Analysis of the Ross Ice Shelf Airstream Forcing Mechanisms Using Self-Organizing Maps

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ABSTRACT

The Ross Ice Shelf airstream (RAS), a prominent transport mechanism of cold, continental air to the north, is the most common wind pattern over the Ross Ice Shelf, Antarctica. The forcing mechanisms of the RAS include katabatic drainage, mesoscale forcing, and synoptic forcing. This paper uses the 15-km output from the Antarctic Mesoscale Prediction System (AMPS) and the method of self-organizing maps (SOM) to analyze how the combination of these forcing mechanisms impacts the strength and position of the RAS. It is found that the strength and position of the RAS is mainly driven by the thermal forcing in the region of the Transantarctic Mountains. This forcing includes the pressure gradient associated with cold air pooling at the base of the Transantarctic Mountains, as well as, the pressure gradient associated with the temperature contrast between the cold air located over the East Antarctic Plateau and the warm ambient air over the Ross Ice Shelf. These forcing mechanisms are analyzed in a region near the southern tip of the Ross Ice Shelf. In this region, the pressure gradient associated with the temperature contrast between the East Antarctic Plateau and the ambient air over the ice shelf is usually present during RAS events, while the pressure gradient associated with the cold air pooling varies between RAS events. The analysis shows that, in the region of the southern Ross Ice Shelf, RAS events can occur without the presence of cold air pooling.

1. Introduction

The Ross Ice Shelf airstream (RAS), the most common wind pattern over the Ross Ice Shelf (RIS) (Fig. 1a) (Seefeldt and Cassano 2012; Nigro and Cassano 2014, hereafter NC14), is one of the prominent transport mechanisms of cold, continental air from the interior of the Antarctic continent to lower latitudes (Parish and Bromwich 1998; Parish et al. 2006; Seefeldt and Cassano 2012). The origin of the RAS is located in the Siple Coast confluence zone, which is a large drainage basin of cold, continental air from both East and West Antarctica (Bromwich and Liu 1996; Parish and Bromwich 1998). The RAS transports the cold, continental air from the base of the Siple Coast confluence zone, over the western or central RIS, to the Ross Sea (Bromwich et al. 1992, 1993; Parish and Bromwich 1998; Parish et al. 2006; Seefeldt and Cassano 2012). The drainage of cold, continental air from the interior of the continent to lower latitudes is a major component of the thermally direct meridional circulation in the Southern Hemisphere (Parish 1992; Parish and Bromwich 1987; Parish and Cassano 2001; Parish and Bromwich 2007), therefore, making the RAS an important part of the Southern Hemisphere climate system.

The RAS, which flows parallel to the Transantarctic Mountains, is driven by a combination of katabatic drainage, mesoscale, and synoptic forcing (Parish et al. 2006; Nigro et al. 2012; Seefeldt and Cassano 2012). Katabatic winds are forced by negatively buoyant air flowing over sloped terrain (Parish 1988). In the region of the RIS, katabatic winds flow through the Siple Coast confluence zone and the glacial valleys of the Transantarctic Mountains. These winds drain the cold, continental air from the interior of the continent onto the RIS, providing a source of cold, continental air and momentum to the RAS.

The mesoscale and synoptic forcing mechanisms of the RAS include barrier winds, synoptic and mesoscale cyclones located in the region of the RIS and the Ross Sea (Parish et al. 2006; Nigro et al. 2012; Seefeldt and Cassano 2012), and associated fronts. In general, these
forcing mechanisms are driven by horizontal pressure gradients that occur in the region of the RIS.

Barrier winds form when stably stratified flow is directed toward a barrier. If the flow directed toward the barrier does not possess enough kinetic energy to traverse the barrier, the flow is blocked and mass convergence occurs. The mass convergence creates a region of cold air and high pressure at the base of the barrier. The region of high pressure induces a pressure gradient force (PGF) directed away from the barrier, which drives the wind. This wind becomes approximately geostrophic, in the cross-barrier direction and flows parallel to the barrier with the barrier located to the left of the wind in the Southern Hemisphere (O’Connor et al. 1994; Parish et al. 2006; Seefeldt et al. 2007; Steinhoff et al. 2009). Barrier winds are located within the Rossby radius of deformation of the barrier. The Rossby radius of deformation describes the horizontal length over which the atmosphere adjusts due to flow blocked by a barrier and, therefore, determines the distance away from the mountains that the barrier wind can form (King and Turner 1997).

In the region of the RIS, barrier winds form when the semipermanent synoptic cyclone located off the coast of West Antarctica directs flow over West Antarctica, toward the Transantarctic Mountains (Nigro et al. 2012). If this flow is blocked, mass convergence occurs and a barrier wind forms along the base of the Transantarctic Mountains. This barrier wind is a southeasterly wind that flows parallel to the Transantarctic Mountains.

Forcing from synoptic and mesoscale cyclones in the region of the RIS and the Ross Sea provide forcing for the RAS by altering the magnitude and orientation of the pressure gradient over the RIS (Parish et al. 2006; Nigro et al. 2012; Seefeldt and Cassano 2012). The presence of cyclones in these regions lowers the pressure over the RIS and the Ross Sea. This typically enhances the pressure gradient over the RIS, with high pressure located adjacent to the mountains and low pressure located away from the mountains. This pressure gradient induces a geostrophic wind that is roughly parallel to the mountains, which is consistent with the path of the RAS. Parish et al. (2006) and Seefeldt and Cassano (2012) concluded that the semipermanent synoptic-scale cyclone located off the coast of West Antarctica is the dominant forcing mechanism driving the RAS.

This paper will use the results of the self-organizing maps (SOM) analysis presented by NC14 to analyze the forcing mechanisms (katabatic drainage, mesoscale forcing, and synoptic forcing) of the RAS. NC14 used the method of SOMs to identify the dominant RAS patterns that occur over the RIS (see NC14, their Fig. 5a). The RAS patterns were defined as patterns with a dominant stream of southerly flow over either the western or central RIS. The results showed variations in the strength and position of the RAS. This study will analyze how the contribution of the individual forcing mechanisms of the RAS (katabatic drainage, mesoscale forcing, and synoptic forcing) influences the variation of the strength and position of the RAS. The analysis investigates the thermal forcing of the PGF that drives the RAS and is conducted on a section of the RAS in the southern RIS (see Fig. 1a for location of cross section and Fig. 1b for the topography along the cross section).

In the southern RIS, the RAS winds are both barrier
parallel and located within the Rossby radius of deformation of the Transantarctic Mountains, indicating the potential for forcing from both katabatic drainage and barrier winds in this region.

2. Data

The RAS SOM analysis presented by NC14, used output from the polar-modified version of the Weather Research and Forecasting (WRF) Model run within the Antarctic Mesoscale Prediction System (AMPS; Powers et al. 2012). AMPS is run twice daily and the output is archived and available through the National Center for Atmospheric Research. The archived 3-hourly AMPS output from the AMPS continental domain was used for the RAS SOM analysis. This data have 15-km grid spacing and 44 terrain-following eta levels. The 12–21-h forecasts from October 2008 to September 2010 were used. The 12–21-h forecasts were used in order to allow the model sufficient time to adjust from the initial conditions (Bromwich et al. 2005; Seefeldt and Cassano 2008). [For additional information on this AMPS dataset see NC14, Powers et al. (2012), and the AMPS website at http://www.mmm.ucar.edu/rt/amps.]

NC14 used automated weather station (AWS) observations to evaluate the accuracy of the AMPS modeled low-level winds over the RIS. Over the 2-yr 15-km AMPS-WRF dataset, the model root-mean-square error (RMSE) was 2.8 m s$^{-1}$ for the $u$ component of the wind, 2.9 m s$^{-1}$ for the $v$ component of the wind, 2.8 m s$^{-1}$ for the wind speed, and $-6^\circ$ for the wind direction. The correlation was 0.58 for the $u$ component of the wind, 0.72 for the $v$ component of the wind, and 0.69 for the wind speed (NC14). Therefore, AMPS reasonably predicts the low-level wind field over the RIS. Because of the lack of observations in Antarctica, the 15-km AMPS-WRF output will be used to diagnose the forcing mechanisms of the RAS in this study.

3. Methods

NC14 used the method of SOMs to identify the dominant RAS patterns that occur 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 (Kohonen 2001). The analysis used the AMPS forecast $u$ and $v$ components of the 10-m wind over the RIS for the SOM training. The authors determined that a $5 \times 4$ SOM grid was sufficient for capturing each of the RAS patterns that occur over the RIS. Figure 2a shows the 10-m wind pattern identified by the NC14 RAS SOM analysis. In this figure, each pattern is referred to as a node and is referenced by its column number (zero is the leftmost column) and its row number (zero is the top row). See NC14 for additional details on the method of SOMs.

For the analysis presented in this paper, the individual forcing mechanisms (katabatic drainage, mesoscale forcing, and synoptic forcing) will be analyzed for each RAS pattern identified by the NC14 RAS SOM analysis. The katabatic force (KF) is calculated using the following equation:

$$
KF = g \frac{d}{\theta_o} \sin \alpha,
$$

where $g$ is the acceleration due to gravity, $d$ is the potential temperature deficit, $\theta_o$ is the background potential temperature, and $\alpha$ is the terrain slope. The potential temperature deficit is calculated following the method described by Parish and Cassano (2001) by fitting a regression line to the potential temperature profile of the free atmosphere. The difference between the environmental potential temperature and the potential temperature calculated from the regression line at a given height is the potential temperature deficit. For the work presented in this paper, the free atmosphere was defined as the layer between 1000 and 2000 m above ground level. These values were chosen using the average potential temperature profile from the October 2008 to September 2010 dataset and are consistent with the range of values used by previous studies (Parish and Cassano 2001, 2003a,b). The potential temperature deficit was calculated at the height of the lowest eta level. Therefore, the background potential temperature $\theta_o$ was set to the potential temperature at the lowest eta level. For plotting purposes, the KF was transformed into an equivalent geostrophic wind by dividing the KF by the Coriolis parameter and multiplying by negative one.

The synoptic and mesoscale forcing mechanisms (barrier winds, synoptic cyclones, and mesocyclones) are analyzed using a combination of the Rossby radius of deformation and a cross-sectional analysis of the RAS in the southern RIS. The Rossby radius of deformation is used in this study to determine the approximate maximum width of potential barrier winds 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 calculated using the following equation from King and Turner (1997):

$$
R_R = \left( \frac{1}{f} \left( gH \frac{\Delta \theta}{\theta} \right)^{0.5} \right),
$$

where $f$ is the Coriolis parameter, $g$ is the acceleration due to gravity, $H$ is the height is the potential temperature deficit. For the work presented in this paper, the free atmosphere was defined as the layer between 1000 and 2000 m above ground level. These values were chosen using the average potential temperature profile from the October 2008 to September 2010 dataset and are consistent with the range of values used by previous studies (Parish and Cassano 2001, 2003a,b). The potential temperature deficit was calculated at the height of the lowest eta level. Therefore, the background potential temperature $\theta_o$ was set to the potential temperature at the lowest eta level. For plotting purposes, the KF was transformed into an equivalent geostrophic wind by dividing the KF by the Coriolis parameter and multiplying by negative one.

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R_R = \left( \frac{1}{f} \left( gH \frac{\Delta \theta}{\theta} \right)^{0.5} \right),
$$
where $f$ is the Coriolis parameter, $H$ is the height of the barrier, $\Delta \theta$ is the change in potential temperature over the depth of the layer of blocked flow, and $\theta$ is the average potential temperature of the layer of blocked flow. To determine the maximum possible width of a barrier wind, the Rossby radius of deformation was calculated with the assumption that the layer of blocked flow extended from the surface to the top of the mountain (the maximum possible height of the blocked flow).

The SOM node-averaged cross-sectional perpendicular winds will be analyzed to determine if the RAS is located within the Rossby radius of deformation of the Transantarctic Mountains. If the RAS is located within the Rossby radius of deformation and below the height of the mountain, it may be driven by barrier wind dynamics. To determine if the RAS is located within the Rossby radius of deformation of the barrier, the edges of the RAS must be defined. Using the method presented by Olson and Colle (2009), the horizontal and vertical extents of the jet are defined as the location where the maximum speed in the jet has decreased by a factor of $e^{-1}$. Using this method and the vertically interpolated model cross-sectional perpendicular winds, the edge of the jet is evaluated every 10 m from the surface up to the top of the RAS.

As mentioned, the synoptic and mesoscale forcing mechanisms of the RAS are generally driven by horizontal

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**Fig. 2.** (a) Node-averaged 10-m wind speeds. The black dots indicate the location of the cross section and the black asterisk indicates the location of the Rossby radius of deformation with respect to the Transantarctic Mountains. (b) Node-averaged cross-sectional perpendicular wind speeds. Colored contours indicate southeasterly flow into the cross section. Hatched contours indicate northwesterly flow out of the cross section. The horizontal black line indicates the top of the mountain. The vertical black line indicates the location of the Rossby radius of deformation. The black contour line indicates the edge of the jet. White regions show areas of topography along the cross section. The $x$ axis is plotted as horizontal distance in kilometers and the $y$ axis is plotted as height above the ground in meters. In (a) and (b), the colored outlines around the nodes indicate the groups discussed in the text [group 1 (red), group 2 (blue), and group 3 (green)].
pressure gradients. For the analysis presented in this paper, the PGF at the surface, the PGF in the layer below the top of the mountain, and the PGF in the layer above the top of the mountain are analyzed for the cross section in the southern RIS in each SOM identified RAS pattern. The surface PGF provides information about the forcing for the RAS, while the PGFs in the layers above and below the top of the mountain provide information about the dynamics in the layers above (no barrier wind forcing) and below (potential barrier wind forcing) the top of the mountain. Other topographic forcing mechanisms, such as gravity waves, were not analyzed.

For these PGF calculations, the surface is defined as a constant elevation that is representative of the surface height along the length of the entire cross section. The layer below the top of the mountain is defined as the layer from the surface to the top of the mountain. The layer above the top of the mountain is defined as the layer from the top of the mountain to the height at which the layer above the mountain has the same depth as the layer below the top of the mountain. The surface PGF is calculated from surface pressures that have been interpolated to the constant surface height along the cross section. In each of the two layers, the PGF is calculated using the pressure differences between the bottom and the top of the layer.

To determine how much of the forcing is coming from either above or below the top of the mountain, the surface PGF is correlated with the PGF in the layers above and below the top of the mountain. A strong correlation between the surface PGF and the PGF in one of the two layers indicates that the dynamics in this layer strongly contributes to the forcing of the RAS at the surface. If the correlation is strong with the layer above the top of the mountain and low in the layer below the top of the mountain, the RAS is not driven by barrier wind dynamics and if the correlation is strong with the
layer below the top of the mountain, the RAS is potentially forced by barrier wind dynamics.

4. Results

a. RAS patterns

Figure 2a of the node-averaged 10-m winds shows the various positions, widths, and strengths of the RAS. Moving from left to right across the SOM the RAS becomes narrower and shifts westward toward the mountains. Moving from top to bottom across the SOM the winds shift westward toward the mountains and become more southeasterly, or more terrain parallel. From top to bottom, the strength of the RAS weakens except in column 0.

Figure 2b of the cross-sectional perpendicular winds, shows the various widths, heights, and strengths of the RAS. The solid contours show winds into the page (the direction of the RAS) and cross-hatched contours show winds out of the page. The vertical black line shows the location of the Rossby radius of deformation and the horizontal black line shows the height of the mountains in the cross section. The thick, black contour indicates the edges of the RAS as defined by the e-folding length (Olson and Colle 2009) and if the RAS is below the height of the mountains and within the Rossby radius of deformation this may reflect barrier wind forcing. In general, the RAS in this cross section is located adjacent to the mountains and within the Rossby radius of deformation for all nodes. Moving from left to right across the SOM, the width of the RAS decreases. Moving from top to bottom across the SOM the wind speeds within the RAS decrease. The deepest RAS are located in the top-right corner of the SOM and the shallowest RAS are located in the bottom-left corner of the SOM.

b. Katabatic forcing

Figure 3 shows the equivalent geostrophic winds with respect to KF for each node. The winds that are shown are stronger than the actual winds because turbulence and advection act to reduce the speed of the katabatic winds. The KF indirectly contributes to the RAS because KF only exists over sloping terrain and, therefore, does not exist over the nearly flat RIS where the RAS is located. The katabatic drainage from the glacial valleys of the Transantarctic Mountains feeds into the RAS, providing a source of cold, continental air and momentum for the RAS. The winds shown in Fig. 3 indicate that all of the nodes have some KF. The strength of the KF varies across the nodes, indicating that the contribution of cold air and momentum to the RAS by the KF varies across the RAS patterns. Each node has strong KF over the steep topography of the Transantarctic Mountains and very little to zero KF over the nearly flat RIS. The nodes in the top left have the strongest KF. Moving from top to bottom and from left to right across the SOM the KF decreases, indicating that KF contributes the most to the RAS patterns in the top left of the SOM.

c. Synoptic and mesoscale forcing

This analysis is conducted over a cross section in the southern RIS (location shown in Fig. 1a). In the southern RIS the RAS is both barrier parallel and located within the Rossby radius of deformation (the location of the Rossby radius of deformation is indicated by a black asterisk in Fig. 2a), indicating the RAS has the potential for barrier wind forcing in this region. The cross-sectional analysis allows for the investigation of the horizontal pressure gradients that drive the synoptic and mesoscale forcing mechanisms of the RAS.

Assuming a hydrostatic atmosphere, the horizontal temperature gradients that occur in the region of the RIS create pressure gradients that contribute to the forcing of the RAS. Therefore, throughout the remainder of the paper, the thermal forcing of the RAS will be described in terms of the temperature gradients in the region of the RIS. Figure 4 shows the horizontal temperature anomalies (calculated with respect to the average temperature within the width of the RAS, using vertically interpolated model temperatures every 10 m from the surface up to 6000 m) that occur along the cross section for each SOM identified RAS pattern. All of the nodes show cold air located over the East Antarctic Plateau and adjacent to the Transantarctic Mountains (the blue colors on the left side of the cross section) and warm air over the RIS (the red colors in the center of the cross section). In each cross section, the thick, black contour indicates the edges of the RAS shown in Fig. 2. This shows that the RAS flows with the cold air to the left and the warm air to the right of the wind in each of the SOM identified RAS patterns.

A cold column of air with a larger vertical pressure difference from bottom to top of the column than a warm column of air with an equivalent thickness is denser and has a higher pressure, creating a PGF at the surface directed from the cold column toward the warm column. This PGF induces a wind directed from the cold column toward the warm column. If the conditions are such that the wind is geostrophic (Coriolis force balances the PGF), the wind flows with the cold column to the left and the warm column to the right of the wind in the Southern Hemisphere.

The relationship between the horizontal temperature gradients and the pressure gradients that drive the RAS can be further investigated using the cross-sectional
plots in Fig. 5, which analyze the temperature, pressure, and PGF in the layers below and above the top of the mountain. Figure 5a shows the layer-average temperature anomalies (calculated with respect to the average temperature within the width of the RAS) for the layer above (the pink dashed line) and the layer below (the blue dashed line) the top of the mountain. Figure 5a shows that within the Rossby radius of deformation (the region to the left of the vertical dashed black line in Fig. 5a), the temperature increases with distance away from the mountains in both the layer above and below the top of the mountain for each of the SOM-identified RAS patterns. This is consistent with the horizontal temperature anomalies shown in Fig. 4.

As mentioned above, the increase in temperature with distance away from the mountains in each layer should indicate a decrease in pressure with distance away from the mountains in each layer. Figure 5b shows the pressure anomalies for the layer above (the pink dashed line) and the layer below (the blue dashed line) the top of the mountain, as well as, the surface pressure anomaly (the blue solid line). The pressure anomalies for the layers above and below the top of the mountains are calculated using the pressure differences between the bottom and top of the layer. Figure 5b shows that within the Rossby radius of deformation (the region to the left of the vertical dashed black line in Fig. 5b), the pressure decreases with distance away from the mountains in both the layer above and below the height of the mountain, as well as in the surface pressure. This is consistent with the colder column of air adjacent to the mountains having a higher pressure than the warmer column of air over the RIS.

The decrease in pressure with distance away from the mountains creates a PGF in this region. Figure 5c shows...
the PGF for the layer above the top of the mountain (the pink dashed line), the layer below to the top of the mountain (the blue dashed line), and the surface (the blue solid line). Figure 5c shows that within the Rossby radius of deformation (the region to the left of the vertical dashed black line in Fig. 5c) the PGF in the layers above and below the height of the mountain and the PGF at the surface are positive. A positive PGF in this region is directed away from the mountains, or from the column of cold air toward the column of warm air. A PGF directed away from the mountains would induce a geostrophic wind that flows with the Transantarctic Mountains to the left of the wind, or with the column of cold air to the left and the column of warm air to the right. This is consistent with the position and direction of the RAS (Fig. 2b). The temperature, pressure, and PGF analysis (Fig. 5) indicates the RAS is driven by the horizontal temperature gradients that occur in the region of the RIS.

The remainder of this section will investigate how the variation of the temperature field between the RAS patterns is related to the variation of the strength and the position of the RAS in the SOM-identified patterns. For this discussion, the nodes are subjectively broken into groups according to the forcing mechanisms that drive the RAS for each group. In Figs. 2–5, the patterns in group 1 are outlined in red and include nodes [0, 0] to [1, 3] (the two leftmost columns), the patterns in group 2 are outlined in blue and include nodes [2, 0] to [4, 2] with the exception of node [3, 1], and the patterns in group 3
are outlined in green and included nodes [2, 3] to [4, 3] and node [3, 1].

Each of the groups will be analyzed using the plots in Fig. 5. In Fig. 5a the numbers in the top left of each node show the temperature gradient over the width of the jet for the layer above (pink) and below (blue) the top of the mountain and the difference (layer below minus layer above) between the two temperature gradients (black). In Fig. 5b the numbers in the top right of each node show the pressure gradient over the width of the jet for the layer above (pink) and below (blue) the top of the mountain and the difference (layer below minus layer above) between the two pressure gradients (black). In Fig. 5c the numbers in the top right of each node show the correlation (over the Rossby radius of deformation) between the surface PGF and the PGF in the layer above the top of the mountain (pink), the correlation between the surface PGF and the PGF in the layer below the top of the mountain (blue), and the difference (layer below minus layer above) between the two correlations (black).

1) GROUP 1 PATTERNS

The RAS patterns in group 1 have a RAS that is located within the Rossby radius of deformation and below the height of the mountain (with the exception of node [0, 0]) (Fig. 2b). The PGF correlations (Fig. 5c) indicate that the RAS patterns in group 1 have a higher correlation (>0.89) between the surface PGF and the PGF in the layer below the top of the mountain (blue) than between the surface PGF and the PGF in the layer above the top of the mountain (<0.77) (pink). The differences between the correlations, which are shown in black, indicate that the correlation with the layer below the top of the mountain is at least 0.19 greater than with the layer above the top of the mountain. The strong correlation between the surface PGF and the PGF in the layer below the top of the mountain indicates the dynamics in the layer below the top of the mountain dominate the forcing of the RAS in these patterns. The positive PGF in the layer above the top of the mountain (pink dashed line), indicates that the dynamics in the layer above the top of the mountain also contribute to the forcing of the RAS, but are not the dominant forcing mechanism of the RAS in these patterns.

The difference between the forcing mechanisms in the layers below and above the top of the mountain will be investigated using the plots in Fig. 5. Figure 5a shows that, for the patterns in group 1, the temperature gradient over the width of the jet is larger in the layer below the height of the mountain (ranging from 0.0072 to 0.0144 K km$^{-1}$) than in the layer above the height of the mountain (ranging from 0.0048 to 0.0111 K km$^{-1}$), with the difference between the two gradients ranging from 0.0009 to 0.0047 K km$^{-1}$. Figure 5a also shows that the rate of the temperature change with distance away from the mountain (within the Rossby radius of deformation) differs in the layers above and below the top of the mountain. In the region closest to the Transantarctic Mountains, the temperature increases with distance away from the mountains at a slower rate in the layer below the top of the mountain than in the layer above the top of the mountain (Fig. 5a). Moving farther away from the mountain, but staying within the Rossby radius of deformation, the temperature increases at a faster rate in the layer below the top of the mountain than in the layer above the top of the mountain. This type of temperature change is consistent with cold air pooling against the Transantarctic Mountains. When cold air is pooled against a mountain, the layer-averaged temperature below the height of the mountain remains fairly constant or increases slightly with distance away from the mountain until the edge of the cold pool is reached, at which point, the temperature increases quickly [see Fig. 22 in Bell and Bosart (1988) and Fig. 3 in Xu et al. (1996)]. Additionally, for cold air pooling, the temperature change below the height of the mountain is expected to be larger than the temperature change above the height of the mountain. Therefore, the temperature anomalies shown in Fig. 5a indicate that the RAS patterns in group 1 have cold air pooled against the Transantarctic Mountains.

The pressure changes in the layer below the height of the mountain, shown by the blue dashed line in Fig. 5b, are consistent with the temperature changes in Fig. 5a. In the region closest to the mountain, the pressure in the layer below the height of the mountain decreases at a slower rate than the layer above the height of the mountain. Moving farther away from the mountain, the pressure in the layer below the height of the mountain decreases at a faster rate than in the layer above the mountain. This type of pressure change is consistent with the concept of cold air pooled against the Transantarctic Mountains, where the pressure in the layer below the top of the mountain decreases only slightly through the cold pool region and then decreases quickly at the edge of the cold pool (Bell and Bosart 1988; Xu et al. 1996).

The pressure gradient in the layer below the top of the mountain for the RAS patterns in group 1 (Fig. 5b) induces a PGF directed away from the mountains. This is shown by a positive PGF in the blue dashed line in Fig. 5c. Figure 5c shows that in the layer below the height of the mountain (blue dashed line), the PGF is small closest to the mountain, or in the region of the cold pool where the pressure decreases only slightly with distance away from the mountain. The PGF reaches its
FIG. 5. (a) Node-averaged cross section of layer-averaged temperature anomalies (K) (with respect to the average temperature within the width of the jet) in the layer below the mountain (dashed blue line) and in the layer above the mountain (dashed pink line). The numbers in the top-left corner show the temperature gradient (K km⁻¹) over the width of the jet in the layer above the mountains (pink), in the layer below the mountains (blue), and the difference between the two temperature gradients (layer below minus layer above) (black). Positive differences indicate larger temperature gradient in the layer below the top of the mountain. In (a), (b), and (c), the vertical dashed line indicates the location of the Rossby radius of deformation. The x axis is plotted as horizontal distance in kilometers. The edge of the Transantarctic Mountains is located at approximately 200 km. Colored outlines around the nodes indicate the groups discussed in the text [group 1 (red), group 2 (blue), and group 3 (green)].
FIG. 5. (b) Node-averaged cross-sectional pressure anomalies (hPa) (with respect to the average pressure along the cross section) at the surface (solid blue line), in the layer below the mountain (dashed blue line), and in the layer above the mountain (dashed pink line). The numbers in the top-right corner show the pressure gradient (hPa km$^{-1}$) over the width of the jet in the layer above the mountains (pink), in the layer below the top of the mountain (blue), and the difference between the two pressure gradients (layer below minus layer above) (black). Negative difference indicates larger pressure gradient in the layer below the top of the mountain.
Fig. 5. (c) Node-averaged cross-sectional pressure gradient force (m s\(^{-2}\)) at the surface (solid blue line), in the layer below the mountain (dashed blue line), and in the layer above the mountain (dashed pink line). The numbers in the top-right corner show the correlation between the surface PGF and the PGF in the layer above the mountains (pink), the surface PGF and the PGF in the layer below the mountains (blue), and the difference (layer below minus layer above) between the two correlations (black). Positive difference indicates larger correlation of surface PGF with the layer below the top of the mountain.
maximum at a distance away from the mountains, but within the Rossby radius of deformation. The location of the maximum PGF coincides with the location of the edge of the cold pool where the pressure decreases quickly. This indicates that in group 1, the magnitude of the PGF in the layer below the height of the mountain is mainly driven by the dynamics associated with the cold pool of air located adjacent to the Transantarctic Mountains. The strong correlation between the surface PGF and the PGF in the layer below the height of the mountain then suggests that the RAS is mainly driven by the PGF associated with the cold air pooled against the Transantarctic Mountains for the patterns in group 1.

The dynamics associated with the cold air pooled against the Transantarctic Mountains are consistent with the dynamics associated with a barrier wind. These dynamics include cold air and high pressure located adjacent to the mountains, a PGF directed away from the mountains, and a geostrophic wind that flows with the mountains to the left of the wind in the Southern Hemisphere (consistent with the RAS wind patterns shown in Fig. 2b). Additionally, the RAS is located within the Rossby radius of deformation and below the height of the mountain for these patterns (Fig. 2b). The origin of the cold air pooled against the Transantarctic Mountains cannot be determined from the SOM analysis presented here and, therefore, it cannot be determined if the forcing of the RAS in the group 1 patterns is driven by barrier winds. Therefore, we conclude that the RAS patterns in group 1 are mainly driven by the barrier wind–like dynamics associated with the cold air pooled at the base of the Transantarctic Mountains.

The dynamics in the layer above the top of the mountain also provides some forcing for the RAS patterns in group 1 (as shown by the positive PGF in the layer above the mountains for these patterns). In this layer, the temperature increases quickly in the region closest to the plateau or at the edge of the mountains (Fig. 5a). This temperature increase is due to the proximity of the cold air located over the plateau to the relatively warm ambient air over the RIS (due to surface inversion conditions, the atmosphere warms with height above the ground). The influence of the cold air from the plateau enhances the temperature gradient at the edge of the mountains, creating a strong pressure gradient in the region of the RAS (Fig. 5b). The pressure gradient creates a PGF directed from the cold air over the plateau toward the warm air over the RIS (Fig. 5c), enhancing the magnitude of the RAS.

Additionally, the influence of the synoptic-scale pressure gradient on the RAS can be inferred by looking at the PGF in the region outside of the Rossby radius of deformation (Fig. 5c). Essentially, it can be assumed that the PGF outside of the Rossby radius of deformation is free of topographic influences and is therefore driven by the large-scale pressure field of the region. It is then assumed that this forcing is representative of the synoptic forcing within the region of the RAS. In the group 1 patterns, the PGF outside of the Rossby radius of deformation is positive (directed away from the mountains), indicating the synoptic forcing enhances the magnitude of the RAS. The magnitude of the PGF outside of the Rossby radius of deformation varies across the group 1 nodes, indicating that the impact of synoptic forcing on the RAS varies over these nodes. The synoptic forcing ranges from weak in nodes [1, 0] and [1, 1] (where the PGF outside of the Rossby radius of deformation is close to zero) to fairly strong in nodes [0, 1] and [0, 2] (where the average PGF outside of the Rossby radius of deformation is roughly half of the average PGF inside of the Rossby radius of deformation).

In conclusion, the RAS patterns in group 1 are mainly driven by the dynamics in the layer below the height of the mountains (as shown by the strong correlation with the layer below the height of the mountains) and, in some nodes, the large-scale synoptic forcing in the region of the RIS (as shown by the PGF in the region outside of the Rossby radius of deformation). In these patterns, there is strong cold air pooling that provides barrier wind–like forcing for the RAS. In these nodes, there is also some forcing from the layer above the height of the mountain. In the layer above the mountain, the forcing is driven by the PGF associated with the temperature gradient between the cold air located over the plateau and the warm air located over the RIS.

2) GROUP 2 PATTERNS

The RAS patterns in group 2 show a RAS that is located within the Rossby radius of deformation and that extends above the height of the mountain (Fig. 2b). The PGF correlations (Fig. 5c) indicate the RAS patterns in group 2 have a high correlation (>0.86) with the layer below the top of the mountain and a relatively (with respect to group 1) high correlation (>0.75) with the layer above the top of the mountain. The difference between the correlations above and below the top of the mountain is at most 0.12, indicating a smaller difference between the correlations in the group 2 patterns than in the group 1 patterns. The strong correlation with the layers above and below the top of the mountain indicates that the dynamics in the two layers have a similar impact on the forcing of the RAS in the group 2 patterns.

The group 2 layer-averaged temperature anomalies (Fig. 5a) show that the temperature change over the Rossby radius of deformation is fairly similar in the
layers below and above the top of the mountain. The difference between the temperature gradient over the width of the jet in the layers below and above the top of the mountain range from $-0.0035$ to $0.0012 \text{ K km}^{-1}$ (black numbers in Fig. 5a), which is generally smaller than the temperature gradient differences in group 1.

Similar to group 1, the group 2 temperature anomalies in the layer below the top of the mountain are consistent with cold air pooling. In the region closest to the Transantarctic Mountains the layer-averaged temperature remains fairly constant or increases slightly with distance away from the mountains and then increases quickly in the region farther away from the mountains. The temperature gradient in the layer below the top of the mountain is $0.0007$–$0.0038 \text{ K km}^{-1}$, which is smaller than the temperature gradient in the layer below the top of the mountain in the group 1 patterns ($0.0072$–$0.0144 \text{ K km}^{-1}$). This suggests that the amount of cold air pooled against the mountains in group 2 is less than the amount of cold air pooled against the mountains in group 1.

The group 2 pressure changes (blue dashed line in Fig. 5b) and the PGF (blue dashed line in Fig. 5c) in the layer below the height of the mountain are consistent with the temperature changes in the layer below the height of the mountain (blue dashed line in Fig. 5a). The pressure in the layer below the height of the mountain decreases slightly through the region of the cold pool and decreases quickly at the edge of the cold pool. Similar to the patterns in group 1, the PGF in the layer below the height of the mountain is small in the region of the cold pool where the pressure decreases slightly and has a maximum at the edge of the cold pool where the pressure changes quickly. Therefore, the RAS patterns in group 2 are partially driven by barrier wind–like dynamics associated with cold air pooled at the base of the Transantarctic Mountains. The contribution of the cold pool dynamics to forcing the RAS is smaller in the group 2 patterns than in the group 1 patterns due to the reduced cold air pooling and resultant weaker PGF.

In group 2, the dynamics in the layer above the height of the mountains provides additional forcing for the RAS. Similar to group 1, this forcing is associated with the temperature gradient between the cold air located over the East Antarctic Plateau and the warm air located over the RIS. The temperature anomalies for the group 2 patterns show a sharp increase in temperature at the edge of the plateau (Fig. 5a). This coincides with a sharp decrease in pressure (Fig. 5b) and a positive PGF (Fig. 5c) in this region. Therefore, the RAS patterns in group 2 are partially driven by the PGF associated with the temperature gradient between the cold air over the plateau and the warm air over the RIS.

In contrast to group 1, the group 2 PGF values outside of the Rossby radius of deformation (Fig. 5c) are very small or close to zero. This indicates that the large-scale synoptic pressure gradient provides weak forcing of the RAS in these patterns.

In conclusion, the nodes in group 2 are forced by the combination of the barrier wind–like dynamics associated with cold air pooled against the Transantarctic Mountains and the PGF associated with the temperature contrast between the cold air over the plateau and the warm air over the RIS. Differing from group 1 where the forcing below the top of the mountain dominated the forcing of the RAS, neither the forcing from above or below the mountains dominates the forcing of the RAS in group 2. Additionally, the large-scale synoptic pressure gradient of the region provides weak forcing for the RAS in these patterns.

3) GROUP 3 PATTERNS

The RAS patterns in group 3 show a RAS that is located within the Rossby radius of deformation and that extends above the height of the mountain (Fig. 2b). Figure 5c shows that the patterns in group 3, unlike the patterns in groups 1 and 2, have a higher correlation between the surface PGF and the PGF in the layer above the mountains than between the surface PGF and the PGF in the layer below the height of the mountains. The strong correlation between the surface PGF and the PGF in the layer above the mountain indicates that the dynamics in the layer above the mountains dominate the forcing of the RAS in these patterns.

The temperature gradient over the width of the jet in the layer below the top of the mountain for the patterns in group 3 ranges from $-0.0021$ to $0.0001 \text{ K km}^{-1}$ (black numbers in Fig. 5a). This indicates that the temperature below the top of the mountain either remains almost constant or decreases through the width of the jet, providing no evidence of cold air pooling in these RAS patterns. The pressure anomalies in these patterns (Fig. 5b) show a pressure that decreases away from the mountains at a slower rate than the pressure in the layer above the mountain (as shown by the positive pressure gradient differences in black). This is consistent with the small PGF values (Fig. 5c) in the region of the RAS winds in group 3 (the RAS is very narrow in these patterns). Therefore, the layer below the height of the mountain provides little to no forcing for the RAS in the group 3 patterns.

Therefore, the dynamics in the layer above the mountain dominate the forcing of the RAS in the group 3 patterns. Similar to groups 1 and 2, this forcing is driven by the PGF associated with the temperature contrast between the cold air located over the East Antarctic Plateau and the warm air over the RIS. This is
shown by the sharp increase in temperature, the sharp decrease in pressure, and the positive PGF in the region at the edge of the plateau in the layer above the height of the mountain (Fig. 5).

In the group 3 patterns, the PGF is positive in the region outside of the Rossby radius of deformation (Fig. 5c), indicating the synoptic forcing enhances the magnitude of the RAS in these patterns. The magnitude of the PGF outside of the Rossby radius of deformation varies across these nodes, with the PGF in nodes [3, 3] and [4, 3] being slightly stronger than in nodes [3, 1] and [2, 3]. The magnitudes of the PGF outside of the Rossby radius of deformation are approximately equal to the PGF inside of the Rossby radius of deformation for each of these nodes, with the exception of the PGF adjacent to the mountains in the layer above the height of the mountain (pink dashed line). As mentioned above, the peak in the PGF in the layer above the height of the mountains is due to the pressure gradient associated with the temperature contrast between the cold air over the plateau and the warm air over the RIS. The location of the peak in the PGF in the layer above the height of the mountains coincides with the location of the RAS, which is narrow in these patterns (Fig. 2a).

In conclusion, the group 3 patterns are forced by a combination of the PGF associated with the temperature contrast between the cold air located over the East Antarctic Plateau and the warm ambient air over the RIS and the synoptic pressure gradient in the region of the RIS. Differing from groups 1 and 2, the RAS patterns in group 3 have no cold air pooling against the Transantarctic Mountains.

5. Conclusions

This paper used the NC14 RAS SOM results to investigate how the individual forcing mechanisms (katabatic drainage, mesoscale forcing, and synoptic forcing) influence the variation of the strength and position of the RAS. The analysis was conducted over a cross section in the southern RIS, where the RAS is both barrier parallel and located within the Rossby radius of deformation of the mountains.

The KF was calculated for each RAS pattern. The KF is strongest over the steep topography of the Transantarctic Mountains and weakest over the nearly flat RIS. Each RAS pattern has some KF, but moving from top to bottom and from left to right across the SOM the KF decreases. This indicates that KF contributes the most to the RAS patterns in the top left of the SOM.

The thermal forcing associated with the RAS was investigated by analyzing the temperature, pressure, and PGF anomalies in the layers above and below the top of the mountain and at the surface. This analysis concluded that the variation in the strength and position of the RAS is mainly driven by a combination of the PGF associated with cold air pooling against the Transantarctic Mountains, the PGF associated with the temperature contrast between the cold air located over the East Antarctic Plateau and the warm ambient air over the RIS, and the large-scale synoptic pressure gradient. In the region of the southern RIS, the PGF associated with this temperature contrast between the East Antarctic Plateau and the ambient air over the ice shelf was present for all RAS patterns, while the amount of PGF associated with the cold air pooling at the base of the mountains varied between RAS patterns. The RAS patterns in group 1 had the most cold air pooled, the patterns in group 2 had less cold air pooled, and the patterns in group 3 had no cold air pooled against the Transantarctic Mountains. The forcing from the large-scale synoptic pressure gradient also varied across RAS patterns. Groups 1 and 3 both had a synoptic pressure gradient that enhanced the magnitude of the RAS, while the patterns in group 2 had a synoptic pressure gradient that provided very weak forcing for the RAS.

The analysis of the RAS patterns in group 3, shows that the barrier parallel flow, which appears to be a classic barrier wind when considering only the surface wind field, occurred without the presence of cold air pooling. The lack of cold air pooling in these patterns indicates that these RAS patterns are not forced by barrier winds. Instead, the barrier parallel flow is driven by the PGF associated with the temperature gradient between the cold air located over the plateau and the warm ambient air located over the RIS. This temperature gradient persists across all RAS patterns (groups 1, 2, and 3) providing a persistent forcing for the barrier parallel surface winds observed in this region.

The persistent forcing associated with the temperature gradient between the air located over the plateau and the ambient air located over the RIS across all RAS patterns indicates that the PGF at the edge of the plateau usually contributes to RAS forcing. This finding builds on previous research, which indicates the RAS is driven by a combination of katabatic drainage, barrier winds, and synoptic forcing (Parish et al. 2006; Nigro et al. 2012; Seefeldt and Cassano 2012). The SOM analysis presented here indicates that the RAS is usually forced by the PGF associated with the temperature gradient at the edge of the plateau (this forcing is present for all of the SOM identified RAS patterns) and that the contribution of the katabatic drainage, barrier winds, and synoptic forcing varies between RAS events (the magnitude of these three forcing mechanisms varies over the SOM-identified RAS patterns).
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