Diagnosis of the Katabatic Wind Influence on the Wintertime Antarctic Surface Wind Field from Numerical Simulations

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ABSTRACT

Katabatic winds have long been recognized as one of the key climatic variables of the low-level Antarctic environment. Antarctic surface winds display a high degree of persistence with mean directions related to the local topographic configuration of the ice sheet, consistent with katabatic forcing. Continental orography also constrains the atmospheric boundary layer motions through blocking and cold air damming. Finally, the coastal rim about the Antarctic continent is among the most active baroclinic zones on Earth. The establishment of the low-level wind field over Antarctica is thus potentially the result of a number of interacting processes.

To quantify the forcing of the wintertime surface wind field over the Antarctic continent, two numerical strategies are presented. First, idealized numerical simulations are conducted to illustrate the strong orographic control of the low-level wind field. Second, a series of daily numerical simulations using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) has been performed for the midwinter month July 2001. The horizontal pressure gradient as depicted in MM5 was added to the standard output and an analysis was conducted to understand the forcing of the low-level wind field. Horizontal pressure gradients at the lowest sigma level (6 m above the surface) revealed a net forcing primarily down the local topographic fall line. Analyses of the katabatic forcing showed that it was a significant component of the total horizontal pressure gradient force over the interior of the continent. Near the coast and extending several hundred kilometers inland, however, effects of the ambient pressure gradient force were typically comparable to the katabatic forcing and often considerably more important. This suggests that the role of topography in shaping the Antarctic boundary layer winds through blocking and subsequent adjustment is critical to the establishment of the low-level wintertime Antarctic wind field.

1. Introduction

Local topography is vital to the development of the near-surface wind field over the Antarctic continent. From the earliest explorations onto the face of the Antarctic ice sheet, it has been clear that the persistent and often intense surface winds were coupled to the underlying terrain. Winds follow pathways controlled by the local topographic configuration, being deflected to the left of the local downslope direction by 20°–50° owing to the Coriolis force. The linkage between the Antarctic terrain and the surface wind has inspired the use of “katabatic” to characterize the low-level flow regime. Here the term katabatic is used to describe flows that are forced by the radiational cooling of the lower atmosphere adjacent to the sloping terrain. This cooling of the low-level environment implies that the density of the air near the surface is greater than that situated some horizontal distance away from the sloping terrain and hence a horizontal pressure gradient force becomes established in a downslope direction. Katabatic winds are thus directed with a downslope component although, owing to the expanse of the Antarctic ice slopes, Coriolis effects become important. Ball (1960) has shown that the persistent surface wind over the Antarctic continent can be represented by a three-way balance between the forces due to the horizontal pressure gradient, Coriolis force, and friction with a resulting flow direction to the left of the fall line.

Much of the discussion regarding Antarctic winds pertains to katabatic forcing (e.g., Ball 1960; Mather and Miller 1966; Parish 1984; Parish and Bromwich 1987). Stone and Kahl (1991) and Neff (1992, 1999) noted, however, that synoptic-scale disturbances are also critical to the observed behavior of the wind regime at the South Pole. In addition, laboratory experiments by Baines and Fraedrich (1989) and Chen et al. (1993) clearly illustrate the topographic influence of the ele-
Vibrated Antarctic ice sheet on the circulation of the high
southern latitudes. The former noted that the resulting
large-scale circulation about their model of the Antarctic
continent is constrained by the underlying topography
and that the low-level flows tend to follow topographic
countours. Similar strong topographic control of the low-
level Antarctic wind field was also revealed in the ide-
alized numerical simulations described in Parish and
Cassano (2003).

The purpose of this paper is to examine the forcing of
the Antarctic surface wind regime during winter as
simulated by mesoscale models (see also Bintanja
2000). In particular, a budget analysis of the horizontal
pressure gradient force (PGF) as determined from nu-
merical models will be conducted to establish the role
of katabatic forcing in the winter surface winds. An
assessment will be made whether the surface winds
over Antarctic are the result of diabatic cooling of the
terrain slopes and adjacent atmosphere over Antarctica
(katabatic forcing) or due to effects of the pressure gra-
dient force in the ambient atmosphere (ambient or syn-
optic forcing) or some combination of each. This work
extends that of Parish and Cassano (2001) who showed
that the wintertime Antarctic surface wind regime as
diagnosed from the National Centers for Environmental
Prediction–National Center for Atmospheric Research
(NCEP–NCAR) Global Reanalysis Project (Kalnay et
al. 1996) was the result of a combination of both kat-
абatic and ambient forcing. One shortcoming of the
analyses contained in Parish and Cassano (2001) was
that the PGF was not part of the standard output in the
NCEP–NCAR reanalysis and thus had to be estimated.

Here, the primary analyses will be based on a series of
daily wintertime simulations using the fifth-generation
Pennsylvania State University–NCAR Mesoscale Mod-
el (MM5) in which all terms in the prognostic equations
including the PGF have been output. This model has
been used previously in simulating Antarctic processes
(Cassano 1998; Cassano and Parish 2000) and more
recently has been used in real-time mode to provide
twice-daily numerical guidance for Antarctic forecasters
(e.g., Guo et al. 2001, 2003; Powers et al. 2001; Wei
et al. 2001). The simulations discussed here were made
using the same version as that used in the Antarctic
Mesoscale Prediction System (AMPS) real-time appli-
cations. Section 2 contains an examination of the strong
topographic control of the Antarctic surface wind re-
gime using simple model simulations of idealized flow
situations. Results from daily MM5 Antarctic simula-
tions and analyses of the forcing of the surface wind
regime are presented in section 3. A summary is in-
cluded as section 4.

2. Topographic control of the Antarctic wind
regime

Laboratory experiments conducted by Baines and
Fraedrich (1989) and Chen et al. (1993) show that the
elevated Antarctic continent strongly constrains tropo-
spheric circulations. Rotating tank experiments de-
scribed in Baines and Fraedrich (1989) consist of a near-
steady clockwise moving fluid with an imposed coun-
terclockwise-moving topographic representation of Ant-
arctica. This replicates a mean cyclonic circulation that
is imposed on the continent. Cyclonic eddies, similar in
position to those revealed in mean 700-hPa isobaric
maps, were produced in the rotating tank experiments,
suggesting that the mean tropospheric circulation pat-
tern about the Antarctic continent is strongly influenced
by topographic forcing.

Figure 1 depicts the general topography of the Ant-
arctic ice sheet and selected geographical features. As
noted in King and Turner (1997, hereafter KT), the
sparse observational dataset and the elevated continental
orography combine to complicate estimates of mean sea
level (MSL) pressure over Antarctica. MSL pressures
over the interior of the continent are the result of the
reduction of surface pressures to sea level and often the
values are highly suspect. Maps shown in KT depict a
surface anticyclone over the East Antarctic plateau
throughout the year although the authors caution that
such a representation is only a qualitative indication of
the circulation. The cold core nature of the surface high
pressure implies a decrease in the anticyclonic vorticity
with height and a gradual transition to a cyclonic cir-
culation at upper levels. Mean 300-hPa maps over the
Antarctic continent shown in KT reveal a cyclonic cir-
culation although geostrophic wind speeds are much less
than seen to the north of the continent.

Simple numerical experiments have been conducted
to examine the adjustment of the mean circulation by
the Antarctic orography using the hydrostatic primitive
equation model described in Parish et al. (1997). The
same model has been used previously (Parish 1984; Par-
ish and Waight 1987; Parish and Bromwich 1991) in
simulations of Antarctic processes and studies of kat-
абatic winds. Details of the model can be found in Parish
and Waight (1987). Here, it is sufficient to note that the
model is written in sigma coordinates and contains prog-
nostic equations for the two horizontal scalar equations
of motion, temperature, and pressure. Explicit longwave
radiation is simulated following Cerni and Parish (1984)
and the boundary layer parameterization is from Brost
and Wyngaard (1978) with fluxes of heat and moment-
tum determined from the similarity theory of Businger
et al. (1971). The model domain considered consists of
a 110 × 110 horizontal grid points centered on the South
Pole with a grid spacing of 60 km. This covers the entire
continent and extends into the Southern Ocean. It is
assumed for the purposes of these simulations that the
ocean is ice covered. Ten vertical levels are used with the
first approximately 10 m above the surface.

Two numerical experiments will be discussed. First,
a 24-h control simulation of wintertime katabatic winds
was conducted. This numerical experiment was initial-
ized about a state of rest with the PGF at all levels set
to zero. Effects of moisture and solar radiation were not considered. The vertical temperature structure consists of a constant Brunt–Väisälä frequency of 0.0125 with a 1000-hPa temperature of 258 K. At the start of the model simulation, therefore, no surface temperature inversion was specified. The primary forcing mechanism for this control simulation is the longwave radiative cooling at the surface and attendant cooling of the low levels of the overlying atmosphere. Diabatically cooled air over sloping terrain implies the existence of a horizontal temperature gradient. A downslope-directed PGF in the lowest model levels over the sloping Antarctic ice surface thus becomes established. As the low-level flows accelerate in the direction of the terrain-induced PGF, a continental katabatic wind circulation becomes developed within a few hours form the start of the simulation. By 24 h, near steady conditions prevail.

Figure 2a illustrates the 24-h wind vectors and streamlines at the lowest sigma level from the first simulation, which represents a purely katabatic-driven wind system. Winds are clearly influenced by the terrain, with flows directed with a downslope component deflected to the left of the fall line by approximately 20° near the steep coastal perimeter to 50° over portions of the high interior of the continent owing to effects of the Coriolis force. Wind speeds at the lowest sigma level are proportional to the underlying terrain with strongest winds seen near the steep coastal margin. Wind speeds near the coast are typically 10–15 m s⁻¹, which seems reasonable based on the long history of observations at coastal stations. Streamlines match well with that depicted in Parish and Bromwich (1987). Streamline confluence is present along the periphery of the continent at several locations such as onto the Ross Ice Shelf near 160°W, upslope of Adélie Land (140°E), and onto the Amery Ice Shelf (70°E), in agreement with previous depictions of the mean low-level winter winds over Antarctica.

PGF vectors are shown in Fig. 2b and can be seen to be directed downslope at nearly every grid point. The magnitudes of the PGF vectors shown are computed by dividing the PGF by the Coriolis parameter and thus the vector lengths represent the geostrophic wind magnitude. Since no initial PGF was present at the start of the model simulation, the vectors represent the effect of pure katabatic forcing. It can be seen that the magnitude of the PGF is roughly proportional to the underlying terrain slope with values increasing from relatively small values over the interior of the continent to largest values over the steep coastal slopes. Magnitudes equivalent to geostrophic winds in excess of 30 m s⁻¹ are found near the coast of East Antarctica.

A second experiment was then conducted to illustrate the potential effects of the terrain on the establishment
of the surface wind field without consideration of the katabatic influence. This experiment is identical to the previous simulation except that (i) no longwave radiative cooling is present, (ii) a PGF exists at the start of the numerical experiment, and (iii) initial wind components are assumed to be in geostrophic balance with the pressure field. The purpose of this simulation is to examine how the Antarctic topography is able to force an adjustment in the wind and pressure fields that is independent of katabatic forcing. By turning off the longwave radiation parameterization, no surface cooling and hence no katabatic acceleration can be produced in this simulation. The initial PGF field was specified following KT and consists of a surface anticyclonic circulation centered about the South Pole that decreases with height and becomes cyclonic in the upper troposphere. Estimates of the initial PGF were obtained from maps shown in KT. An equivalent geostrophic wind of 20 m s⁻¹ (counterclockwise) was set at 1000 hPa, decreasing linearly with height to a 10 m s⁻¹ geostrophic wind (clockwise) at 250 hPa. Wind components were set to geostrophic values to start the simulation. This numerical simulation is thus similar to the rotating tank experiments discussed previously in that an initial circulation is impulsively projected against the Antarctic orography.

Rapid adjustment of the low-level wind field occurs as the stable airstreams are directed toward the Antarctic continent. As stable air is forced against a topographic obstacle, it must rise and hence decelerate since a fraction of the kinetic energy of the flow is converted into potential energy. Convergence adjacent to the terrain results in an increase in pressure along the windward face of the barrier and thus an adjustment of the horizontal pressure field. A PGF becomes established such that the direction of the force is down the local fall line. If such forcing persists over several hours, Coriolis effects become apparent. Flows then take on characteristics of geostrophic winds with directions perpendicular to the PGF and hence parallel to the orography. The process described above follows the development of “barrier” winds (Schwerdtfeger 1975) or mountain-parallel motions associated with “cold air damming” (Bell and Bosart 1998).

Simulated flows become dramatically altered over the Antarctic topography from the initial state almost immediately. Although clear topographic control of the low-level wind field is apparent after 24 h, simulations were carried out for longer time periods to allow ample time for spinup of topographically induced circulations. Similar to that observed in the rotating tank experiments, low-level winds become constrained by the orography and are directed with components that follow the height contours of the ice sheet. Shown in Fig. 3 are the 0- and 72-h wind vectors and streamlines and the 72-h PGF at the lowest sigma level. The adjustment of the surface wind regime over the continent is apparent during this integration. Circumpolar low-level winds (Fig. 3a) become modified by the Antarctic terrain such that by 72 h (Fig. 3b) the direction of the wind vectors becomes tied to the underlying terrain. In this manner the surface winds take on characteristics of a katabatic wind regime. Despite markedly different initial conditions and physical forcing mechanisms between the two simulations, wind vector depictions in Fig. 2a and Fig. 3b display a similar pattern with radially diffuent flow off the elevated East Antarctic plateau. Note that confluence in the wind field is also present in the streamlines.
Fig. 3. Wind vectors and streamlines at 10 m above surface at (a) $t = 0$, (b) $t = 72$ h, and (c) PGF vectors at 72 h for idealized ambient wind simulation without radiative cooling effects. Terrain contours thin; dashed lines at 400-m intervals as in Fig. 1.

shown in Fig. 3b and in similar locations as seen in Fig. 2a. Many authors have discussed the robust nature of the Antarctic surface wind field. This persistence reflects to a large degree the strong topographic constraints placed on the low-level wind field by the Antarctic orography.

Figure 3c illustrates that the resulting 72-h PGF vectors at the lowest sigma level reflect strong topographic adjustment. As in Fig. 2b, PGF magnitudes are expressed in terms of the geostrophic wind. There appears to be some enhancement of the PGF magnitudes near the coast, yet the outstanding feature is the directional changes that have occurred during the adjustment period. Each PGF vector is directed nearly downslope, comparable to the directions shown in Fig. 2b.

The idealized simulations presented are not meant to convey the complex plethora of processes at work in forcing Antarctic surface winds but rather to show that realistic depictions of the wind over the continent can be obtained without need for representation of katabatic forcing. Simulations such as that shown in Fig. 3 attest to the strong topographic control placed on the low-level wind field by the Antarctic ice sheet that must take place independent of the katabatic forcing. It is also interesting that the flows at the top of the model in the second simulation (not shown) display a clear wave-number 3 pattern about the continent with marked troughs in the Ross Sea and Weddell Sea sectors and a less-defined trough along approximately 70°E. This pattern matches well with the rotating tank experiments of
Baines and Fraedrich (1989) and again underscores their contention that the Antarctic orography influences the circulation throughout a deep vertical column that in the real world extends well into the stratosphere.

3. MM5 simulations of the midwinter Antarctic surface wind regime

To properly simulate the actual forcing of Antarctic low-level winds, the full complement of physical parameterizations and initial conditions representative of the real atmosphere are required. In pursuit of this goal, daily numerical simulations were conducted using the AMPS version of MM5 for the entire month of July 2001. This time period was selected since it represents the midwinter period where the influence of katabatic forcing is at a maximum. A detailed description of MM5 can be found in Grell et al. (1994). The Medium-Range Forecast (MRF) scheme (Hong and Pan 1996) is used to simulate fluxes in the planetary boundary layer (PBL); the multilayer soil model was used to represent heat transfer through the surface. Long- and shortwave radiative fluxes were modeled using the Community Climate Model 2 parameterization (Hack et al. 1993). Moist physics routines used include the simple ice scheme to represent cloud and precipitation processes and the Grell parameterization for subgrid-scale cloud processes. The model domain consists of a horizontal grid of 127 × 127 points with a resolution of 60 km. Such a horizontal grid spacing is insufficient to resolve finescale structure of the Antarctic terrain such as near the Transantarctic Mountains but is more than adequate for representing the broad features of the ice sheet. As shown by Cassano (1998) and Cassano and Parish (2000), this grid resolution is also adequate for capturing katabatic winds over the steep coastal slopes. In the vertical, 25 sigma levels were specified with the lowest level approximately 6 m above the surface over the Antarctic terrain. Seven sigma levels were situated within the lowest 250 m to allow high resolution of the near-surface wind and temperature fields. MM5 was initialized using the daily 0000 UTC NCEP–NCAR global reanalysis output set (Kalnay et al. 1996) for July 2001 and each simulation was carried out for a 36-h model integration period. Following previous MM5 simulations of Antarctic circulations (e.g., Guo et al. 2001), a specific time period was set aside to allow the realistic circulations to spin up. For the analyses to follow, an 18-h spinup time was allowed. Daily averages were then computed from the subsequent 6-h forecast periods commencing with the 18-h values. Monthly means were then determined by averaging each 6-h time period. In addition to the standard MM5 output, forcing terms from prognostic equations were added to the output set. This provides details regarding the exact forcing mechanisms at work within the model to force the flows. In this study, emphasis will be placed on the PGF from the horizontal equations of motion at the lowest sigma level.

Detailed validation of Antarctic simulations has been notoriously difficult owing to the lack of observational data over the high southern latitudes. There exist nearly two dozen manned stations that ring the continental periphery. Approximately 70 automatic weather stations (AWSs) have also been deployed at various sites about the continent, mostly near the coast and over adjacent ice shelves. Data over the interior of Antarctica, however, are especially lacking. Currently, South Pole is the only station in the interior of Antarctica for which soundings are routinely made. The presumption made at the start of this study was that the assembled reanalysis grids were representative of the gross features in the atmosphere surrounding Antarctica and subsequent evaluation of MM5 output provided confirmation. For this study, comparison was made between the July 2001 MM5 output with the limited observational dataset from July 2001 and long-term climatological records for available stations in the continental interior. As an example, Fig. 4a depicts the MM5 mean July 2001 temperatures at the lowest sigma level of σ = 0.999 (corresponding to a height of approximately 6 m AGL). Temperatures range from less than 200 K atop the high plateau to approximately 260 K at the East Antarctic coast. Temperatures near the coast are reasonable based on AWS and manned station records and multiyear climatological averages. Surface temperatures over the interior appear to be approximately 5 K colder than observations suggest, although given the limited observational dataset it is difficult to assess if this represents a cold bias in the model or if July 2001 is colder than normal. Corresponding potential temperatures at σ = 0.999 illustrated in Fig. 4b show that the near-surface environment over the high plateau is potentially colder than that near the coast. This implies that as air moves downslope, it will be more dense than the air it replaces and hence katabatic flow is expected. Again, there is a suggestion that the MM5 representation of the high interior may be slightly too cold. AWS records indicate that Dome C (75.12°S, 123.4°E) at a height of 3250 m has a mean July potential temperature of approximately 240 K compared to 232 K from the MM5 simulations.

For the purposes of this study, it should be noted that any cold bias in temperatures over the high interior would result in a stronger katabatic wind regime.

Soundings at South Pole are conducted once per day during the austral winter season. These data are critical for model validation regarding the vertical structure of temperature over the interior of the continent. A comparison of the monthly mean vertical profiles of potential temperature from MM5 and average 0000 UTC sounding data from July 2001 at South Pole is shown in Fig. 5. There appears to be close agreement between model output and the mean potential temperatures, especially in the lowest 3000 m AGL. This provides confirmation that, at least over this section of the interior, MM5 is able to capture the important thermal forcing. Again, the MM5 surface temperatures are slightly too
cold, being approximately 3K colder than those observed.

Low-level wind speed and direction from the MM5 July 2001 simulations also appear reasonable as compared with climatology. Figure 6 illustrates the mean wind vectors and streamlines at the lowest sigma level of $\sigma = 0.999$ from the monthly MM5 simulations. Winds display a strongly diffluent airflow pattern off the high plateau of East Antarctica, similar to that proposed by Parish and Bromwich (1987) and also representative of that shown in Figs. 2b and 3c for the idealized experiments. Confluence zones are apparent along the coastal margin including to the south of the Ross Ice Shelf, Adélie Land, and the Amery Ice Shelf, similar to those described previously. These confluence regions have been singled out as responsible for the anomalous intensity of near-coastal wind regimes in-
including the infamous Cape Denison site (Parish 1981; Parish and Wendler 1991). Wind speeds show a general increase from the high interior of East Antarctica to the coast. Strongest winds, which in places approach 20 m s⁻¹, are found along the coastal rim of the continent. There is an enhancement of the coastal wind regime downwind of confluence zones such as near Cape Denison that matches observations from AWSs. Results illustrated in Fig. 6 provide testimony that the MM5 July 2001 simulations are able to capture the primary observed features of the Antarctic surface wind regime. Additional analyses regarding the performance of MM5 in simulating Antarctic weather conditions can be found in Guo et al. (2003).

To examine details of the forcing of the low-level winter flows over Antarctica, the PGF from MM5 needs to be evaluated. Figure 7 shows the mean PGF at the $\sigma = 0.999$ for the July 2001 MM5 simulations. Vectors lengths again correspond to geostrophic wind magnitudes. The outstanding feature of the PGF vectors over Antarctica is the alignment down the fall line of the local terrain. PGF vectors over the ice sheet are directed downslope at nearly every grid point similar to that seen in the idealized experiments shown previously. Significant differences are seen in the orientation of the PGF vectors over the continent versus those over the ocean to the north of the coastline, indicating an abrupt change in the horizontal pressure field. It is evident that the continental ice sheet forces a pronounced adjustment of the ambient wind and pressure fields in the lower atmosphere. The magnitudes of the PGF show an increase from atop the continental interior of East Antarctica toward the ocean with largest values in excess of 40 m s⁻¹ equivalent geostrophic winds over the steep coastal slopes. Local maxima in the PGF magnitude are found along confluence zones depicted earlier. From Fig. 7 it is apparent that the PGF near the surface of the Antarctic continent is directed down the fall line with a magnitude proportional to the underlying terrain slope. Both characteristics are what would be expected for a purely katabatic wind regime such as that seen in Fig. 2 and as described in other studies (e.g., Parish and Bromwich 1987).

Given the PGF from MM5, it is logical to inquire as to whether this force is the result of katabatic processes or other processes in the ambient atmosphere. To determine the fraction of the net forcing that can be attributed to katabatic influences, a budget study of the PGF is required. As shown by Mahrt (1982) and Whitelaw (1990), the PGF in the lower layer of the atmosphere over sloping terrain can be expressed by

$$\text{PGF} = \frac{g}{\theta_0} \frac{d}{\cos \alpha} \sin \alpha - \cos \alpha \frac{g}{\theta_0} \frac{\partial(h)}{\partial s} - \frac{1}{\rho} \frac{\partial p}{\partial s_{\text{amb}}}, \quad (1)$$

where $d$ is the potential temperature deficit, the difference in potential temperature at any point in the katabatic layer between the radiatively cooled layer and the ambient atmosphere. The term $\alpha$ refers to the terrain slope, $\theta_0$ is the mean potential temperature in the cooled layer, $h$ is the depth of the radiatively cooled lower layer, and $s$ refers to a horizontal distance. The subscript amb in the third rhs term refers to the ambient environment. Other symbols have their usual meteorological meaning. The first rhs term is the katabatic force (hereafter KF), arising due to radiative cooling of the ice slopes. It is dependent on both the amount of radiative cooling and the slope of the underlying surface. This term is the critical term for inferring the fraction of the PGF that can be explained by katabatic processes in the MM5 simulations. The second rhs term represents the effects of varying layer thickness and/or diabatic cooling in the downslope direction and will be known as the “integrated deficit” term. Cassano (1998) referred to this second term as the “adverse” PGF term, although noting that it can act either to suppress or accelerate downslope flows. The third term refers to the PGF in the free atmosphere above the layer of diabatic cooling, which will be referred to as the ambient PGF component. This term represents the effects of horizontal pressure gradients in the atmosphere that are undisturbed by effects of diabatic cooling over the sloping ice terrain. Such a PGF component does, however, reflect influences such as blocking and adjustment.

Both KF and the integrated deficit PGF terms can be computed from the vertical profile of potential temperature at a particular grid point. Calculation of the potential temperature deficit, $d$, is necessary for evaluation of each term. This quantity can be determined at each level for every grid point by extrapolation of the potential temperature profile in the free atmosphere down...
to the surface. The process for calculating the deficit is shown schematically in Fig. 8. It is assumed that the influence of diabatic cooling and hence the thickness of the katabatic layer is limited to approximately the lowest few hundred meters based on sounding data from over the ice sheet. Here the layer from 750 to 2000 m is assumed to represent the ambient atmosphere just above the diabatically cooled layer. A least squares linear fit to the ambient MM5 potential temperature profile at a particular grid point, $\theta(z)$, is made. Extrapolation of this line to the surface provides an estimate of the ambient thermal structure, $\theta_0(z)$. The difference between the MM5 potential temperatures, $\theta(z)$, and the assumed ambient potential temperature, $\theta_0(z)$, at the same level yields $d$. Calculation of the integrated deficit term requires determination of the integral of the product of the potential temperature deficit and the depth of the overlying radiatively cooled layer, and so the vertical profile of $d$ is required.

Figure 9 depicts the mean potential temperature deficit at the lowest sigma level of $\alpha = 0.999$ for July 2001. Not surprisingly, the largest potential temperature deficits are found over the high plateau of East Antarctica with values in excess of 30 K over a broad section of the interior and reaching 40 K over the highest ridges. Deficits decrease toward the coast and are typically less than 10 K within 200 km of the coast. This pattern is to be expected since the stronger winds over the coastal margin act to mix the lower atmosphere more than that seen over the high interior of East Antarctica (e.g., Schwerdtfeger 1984). Such potential temperature deficits are related to the strength of the surface temperature inversion, a common measure over the Antarctic that has been discussed in the early literature (Schwerdtfeger 1970). Inversion strength can be defined as the difference between the surface temperature and the warmest temperature as recorded by a sounding. Phillpot and Zillman (1970) have estimated the strength of the mean winter inversion and found values in excess of 25 K over the highest portions of East Antarctica. It should be noted that potential temperature deficits typically have absolute values significantly larger than quoted values of inversion strengths since the latter use actual temperature rather than potential temperature differences. Schwerdtfeger (1970) has commented on the difficulty of defining a standard height by which inversion strength is measured. Here it is sufficient to note that the inferred potential temperature deficits display a pattern similar to that of the inversion strength as shown in Phillpot and Zillman (1970) with isoliths of each roughly following contours of the ice terrain. It should also be noted that calculation of potential temperature deficit is dependent somewhat on the selection of the height and thickness of the layer that is assumed to represent the ambient environment. Sensitivity calculations have been conducted using a variety of layer thickness and height characteristics and show only minor changes from that shown in Fig. 9. As an example, specifying the ambient layer from 1000 to 3000 m results in computed deficits with maximum differences of less than 10% of the values shown.

To quantify the contribution of the katabatic forcing to the total PGF, each term in (1) must be determined. The first two terms, $KF$ and the integrated deficit, were calculated directly from the MM5 potential temperature profiles at each grid point. It is not possible to directly determine the third term in (1). Since, however, the total PGF is known and the $KF$ and integrated deficit terms in (1) can be calculated, a very good estimate for the third term can be obtained as the difference between the total PGF as shown in Fig. 7 and the first two terms in
Figure 10. As in Fig. 7 except for katabatic component of the horizontal pressure gradient force.

Figure 11. As in Fig. 7 except for ambient component of the horizontal pressure gradient force.

To ensure consistency with model output, finite differences for the first two terms in (1) were computed using the same numerical strategy as in MM5. Figure 10 depicts the mean KF vectors for the July 2001 period. As in Fig. 7, each KF vector length corresponds to a geostrophic wind magnitude. By definition, all KF vectors are directed downslope. Largest magnitudes in Fig. 10 are found near the coastal margin over the steeply sloping ice slopes. The KF magnitudes range from geostrophic equivalent wind magnitudes of less than 5 m s\(^{-1}\) over the high interior of East Antarctica to greater than 30 m s\(^{-1}\) at the coast. Maxima in the KF magnitudes again correspond to regions of streamline confluence over the ice sheet, and the patterns of KF maxima are similar to that shown in Parish and Cassano (2001) based on the NCEP–NCAR reanalysis. A comparison between the vector fields of Fig. 10 with those of Fig. 7 provides an estimate of the fraction of the total forcing of the wind field at \(\sigma = 0.999\) that arises due to katabatic forcing. Close inspection shows that KF composes a significant component of the total PGF over the interior. It is evident, however, that the KF component is only a fraction of the total PGF over large coastal sections of East Antarctica. Vectors of the total PGF from the East Antarctic coast to several hundred kilometers inland have magnitudes that are often at least twice as large as corresponding KF vectors.

As noted above, an estimate of the PGF associated with the background environment, the third term in (1), can be obtained by vector subtraction of the sum of KF and the integrated deficit PGF component from the total PGF. The integrated deficit PGF component was found to be considerably smaller than KF. Calculations revealed that this term was insignificant except at grid points over the terminal coastal slopes and, hence, is not shown. This is in accord with previous work regarding the integrated deficit term (e.g., Ball 1960; Cassano 1998; Parish and Cassano 2001) and was expected. The sum of KF and the adverse PGF terms in (1) then provides an estimate as to the total influence of the diabatic cooling process in forcing the motion at the lowest sigma level. Figure 11 illustrates mean July 2001 vectors of the ambient PGF at \(\sigma = 0.999\). The vectors in Fig. 11 have again been scaled by the Coriolis parameter so as to reflect the geostrophic wind magnitude. Note that vector magnitudes increase away from the elevated East Antarctic plateau and reach a maximum over the steep coastal ice slopes, similar to that for KF in Fig. 10. A second noteworthy feature is the orientation of the ambient PGF down the local fall line at nearly every grid point over the model continent. This again is a feature that is seen in the KF vectors over the continent. That the direction of the ambient PGF component is tied to the slope of the underlying terrain implies that significant adjustment of the nonkatabatic component of the PGF has occurred, most likely through blocking effects of the Antarctic orography and subsequent adjustment as shown in Fig. 3. In this manner, ambient PGF forcing masks itself as katabatic although it is forced by topographic adjustment processes. Inspection of the vector field shows that on average the wintertime ambient PGF exceeds KF along the coast. Of special interest are the magnitudes of the ambient PGF along the East Antarctic coastal margin. This region is among the most katabatic-prone regions of Antarctica and an extensive body of literature can be found...
regarding the surface flows along the coast. Figure 11 illustrates that the ambient PGF vectors at the surface for a region extending inland at least 300 km from the East Antarctic coast and stretching from approximately 70° to 150°E have magnitudes that exceed an equivalent geostrophic wind of 25 m s⁻¹. Such values are considerably larger than corresponding KF magnitudes suggesting that the existence of the mean wintertime low-level winds cannot be explained by katabatic processes alone. It appears that the adjustment processes such as seen in the simple numerical experiment illustrated in Fig. 3 are at least as important in the forcing of these surface winds, even during the midwinter period.

It is obvious that the topographic control suggested in the MM5 simulations and those revealed in the idealized experiments shown previously become more dominant on the low-level Antarctic wind field during summer or transitional months when the role of katabatic forcing must be considerably weaker. It appears that the katabatic forcing represents one component of the wind and at times is swamped by effects of the ambient environment through topographic blocking and adjustment such as shown previously. Antarctic surface winds, thus, represent a more complicated wind regime than has often been discussed and represent more than a katabatic wind environment.

4. Summary

Antarctic surface winds, especially those that are found during the nearly 9-month non-summer period, have often been referred to as “katabatic.” Primarily characteristics of the wind, such as the high directional persistence, mean direction that is clearly related to the local fall line of the terrain, and the mean strength of the flow that is proportional to the underlying slope, are all classic signatures of katabatic processes. Parish and Cassano (2001, 2003) have argued that such an interpretation cannot be used to explain the persistence of the wind regime during summer months or during episodes of strong cyclone forcing. Others such as Neff (1999) have also questioned the supposed dominance of katabatic processes in forcing the surface wind over Antarctica.

Simple numerical experiments have been conducted to show that the continental orography is responsible for significant modification of the horizontal pressure field without consideration of katabatic influences. When stable air is directed such that even a small component of motion is directed against the continental ice sheet, significant adjustment in both the motion and pressure fields result. After adjustment, the winds are directed with a significant component parallel to the barrier and with a PGF directed down the local fall line. Surface winds are thus directed along topographic pathways similar to that expected from katabatic flow. Wind direction alone is therefore a poor indicator of katabatic flows.

Daily numerical simulations for the month of July 2001 have been conducted with the AMPS version of MM5 to depict the net forcing of the time-averaged flows over Antarctica. Model output includes the PGF as represented in the horizontal equations of motion in MM5. Results of the July 2001 simulations show that the mean PGF at the lowest model level is directed down the local fall line at nearly every grid point over the continent. Magnitudes of the PGF increase monotonically from the gently sloping high interior to the steep coastal margin, similar to that expected from a purely katabatic-driven wind regime. Separation of the PGF into katabatic and ambient components revealed that the surface wind regime could not be explained entirely by effects of diabatic cooling over the sloping ice fields. Wintertime surface winds over the near coastal ice slopes appear to be forced by both katabatic and ambient components of the PGF with the latter being of larger magnitude. This supports the work presented in Parish and Cassano (2001) that Antarctic surface winds are more complicated than simply drainage flow off the continental ice sheet. Topographic adjustment is an important forcing mechanism in producing the Antarctic low-level winds even during midwinter when katabatic effects are maximized. Reference to Antarctic flows as simply katabatic is an oversimplification of the boundary layer processes acting to produce such motions.

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REFERENCES


