Case Study of a Barrier Wind Corner Jet off the Coast of the Prince Olav Mountains, Antarctica

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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. 1). 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. 1). The following case study focuses on the formation of the wind speed maximum off the coast of the Prince Olav Mountains.

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. 1, 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. 2). 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

![Fig. 2. A schematic depiction of the Greenland easterly tip jet based on Moore and Renfrew (2005).](image-url)
In fact, at the end of the barrier, the PGF aligns with the direction of flow (Fig. 2). 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

a. 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 Arctic. 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. 3). 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 1. A list of the physical parameterizations in AMPS and the upper boundary condition.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Scheme</th>
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<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>
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<tr>
<td>Boundary layer</td>
<td>Mellor–Yamada–Janjic (Eta) turbulent kinetic energy (TKE)</td>
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<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>
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<tr>
<td>Microphysics</td>
<td>WRF Single-Moment 5-Class (WSM5)</td>
</tr>
<tr>
<td>Cumulus parameterization</td>
<td>Kain–Fritsch (new Eta) parameterization</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>
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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:

$$\text{Fr} = U \left[ \frac{gH}{\bar{\Theta} \sum_{i} (\Theta' - \Theta)} \right]^{-0.5}$$.  (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

$$\text{Fr} = U \left[ \frac{g}{\bar{\Theta} \sum_{i} (\Theta' - \Theta)} \Delta \Theta \right]^{-0.5}$$.  (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 $\Delta \Theta$ 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. 1 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 ($\text{Fr} < 1$) or can pass over the barrier ($\text{Fr} > 1$).

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.

b. 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. 1) 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 m s$^{-1}$, 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. 4c) 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. 4c 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. 4a,b,d show the pressure, temperature, and wind direction comparisons, respectively. These observations also indicate that AMPS modeled the high wind event 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.

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.

Figures 5a,b show the AMPS forecasts of SLP and 10-m winds from 0300 UTC 5 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. 5d). 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. 5a) indicates that a synoptic-scale low pressure system was located off the coast of West Antarctica. Figure 5c 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 −25°C to −24°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. 5d). Additionally, the enhanced cloud signature associated with the cyclone and frontal band over West Antarctica can be seen in the satellite infrared image (Fig. 5c). 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.

b. 1200 UTC 5 September 2009: 12 h prior to wind event

The AMPS-forecast SLP plot (Fig. 6a) 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. 4a) is validated by the Sabrina AWS pressure observations (shown by the solid line in Fig. 4a), 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. 6b). This also induced a downslope flow from West Antarctica into the region of the Siple Coast confluence zone (Fig. 6b). 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.

c. 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. 7a). 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 8a–d show the evolution of the 500-mb geopotential heights from 0300 UTC 5 September to 0300 UTC 6 September 2009. Figure 8a shows an area of upper-level low geopotential heights located over the southern tip of the RIS. Moving forward in time, Figs. 8b,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. 8d 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.

The movement of the upper-level wave pattern, as shown in Figs. 8a–d, also contributed to the increase in SLP to the southeast of the Prince Olav Mountains by directing flow from the northeast toward the Transantarctic Mountains. The plots of the AMPS-forecast 700-mb geopotential heights and the 700-mb winds for 2100 UTC 5 September 2009 (Figs. 9a,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 10 shows plots of the cross section parallel winds (Fig. 10a), perpendicular winds (Fig. 10b), Froude number (Fig. 10c), and potential temperature (Fig. 10d) in this region (Fig. 1 shows the location of the cross

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**Fig. 8.** 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.
section) for this time period. In Fig. 10a, 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 10b 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 10d 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. 10b in the region adjacent to the mountains. Figure 10c 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. 10 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.

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. 7a. 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. 7b and the further development of the barrier wind will be discussed in the next section.

d. 0300 UTC 6 September 2009: Barrier wind strengthening

The AMPS-forecast SLP and 10-m wind plots (Figs. 11a,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

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**Fig. 9.** AMPS forecast of (a) 700-mb geopotential heights and (b) 700-mb winds for 2100 UTC 5 Sep 2009. The black dots in the AMPS plots indicate the location of the Sabrina and Eric AWS locations.
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. 11b). 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. 8d) 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. 11b. Additionally, it can be seen that these strong winds have reached the location of the Sabrina AWS (Fig. 11b), initiating the start of the high wind event at Sabrina (Fig. 4c).

e. 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. 4c). The SLP plot for this time period (Fig. 12a) 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. 1) and can be seen in the plot of SLP at this time. The 10-m wind plot for this time period (Fig. 12b) 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.

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. 12a,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. 13c,d with the corresponding SLP and 10-m winds shown in Figs. 13a,b. Figure 14 shows the cross-jet and along-jet forces for the cross section depicted in Figs. 13a,b.

The PGF plot shown in Fig. 13c 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. 13d) 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. 13b) shows the presence of the barrier wind off the coast of the Prince Olav Mountains.

The SLP plot for 2100 UTC 6 September 2009 (Fig. 13a) 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. 13b,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. 13a) 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. 13b. 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. 14a,b show the individual terms of the horizontal momentum equations along the cross section shown in Figs. 13a,b. The cross section aligns with the strongest 10-m winds within the jet (Fig. 13b). 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. 14a) and parallel (Fig. 14b) to the jet and is similar to the method used by Outten et al. (2009). For the along-jet force balance (Fig. 14b) forces that are positive are directed in the direction of the wind vector, and for the cross-jet force balance (Fig. 14a) 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. 14a), 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

FIG. 12. As in Fig. 6, but for 2100 UTC 6 Sep 2009.
increased in magnitude and was negative throughout the remainder of the jet.

In the analysis of the along-jet forces (Fig. 14b), 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. 13a) 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. 13a) and accelerated the jet (Fig. 13b). 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.

Fig. 13. (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.
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. 13b). 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. 13b–d). The turning of the winds in this region directed the low-level flow toward the southeastern base of the Queen Alexander Mountains (Fig. 13b). 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. 13a). As shown in Figs. 13a,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. 13c). 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. 13b). 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. 1) 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. 15) 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 Transantarctic Mountains. Additionally, the AMPS wind speed forecasts (shown by the dashed line in Fig. 15) 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.

![Fig. 15. Eric AWS wind speed observations (solid line) and AMPS forecast of 10-m wind speeds (dashed line) for a portion of September 2009.](image-url)
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 Alexandra 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.

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