGF with the direction of flow increases the magnitude of the wind speed, creating the BWCJ. In Nigro et al. (2012) this was depicted using the PGF plot in their Fig. 13c. The PGF plot shows a strong PGF in the southern tip of the RIS (the region of the southern RIS barrier wind) and a weaker PGF to the northwest of the Prince Olav Mountains (the region of the BWCJ). The nodes in columns 3 and 4 of Fig. 31a show a similar pattern in the PGF. These nodes have a strong PGF in the southern tip of the RIS and a weaker PGF to the northwest of the Prince Olav Mountains. Additionally, the 10 m wind plots (Fig. 29a) for the nodes in columns 3 and 4 show the maximum winds are located in the region of weaker PGF, or directly to the northwest of the Prince Olav Mountains. Therefore, the patterns in columns 3 and 4 of the subset SOM (Fig. 29a) are consistent with the characteristics for BWCJs to the northwest of the Prince Olav Mountains.
Using common characteristics of barrier winds and BWCJs, the nodes in columns 3 and 4 of the subset SOM (Fig. 29) have been identified as patterns containing both barrier winds and BWCJs. The barrier wind and BWCJ patterns occur approximately 14% of the time (Fig. 30a) and approximately 41% of the time when a RAS pattern occurs. The barrier wind and BWCJ patterns are mostly non-summer patterns (Fig. 30b), although barrier wind patterns can occur during the summer months.

4.3.2 Barrier Wind Forcing

As previously mentioned, barrier winds are best identified using data that shows the evolution of the flow from when the wind is directed towards the barrier to the eventual barrier parallel flow. Therefore, this section will begin by discussing the typical evolution of the winds through the SOM patterns, followed by a discussion of winds directed towards a barrier.

Figure 32 shows the node transitions, or how weather patterns evolve, through the RAS patterns shown in Fig. 29a. The arrows indicate transitions from one pattern to the next, with the size of the arrows indicating the frequency at which the transition occurs, with larger arrows indicating transitions that occur more often. Due to limitations of using a two-dimensional plot, the transition arrows only show the transitions between neighboring, or adjacent, nodes. The labels in the top left of each node indicate the frequency of transitions to a neighboring node (what is shown by the arrows), to a non-
neighboring node (any non-adjacent node on the subset SOM), to itself (this is the most frequent transition for each node) and to a pattern on the full SOM (Fig. 26a). This information helps to give a complete picture of the transitions from a particular node.

Fig. 35. Transition analysis of the subset SOM RAS patterns. The arrows indicate a transition from one pattern to another. The larger an arrow, the more frequent that transition occurs.

The most dominant transitions move from left to right across the top row of Fig. 32 (node [0,0] to [4,0]) and then down column four (node [4,0] to [4,3]). Row two also shows general transitions from left to right. Moving through these dominant transitions provides
information on how the wind patterns evolve into barrier winds (nodes in columns three and four).

In addition to the transition analysis presented above, a manual analysis of the transitions from the full SOM (Fig. 26) to the subset SOM (Fig. 29) was conducted. This analysis indicated that the mesocyclone patterns transition from node [3,3] to node [2,3] to node [1,3] on the full SOM (Fig. 26) and then to the barrier wind patterns in nodes [4,1], [4,2] and [4,3] on the subset SOM (Fig. 29), suggesting that a mesocyclone moving from the southeastern RIS towards the west and north is another typical evolution leading up to the formation of a barrier wind in the region of the RIS.

The typical evolution of the wind patterns through the SOM is useful in diagnosing flow directed towards a barrier prior to the development of a barrier wind. As discussed, the patterns of the RAS climatology tend to transition from left to right across the top row of the subset SOM (from node [0,0] to node [4,0]). The 10 m winds, SLP anomalies, 700 mb geopotential heights with the 700 mb wind vector overlaid, and surface PGF plots for these nodes are shown in Fig. 33 and are arranged such that the dominant transitions are down the columns. Moving down the columns, the 10 m winds show the strongest winds dominating the majority of the RIS, then becoming narrower and moving towards the center of the RIS, and then shifting westward, becoming adjacent to the Transantarctic Mountains in the southern RIS. The SLP anomalies and the 700 mb geopotential heights (a level above the West Antarctic Ice Sheet and below the tops of the Prince Olav Mountains) show the trough from the synoptic cyclone located off the coast of West Antarctica shifting
westward over the RIS. The movement of this trough towards the west alters the direction of the 700 mb winds located to the east of the southern tip of the RIS. The 700 mb winds to the east of the southern tip of the RIS change from southeasterly (parallel to the Transantarctic Mountains) in node [0,0] to easterly (with a component directed towards the Prince Olav Mountains) in nodes [3,0] and [4,0]. As the flow directed towards the Prince Olav Mountains in nodes [3,0] and [4,0] is blocked, mass convergence occurs and increases the pressure (the SLP anomalies in nodes [0,0] through [4,0] each show high pressure adjacent to the Transantarctic Mountains in the southern RIS) and the PGF adjacent to the mountains in the southern RIS. The increase in the PGF is shown moving down the column of PGF plots in Fig. 33, where an area of relatively weak PGF is located adjacent to the mountains in the southern RIS in nodes [0,0], [1,0] and [2,0] and the strongest PGF is located adjacent to the mountains in nodes [3,0] and [4,0]. It is the strongest PGF located adjacent to the mountains in nodes [3,0] and [4,0] that creates barrier winds in the southern RIS of these nodes. This analysis is consistent with the September 2009 case study presented by Nigro et al. (2012). That case study concluded that a barrier wind formed in the southern RIS after a synoptic cyclone traversed the Ross Sea and became stationary off the coast of West Antarctica. As the cyclone was stationary off the coast of West Antarctica, upper level winds were directed over the West Antarctic Ice Sheet towards the Prince Olav Mountains. The upper level winds directed towards the Prince Olav Mountains were blocked, causing mass convergence and an increase in the pressure and PGF in the southern RIS, providing the setup for a barrier wind to the southeast of the Prince Olav Mountains.
The manual transition analysis also indicated that a mesocyclone moving from the southeastern RIS towards the west and north (nodes [3,3], [2,3] and [1,3] of the full SOM, Fig. 26) is also a precursor to the barrier winds in nodes [4,1], [4,2] and [4,3] of the subset SOM (Fig. 29). Figure 34 shows the node averaged 10 m winds, SLP anomalies, 700 mb geopotential heights, 700 mb winds, and surface PGF for nodes [3,3], [2,3] and [1,3] of the full SOM (Fig. 26) and are arranged such that the dominant transitions move down the columns. Moving down the columns of node averaged 10 m winds and SLP anomalies, the mesocyclone moves from the southeastern RIS towards the west and the north. The 700 mb geopotential heights also show a trough moving from the southeastern RIS towards the west and north, altering the 700 mb winds. The 700 mb winds in node [3,3] show easterly and northeasterly flow directed towards the Prince Olav Mountains. As the low pressure moves towards the west and north, the amount of flow directed towards the Prince Olav Mountains increases. As the flow directed towards the mountains increases and is blocked, enough mass convergence occurs to increase the pressure and PGF adjacent to the mountains in the southern RIS. This is shown transitioning from node [2,3] to node [1,3] where a weak PGF is shown in the southern RIS in node [2,3] of Fig. 34, and a slightly increased PGF located adjacent to the mountains is shown in node [1,3]. This increase in the PGF adjacent to the mountains provides the forcing for the southern RIS barrier wind shown in nodes [4,1], [4,2] and [4,3] of the subset SOM. Therefore, a mesocyclone moving westward and northward over the RIS can direct flow towards the Prince Olav Mountains, inducing a barrier wind in the southern RIS.
Fig. 37. The node averaged 10 m winds, SLP anomalies, 700 mb geopotential heights, 700 mb winds and PGF for nodes [3,3], [2,3], and [1,3] of the full SOM in Fig. 26.
Nodes [4,2], [4,3], and [3,3] of the subset SOM (Fig. 29) show barrier winds extending from the southern RIS to the region near Byrd Glacier. These patterns evolve following the dominant node transitions down column 4 of the subset SOM (Fig. 32). Figure 35 shows the node averaged 900 mb to 700 mb thermal winds, SLP anomalies, 850 mb geopotential heights with 850 mb wind vectors overlaid, and surface PGF plots from column 4 of the subset SOM (node [4,0] to node [4,3]). The 850 mb winds are used in this analysis because the Transantarctic Mountains to the northwest of the Prince Olav Mountains are not as tall as the Prince Olav Mountains. Therefore, 700 mb winds are located above these mountains and cannot be blocked by the mountains.
The barrier winds in column 4 start as a barrier wind confined to the southern RIS (node [4,0] and node [4,1]) and transition into a barrier wind extending from the southern RIS to Byrd Glacier (node [4,3]) (Fig. 35). This evolution can be seen when following the
dominant transitions down column 4 of the thermal wind (the leftmost column in Fig. 35). The thermal winds in nodes [4,0] and [4,1] are barrier parallel and adjacent to the Transantarctic Mountains in the southern RIS and pull away from the Transantarctic Mountains in the regions to the northwest of the Prince Olav Mountains. In node [4,2] the thermal wind is barrier parallel and adjacent to the Transantarctic Mountains extending from the southern RIS to the region of the Queen Alexander Mountains (see Fig. 25 for location). In node [4,3] the thermal wind is barrier parallel extending from the southern RIS to Byrd Glacier.

Moving down the column of SLP anomalies and the 850 mb geopotential heights in Fig. 35, it is shown that the elongated trough from the synoptic cyclone off the coast of West Antarctica tends to move westward and northward over the RIS, altering the 850 mb winds. Specifically, the 850 mb winds show north, northeasterly flow over the eastern RIS in node [4,1], northeasterly flow over the eastern and central RIS in node [4,2], and northeasterly flow over the majority of the RIS in node [4,3]. The northeasterly winds in nodes [4,2] and [4,3] direct flow towards some regions of the Transantarctic Mountains. In node [4,2] the northeasterly flow is directed towards the region of the Queen Alexander Mountains (Fig. 25) and in node [4,3] the northeasterly flow is directed towards the regions extending from the Queen Alexander Mountains to Byrd Glacier. The flow directed towards these mountains is blocked, causing mass convergence and an increase in pressure adjacent to the Transantarctic Mountains (nodes [4,2] and [4,3] show high pressure adjacent to the mountains in these regions). The increase in pressure creates a strong PGF located adjacent to the mountains. This is shown when transitioning from node [4,1],
where the strongest PGF is located away from the Transantarctic Mountains, to nodes [4,2] and [4,3], where the strongest PGF is located adjacent to the Transantarctic Mountains and is consistent with a barrier wind in this region. Therefore, the movement of the trough (seen at 850 mb and in the SLP anomalies) westward and northward over the RIS directs flow towards the Transantarctic Mountains, inducing a barrier wind extending from the southern RIS to the region near Byrd Glacier.

5. Conclusion

This paper presents a 10 m wind climatology over the RIS. In the wind climatology, patterns are grouped into the following categories, “RAS”, “northward transport over the eastern RIS”, “katabatic winds”, “mesocyclones over the RIS”, and “weak winds” patterns over the RIS. The patterns within the “RAS” category occur approximately 34% of the time and are more frequent than the patterns in any of the other categories. To investigate this dominant wind regime over the RIS, a RAS climatology over the RIS was developed and presented. The RAS climatology indicated the RAS varies in strength and position over the RIS. The patterns within the RAS climatology range from dominant winds over the majority of the RIS, to a narrow corridor of strong winds through the center of the RIS, to the strongest winds located adjacent to the Transantarctic Mountains.

The barrier wind and BWCJ patterns within the RAS climatology were studied in more detail. The barrier wind and BWCJ patterns were identified using a barrier wind
characteristics analysis and shown to occur approximately 14% of the time and
approximately 41% of the time when a RAS pattern occurs.

The forcing associated with the barrier wind patterns was investigated. The barrier winds in the southern RIS form when an elongated trough (seen at 700 mb and in the SLP anomalies) from the semi-permanent synoptic cyclone located off the coast of West Antarctica protrudes onto the RIS and directs easterly flow from the West Antarctic Ice Sheet towards the Prince Olav Mountains. A similar setup occurs when a mesocyclone moves from the southeastern RIS towards the west and north. This also directs easterly flow from the West Antarctic Ice Sheet towards the Prince Olav Mountains. The flow directed towards the mountains is blocked, causing mass convergence and the setup for a barrier wind in the southern RIS. Barrier winds that extend from the southern RIS to the region near Byrd Glacier were also investigated. These barrier winds form when the trough (seen at 850 mb and in the SLP anomalies) moves westward and northwards over the RIS, directing flow towards regions of the Transantarctic Mountains. The flow directed towards the mountains is blocked, causing mass convergence and the setup for a barrier wind extending from the southern RIS to the region near Byrd Glacier.

Future work by the authors will use this SOM analysis to analyze the katabatic wind and synoptic forcing components of the RAS. The analysis will investigate the forcing associated with the strong winds that dominate the majority of the RIS (column 0 of Fig. 29a) and the corridor of strong winds through the center of the RIS (columns 1 and 2 of Fig.
29a). This future analysis combined with the results presented here will provide a fairly comprehensive description of the dynamics that drive the RAS over the RIS.

6. Acknowledgements

NSF Grants ANT-0943952 and ANT-0636811 supported this work. The AMPS data was retrieved from the National Center for Atmospheric Research-Computational and Information Systems Laboratory. The authors appreciate the support of the Automatic Weather Station Program and the Antarctic Meteorological Research Center for providing the quality controlled AWS data set (Matthew Lazzara, Linda Keller, Jonathan Thom and George Weidner, NSF grant number ANT-0944018).
Chapter 5: Conclusion

This dissertation analyzed the dynamics associated with the low-level wind field over the RIS. Although, prior to analyzing the low-level wind field over the RIS, a new technique for evaluating NWP models as a function of weather pattern was investigated. The validation of NWP models is an important part of Antarctic atmospheric science. Due to limited observations in the Antarctic, the output from NWP models is often validated using the available observations before using the output for research purposes. The new technique uses the method of self-organizing maps to identify model errors as a function of weather pattern. This dissertation presented an example of this technique using the method of self-organizing maps and AMPS output to identify the dominant weather patterns over the RIS. The AMPS forecasts were then compared to AWS observations for each of the identified weather patterns to determine if the performance of the model varied by weather pattern. The results indicated the model performance did vary with weather pattern, as well as, with location over the RIS, demonstrating the ability of this new weather pattern based model evaluation technique to identify model errors as a function of weather pattern.

The low-level wind field over the RIS was first analyzed using a September 2009 case study of a high wind event off the coast of the Prince Olav Mountains. The winds to the northwest of the Prince Olav Mountains have some of the strongest mean wind speeds over the RIS. These strong winds had been studied in the past (Seefeldt and Cassano (2008);
Steinhoff et al. (2009)), although these studies were unable to determine the dynamics that caused the strong winds to develop in this region.

The AWS observations were used to validate the performance of the AMPS-WRF 15 km forecasts in the region of the Prince Olav Mountains during the timeframe of the case study and indicated the forecasts represented the high wind event with reasonable accuracy.

The AMPS-WRF 15 km forecasts were used to diagnose the dynamics associated with the high wind event. It was concluded that the forcing associated with a BWCJ created the strong winds to the northwest of the Prince Olav Mountains. The BWCJ was formed when a synoptic cyclone traversed the Ross Sea and became stationary off the coast of West Antarctica. The cyclone off the coast of West Antarctica directed 700 mb flow over the West Antarctic Ice Sheet towards the Prince Olav Mountains. The protruding Prince Olav Mountains blocked this flow, causing mass convergence and a build up of pressure to the southeast of the Prince Olav Mountains. This region of high pressure created a PGF directed perpendicular and away from the Prince Olav Mountains, which induced a barrier wind in this region. This barrier wind flowed around the Prince Olav Mountains towards the northwest. When the barrier wind reached the area of maximum protrusion by the Prince Olav Mountains, the flow transitioned from a region of terrain induced high pressure to the background pressure field of the region and the PGF aligned with the direction of flow. The alignment of the PGF with the direction of flow increased the magnitude of the wind speed in this region, creating the BWCJ and some of the strongest winds over the RIS.
Subsequently, the unbalanced CF in the BWCJ turned the flow to the left, directing it towards the protruding Queen Alexander Mountains. The Queen Alexander Mountains blocked the flow, causing mass convergence and the setup for a second barrier wind. As this barrier wind flowed around the Queen Alexander Mountains, a second BWCJ formed to the northwest of the Queen Alexander Mountains. A similar process created a third barrier wind and BWCJ in the region of the Churchill Mountains.

In addition to determining the dynamics that cause the strong winds to the northwest of the Prince Olav Mountains, this study has shown that the acceleration of a barrier wind when it reaches the end of a barrier and the PGF aligns with the direction of flow is not confined to the tip of Greenland. Instead, a protruding mountain range can be sufficient for creating a barrier wind and a subsequent BWCJ. Due to the fact that this type of topographic setup occurs in many regions of the world, the conclusions from this case study are applicable to regions other than the Prince Olav Mountains.

The low-level winds over the RIS were further investigated by developing a low-level wind climatology over the RIS. The method of self-organizing maps and the AMPS-WRF 15 km dataset, an improvement over the previous dataset used to create a low-level wind climatology over the RIS (Seefeldt and Cassano 2012), were used to create the wind climatology. The typical wind patterns over the RIS were grouped into the following categories, “RAS”, “northward transport over the eastern RIS”, “katabatic winds”, “mesocyclones over the RIS”, and “weak winds.” The analysis showed the patterns within the “RAS” category occur approximately 34% of the time and are more frequent than the
patterns in any other category. The RAS patterns transport mass from the Siple Coast confluence zone, northward over the RIS to the Ross Sea and are an important part of the Southern Hemisphere climate system. Therefore, the RAS patterns were further investigated using a RAS climatology. The climatology showed that the RAS varies in its strength and position over the RIS. Additionally, some of the RAS patterns contained barrier winds while others did not, validating the hypothesis that the variability in the RAS patterns is driven by differing atmospheric dynamics.

The dynamics of the barrier wind component of the RAS patterns were further investigated. It was shown that barrier winds form in the southern RIS when an elongated trough (seen at 700 mb and in the SLP anomalies) from a synoptic cyclone located off the coast of West Antarctica moves westward and penetrates the RIS, directing easterly flow from the West Antarctic Ice Sheet towards the Prince Olav Mountains. A similar scenario occurs when a mesocyclone in the southeastern RIS moves westward and northward over the RIS. This movement also directs easterly flow from the West Antarctic Ice Sheet towards the Prince Olav Mountains. The flow directed towards the Prince Olav Mountains is blocked causing mass convergence and an increase in pressure and PGF in the southern RIS. The increased PGF in the southern RIS induces a barrier wind in this region, which flows towards the northwest.

The RAS climatology also showed patterns with barrier winds extending from the southern RIS to the region near Byrd Glacier. The forcing for these barrier winds was investigated and it was determined that the barrier winds occur when the elongated trough
(seen at 850 mb and in the SLP anomalies) from the synoptic cyclone off the coast of West Antarctica moves westward and northward over the RIS. This directs 850 mb northeasterly flow towards regions of the Transantarctic Mountains. As the mountains block this flow, mass convergence occurs creating the setup for a barrier wind that extends from the southern RIS to the region near Byrd Glacier.

In general, this dissertation showed the complexity of the low-level wind field over the RIS. The forcing for these winds is a complex interaction of synoptic, mesoscale, and topographic forcing. Therefore, future work will use the SOM analysis presented in Chapter 4 to understand the dynamics that drive the katabatic wind and synoptic forcing components of the RAS. Specifically, the dynamics that drive the RAS patterns with winds dominating the majority of the RIS (column 0 of Fig. 29a) and the dynamics that drive the RAS patterns with a narrow corridor of winds through the center of the RIS (columns 1 and 2 of Fig. 29a) will be investigated. This information combined with the results presented in Chapter 4 of this dissertation will provide a fairly comprehensive description of the dynamics that drive the RAS over the RIS.

In addition to further investigating the dynamics that drive the RAS, the dynamics associated with the BWCJs located to the northwest of the Prince Olav Mountains, Queen Alexander Mountains and Churchill Mountains will be further investigated. These mountain ranges protrude onto the RIS and therefore block the flow directed towards the mountains in this region. Therefore, an obvious question is, “what would the RAS look like if these protruding mountains ranges were smoothed?” This question will be answered using a
WRF simulation of a BWCJ event where the topography of the Prince Olav Mountains, the Queen Alexander Mountains and the Churchill Mountains is smoothed such that the protrusions of these mountains are removed. The results will be analyzed to understand how the removal of the protrusion alters the flow along the base of the Transantarctic Mountains. If the results of the simulation indicate that the protrusion of the mountains is necessary for the formation of the BWCJs, additional simulations will be run with different combinations of the mountain ranges smoothed. For example, the Prince Olav Mountains will be smoothed while the Queen Alexander and Churchill Mountains will remain in their original configuration. This will help to determine if the protrusion of the Queen Alexander Mountains and the Churchill Mountains provides sufficient forcing to develop a BWCJ or if the forcing from the Prince Olav Mountains (the largest protruding mountain range) is necessary for the development of these subsequent BWCJs.

Lastly, the vertical structure of the RAS will be investigated. In 2011, the Alexander AWS was installed on the RIS (-79.044 S, 170.651 E) and is located within the general path of the RAS. This AWS is 30 m tall and was outfitted with 6 levels of instruments, including either an anemometer or an aerovane at each level. The height of this AWS is not tall enough to fully diagnose the vertical structure of the RAS, but the AWS observations will be combined with observations from unmanned aerial vehicles and AMPS output to better understand the vertical structure of the RAS in this region.
References


