Dynamics of the Katabatic Wind Confluence Zone near Siple Coast, West Antarctica*

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ABSTRACT

The surface wind pattern over the ice sheets of Antarctica is irregular with marked areas of airflow confluence near the coastal margins. Where cold air from a large interior area of the ice sheet converges (a confluence zone), an anomalously large supply of air is available to feed the coastal katabatic winds, which, as a result, are intensified and more persistent. The confluence zone inland of Siple Coast, West Antarctica, differs from its East Antarctic counterparts in that the terrain slopes become gentler rather than steeper as the coast is approached. In addition, synoptic processes exert substantially more impact on the behavior of the surface winds.

A month-long field program to study the dynamics of the springtime katabatic wind confluence zone has been carried out near Siple Coast. Two sites, Upstream B (83.5°S, 136.1°W) and South Camp (84.5°S, 134.3°W), were established roughly perpendicular to the downslope direction. The field program involved the use of the ground-based remote sensing equipment (sodar and RASS) along with conventional surface and balloon observations. Previous analyses revealed the cross-sectional structure of the confluence zone as consisting of a more buoyant West Antarctic katabatic airflow overlying a less buoyant katabatic airflow originating from East Antarctica.

The force balances inside the confluence zone are here investigated for three situations: mean (all available wind profiles from balloon launches), and two extreme cases (light and strong winds). A linear regression method is used to estimate the mean vertical wind shears and horizontal temperature gradients. The vertical wind shears are used to examine whether or not the airflows are in geostrophic balance. The results are 1) the airflow above the surface at both sites is in geostrophic balance for the three situations; 2) inside the West Antarctic katabatic wind zone, there are three forces in the north–south direction—the restoring pressure gradient force associated with blocking of the katabatic and synoptic winds, the downslope buoyancy force, and the synoptic pressure gradient force associated with the time-averaged low in the South Pacific Ocean; 3) above the West Antarctic katabatic wind layer, the observed easterly wind is due to the synoptic pressure gradient force associated with the low; 4) inside the East Antarctic katabatic wind zone, in addition to the above three forces, there is the downslope buoyancy force associated with the inversion; and 5) large-scale transient synoptic systems strongly influence the downslope wind speed and the boundary layer depth, resulting in the light and strong wind cases.

1. Introduction

Katabatic winds, revealed by numerical studies (Parish and Bromwich 1986, 1987, 1991; Parish et al. 1993a; Gallée and Schayes 1994; Hines et al. 1995; Du et al. 1995), are the most significant climatological feature of the boundary layer over Antarctica. The surface winds over the sloping ice fields of Antarctica are not isolated from their surrounding environment. They form the equatorward branch of a thermally direct circulation that has rising motion in the circumpolar trough that surrounds the continent and tropospheric sinking motion over the interior to supply the equatorward-directed surface winds. Thus, the fluctuations of Antarctic surface winds are linked to those of the tropospheric circumpolar vortex and the westerlies over the Southern Ocean (e.g., James 1989; Parish 1992a; Egger 1992; Yasunari and Kodama 1993), although there is no simple, unambiguous relation for this linkage on synoptic timescales (Simmonds and Law 1995). Furthermore, this continent-wide airflow has been inferred in part to be responsible for the semiannual pressure oscillation in high southern latitudes, the circumpolar easterlies adjacent to the continent, and the circumpolar low pressure trough (Parish et al. 1994). In short, the katabatic winds are becoming recognized as an important component of the meteorology of high southern latitudes.

Detailed examination of the surface wind fields over the sloping ice sheets reveals that they are highly irregular with marked areas of confluence just inland from the coastal margins. The confluence zone near Siple...
Coast (see Fig. 1 for geographic location) has been confirmed by observational analyses (Bromwich 1986; Bromwich and Liu 1996, and numerical simulations (Parish and Bromwich 1986; Du et al. 1995; Bromwich et al. 1994). In comparison to the confluence zones over East Antarctica, the Siple Coast confluence zone offers a very different dynamical setting. First, terrain slopes are steeper in the interior than near the margins, and this leads to convergence of the airflow where terrain slopes become gentler rather than steeper. Second, the comparatively low elevation of West Antarctica allows the effects of synoptic disturbances to penetrate deep into the ice sheet interior. This, in combination with relatively gentle terrain slopes, may mean that the winds in the Siple Coast area are more affected by synoptic forcing than those in East Antarctica. Third, the annual-mean potential temperature along the snow surface increases strongly with elevation (Radok 1973); by contrast, in East Antarctica, the potential temperature decreases with elevation. This marked difference appears to imply very different atmospheric dynamics in the two areas (Radok 1973; Kodama and Wendler 1986; Stearns and Wendler 1988).

The influence of large-scale forcing on the katabatic wind regime has been studied in a number of places along the coastal margins of East Antarctica (e.g., Streten 1963, 1990; Loewe 1974; Murphy and Simmonds 1993). Streten (1963) and Loewe (1974) indicated that strong wind events were associated with the influence of low pressure systems. Streten (1968, 1990) discussed the importance for prolonged periods of strong winds of the synoptic pressure gradient. Murphy and Simmonds (1993) used a general circulation model (GCM) to simulate strong wind events near Casey (66.25°S, 110.9°E). They found that the strong wind events were associated with strong katabatic and strong gradient flow operating together. Parish et al. (1993b) carried out observational and numerical studies of the large-scale influence on the katabatic wind regime at Adélie Land, East Antarctica. They found that strong katabatic wind cases are associated with low pressure over the coastal margin. The model results showed that the large-scale forcing plays a key role in the development and intensity of the katabatic wind regime. By contrast, the influence of large-scale forcing on the katabatic wind regime in West Antarctica is poorly understood. Due primarily to the sparse data network, analyses currently are limited to numerical simulations (e.g., Bromwich et al. 1994; Du et al. 1995). The results from these simulations showed that the imposed large-scale pressure gradient substantially accelerates the katabatic flow and allows it to propagate across the Ross Ice Shelf.

In 1992, the Polar Meteorology Group from the Byrd Polar Research Center deployed a four-member team to the Siple Coast area and conducted a month-long field program to study the katabatic wind confluence zone during spring (Bromwich and Liu 1996). Two sites, Upstream B (83.5°S, 136.1°W, 370-m elevation) and South Camp (84.5°S, 134.3°W, 666-m elevation), were established (see Figs. 1 and 2) roughly along the terrain contours. The field program involved the use of the ground-based remote sensing equipment (sodar and RASS) along with conventional surface and balloon observations. The composite analyses (Fig. 13 in Bromwich and Liu 1996) revealed the cross-sectional structure of the confluence zone. The main features are as follows.

1) A relatively cold katabatic flow from East Antarctica occupies the layer between the surface and roughly 500 m above ground level (AGL). 2) Low-level jets are present below 200 m AGL and are stronger near the Transantarctic Mountains. 3) Weak inversion layer tops are found near 500 m AGL. 4) A warmer West Antarctic drainage flow overlies the colder East Antarctic katabatic flow and has a depth of approximately 1000 m at Upstream B and 1750 m at South Camp due to blocking against the Transantarctic Mountains as a result of the West Antarctic airflow convergence. 5) This warm flow is statically stable with the air just above the inversion top originating near the surface but to the west of Byrd and gradually shifting to the east of Byrd as height increases. Note that the Transantarctic Mountains block only the West Antarctic katabatic component.

This study extends the previous observational study conducted by Bromwich and Liu (1996) by analyzing the cross-sectional force balances. The force balances are first examined for the mean situation. Then, the force
balances are examined for two extreme wind cases (light and strong) in order to understand how the katabatic winds respond to different large-scale influences. The present paper is organized as follows. Section 2 describes the data and methodology, and section 3 presents the results. A summary and discussion are given in section 4.

2. Data and methodology

Upstream B and South Camp (Fig. 2) were established from 11 November to 8 December 1992. They are roughly 110 km apart, and South Camp is approximately 100 km from the foot of the Transantarctic Mountains. The terrain slopes are $4.5 \times 10^{-3}$ toward 260° at Upstream B and $7.6 \times 10^{-3}$ toward 285° at South Camp, based on Drewry (1983). Balloon launches at Upstream B and South Camp monitored the winds in both boundary layer and free atmosphere. The balloons were launched several times (see Fig. 3) each day but were concentrated during the local daytime hours (local time is UTC + 15 h). Because the surface diurnal variations of wind and temperature at the two sites were weak (Bromwich and Liu 1996), the flow can be treated as approximately in a steady state. Therefore, the lack of nighttime observations is not a significant limitation in examining the governing dynamics. Figure 3 presents a summary of the pilot balloon launches at both sites. It can be seen that the number of observations at both sites are roughly equal except those at 0800 and 2000 UTC when the number of observations at South Camp exceeds that at Upstream B.

Based on the surface wind observations, three situations are investigated: mean (all the available wind profiles from the pilot balloon launches shown in Fig. 3 were used), and two extreme cases (light and strong winds). A summary of the pilot balloon launches used for the light and strong wind averages is presented in Tables 1 and 2.

The boundary layer temperature, pressure, and humidity were measured with an Air-Sonde system, developed by Atmospheric Instrumentation Research (AIR) in Boulder, Colorado. Air-Sonde launches were only taken at South Camp to provide details of the confluence zone thermal structure. Figure 3 shows a summary of the Air-Sonde launches. The numbers of profiles for the light and strong wind cases are 3 and 9, respectively.

The balloon-measured wind profiles used in the observational study by Bromwich and Liu (1996) were approximate. To precisely determine the katabatic wind depth that is crucial for the force balance study, the balloon ascent rates at both sites were carefully recalibrated based on radiosonde data at South Camp and sodar data at Upstream B. The new balloon ascent rates were reapplied to all the wind profiles including those for the light and strong wind cases.

Temporal averaging is applied to the balloon-measured wind and temperature profiles. During the field program, fair weather and partly cloudy skies prevailed. There were two events (each lasting less than 48 h) in which synoptic and mesoscale disturbances significantly influenced the atmospheric boundary layer (ABL) winds and temperatures. Bromwich and Liu (1996) reported that the spatial distribution of the average surface potential temperature during the field program resembles the annual analysis given by Radok (1973). Therefore, this approach is capable of filtering out most synoptic and mesoscale influences (such as passing cyclonic disturbances), and has been used in other ABL studies, such as Kodama et al. (1989) and May and Wilczak (1993). For the case studies, temporal averaging is also applied to the measured wind and temperature profiles during the events. Because the measurements for the two cases were not taken simultaneously, direct comparison between the two sites is not intended. This is discussed further below.

In addition, the Australian Bureau of Meteorology (ABM) analyses (Guymer 1986) were used to describe the large-scale setup for the mean, light, and strong winds. Surface observations from automatic weather stations (AWSs) (Keller et al. 1994) and other manned stations, Byrd (80°S, 120°W) and Casertz (82.3°S, 117°W), were used for pressure anomaly analysis. Surface winds at Upstream B and South Camp were continuously monitored by Lambrecht recording anemometers. The use of pressure anomaly analysis, which provides more detailed mesoscale information than ABM...
analyses, has appeared in several previous studies (e.g., Bromwich and Parish 1988; Carrasco and Bromwich 1993, 1995) due to its capability of removing the topographic influence on surface pressure observations. The pressure anomalies are calculated by subtracting the monthly mean from the hourly observations.

To examine whether or not the winds in the above-mentioned three situations are in the geostrophic balance, the following procedure is used. First, according to Kodama et al. (1989), we define the “basic state” of the boundary layer as a layer that is free from the influence of the earth’s surface and has characteristics of the free atmosphere. Therefore, the basic state of the boundary layer can be obtained by an extrapolation of the free atmosphere conditions into the boundary layer. Assuming that the thermal wind relationship holds in the free atmosphere, which will be proved a posteriori, the relationship can be written as (Holton 1992)

\[ \frac{\partial U_s(z)}{\partial z} = -\frac{g}{f} \frac{\partial T}{\partial y} = \Lambda_x, \]

\[ \frac{\partial V_s(z)}{\partial z} = +\frac{g}{f} \frac{\partial T}{\partial x} = \Lambda_y, \]

where \( U_s \) and \( V_s \) are the \( x \) and \( y \) components of the geostrophic wind, \( g \) acceleration due to gravity, \( T(K) \) the temperature, \( f \) the Coriolis parameter (negative in the Southern Hemisphere), and \( \Lambda_x \) and \( \Lambda_y \) the vertical shears of the wind components in the free atmosphere.

The \( x \) (\( y \)) axis increases toward the east (north), which is approximately antiparallel (perpendicular) to the fall line of the slope (see Fig. 2). Assuming constant vertical wind shears and integrating Eqs. (1) and (2), we have

\[ U_s(z) = U_s(z_0) + \Lambda_x(z - z_0) \]

\[ V_s(z) = V_s(z_0) + \Lambda_y(z - z_0), \]

where \( z_0 \) is a reference height. In this study, the average \( \Lambda_x \) and \( \Lambda_y \) in the three situations are estimated by assuming the layer between 800 and 1200 m AGL to be free atmosphere and by applying linear regression (Kodama et al. 1989) to the measured \( U, V \) components in this layer. The choice of this layer is somewhat arbitrary. Although this layer is in the upper part of the katabatic wind layer, it will be seen later that the choice of this layer is not that important because the katabatic wind circulation is the primary cause of the constant shears. Once the average \( \Lambda_x \) and \( \Lambda_y \) are known, the \( U, V \) components in the rest of the atmosphere can be extrapolated by assuming uniform vertical wind shears and using Eqs. (3) and (4). With known \( \Lambda_x \) and \( \Lambda_y \) and Eqs. (1)

\[ \text{TABLE 1. Pilot balloon launches during the light wind case.} \]

<table>
<thead>
<tr>
<th>Upstream B (( n = 6 ))</th>
<th>South Camp (( n = 7 ))</th>
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<tbody>
<tr>
<td>1800 UTC 24 November</td>
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<td>0300 UTC 28 November</td>
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\[ \text{TABLE 2. Pilot balloon launches during the strong wind case.} \]

<table>
<thead>
<tr>
<th>Upstream B (( n = 5 ))</th>
<th>South Camp (( n = 3 ))</th>
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<tbody>
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<td>1800 UTC 28 November</td>
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<tr>
<td>0600 UTC 27 November</td>
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<tr>
<td>2200 UTC 27 November</td>
<td>0000 UTC 29 November</td>
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<td>0000 UTC 28 November</td>
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Fig. 3. (a) Summary of available pilot balloon launches at Upstream B and South Camp. (b) Summary of available Air-Sonde launches at South Camp.
and (2), the average horizontal temperature gradients are also obtained.

As described in the introduction, the Siple Coast katabatic wind confluence zone is very complicated, involving converging airflows from East and West Antarctica. To reduce the level of complexity, we use the phraseology easterly wind below (above) the inversion layer to denote the East (West) Antarctic katabatic wind, except where more detail is required.

3. Results for mean wind situation

a. Thermal wind analysis

Figures 4a, 4b, and 4c, respectively, show the derived horizontal resultant (vector-averaged) wind speed, resultant wind direction, and directional constancy profiles for both sites. From Fig. 4a, it can be seen that the resultant wind speed at South Camp is much stronger than that at Upstream B and the resultant wind speed decreases with height at both sites. At around 1750 m AGL at Upstream B and 2600 m AGL at South Camp, the resultant wind speed starts to increase with height. Figure 4b shows that the resultant wind directions below these heights are uniformly from the east. The resultant wind directions abruptly shift to westerlies above these heights. Wind directional constancies (ratio of the magnitude of the mean wind vector to the mean wind speed) in Fig. 4c, in general, show that high values are found near the surface and low values near the wind direction shear layers. The low values reflect the large variability
of wind direction at those heights due to oscillations in the depth of the easterly winds. Based upon Fig. 4, the depths of the easterly winds are approximately 1750 m at Upstream B and 2600 m at South Camp. Note that these depths are several hundred meters higher (approximately 17% increase in depth) than those given by Bromwich and Liu (1996) because the balloon ascent rates at both sites were recalibrated (see section 2).

Applying the linear regression relationship [Eqs. (3) and (4)] to the $U$, $V$ components in Fig. 4, the average vertical wind shears ($\Lambda_x$, $\Lambda_y$) in the upper part of the katabatic wind layer are obtained. They are 1.9 and 0.6 m s$^{-1}$ (1000 m)$^{-1}$ at Upstream B and 2.6 and 1.0 m s$^{-1}$ (1000 m)$^{-1}$ at South Camp. Both $\Lambda_x$ and $\Lambda_y$ at South Camp are larger than at Upstream B. The total vertical shears [$ = (\Lambda_x^2 + \Lambda_y^2)^{1/2}$] are 2.0 m s$^{-1}$ (1000 m)$^{-1}$ at Upstream B and 2.8 m s$^{-1}$ (1000 m)$^{-1}$ at South Camp. By comparison, the total vertical shears in Adélie Land, East Antarctica, were around 8.2 m s$^{-1}$ (1000 m)$^{-1}$ during a similar period (Kodama et al. 1989), which is much larger than at the West Antarctic sites. According to the thermal wind relationship [Eqs. (1) and (2)] described in section 2, the corresponding horizontal temperature gradients for both sites are shown in Fig. 5. The temperature gradients in the $x$ ($y$) direction, $-\partial T/\partial x$ ($-\partial T/\partial y$), are positive (negative) at both sites. In Fig. 5, it is also seen that the magnitude of the temperature gradient in the $y$ direction greatly exceeds that in the $x$ direction for both sites. This means that the warm air is primarily to the north. The magnitudes of temperature gradient are substantially smaller than those in East Antarctica. The small temperature gradients at Siple Coast are due primarily to its distant (800 km or more) location from the ice-free ocean. By contrast, Adélie Land of East Antarctica is adjacent to the ice-free ocean and the larger horizontal temperature gradient is expected. Numerical simulations have been conducted by Parish (1992a,b) using both idealized and three-dimensional models. In these simulations, the Antarctic continent was represented by the standard orography and by a flat ice sheet representation. The results show that the baroclinicity along the coastal margins is more intense than in the interior, which is consistent with the above calculation.

The calculated $\Lambda_x$, $\Lambda_y$ are applied to the rest of the easterly and the westerly wind layers at both sites and the extrapolated resultant wind speed and direction versus the observed are plotted in Figs. 6 and 7. It can be seen that the calculated resultant wind speeds and directions in the easterly wind layer at both sites are close to those observed. This suggests that geostrophic equilibrium dominates in the easterly wind layer at both sites. This also suggests that the Ekman effect can be neglected except right near the surface because of the small wind directional shears (nearly uniform wind directions) and linear decrease of wind speeds. The small Ekman effect is primarily due to the nearly frictionless snow surface (some sastrugi were observed at South Camp) and the stable easterly wind layer that suppresses turbulence. Above the easterly wind layer, the interface between the easterly wind and the westerly wind is well reproduced, as is the westerly wind layer. This good agreement indicates that the westerly flow is in geostrophic balance. In summary, Figs. 6 and 7 suggest that the balloon-observed winds at both sites are geostrophic winds.

b. Comparison with observations at D-47, East Antarctica

The geostrophic wind situation at both sites is similar to the daytime boundary layer winds at D-47 in Adélie Land, East Antarctica (Kodama et al. 1989), which are also in geostrophic balance. However, the diurnal variation of surface temperature at both sites is much smaller than that at D-47. For example, the diurnal temperature ranges are 1.9°C at Upstream B and 1.4°C at South Camp (Bromwich and Liu 1996). By contrast, the ob-
served diurnal temperature range (observed during a similar period, but in a different year) is around 9°C at D-47 (Kodama et al. 1989). This difference is primarily due to the 16° more northerly latitude of D-47. Due to the small diurnal variation of surface temperature at both West Antarctic sites, the strong locally generated katabatic wind due to the strong cooling during nighttime over the slope at D-47 (Kodama et al. 1989) is not present. Another difference between the two areas is that the West Antarctic surface heating due to insolation is small enough to inhibit the formation of a convective boundary layer, such as that observed at D-47. In fact, a weak inversion layer whose top is around 500 m AGL is maintained throughout the day.

Kodama et al. (1989) explain the daytime geostrophic wind profiles at D-47 as being due to a mesoscale horizontal temperature gradient, which is quite obviously due to the closeness of D-47 to the ocean (around 200 km). However, the locations of Upstream B and South Camp are far from the closest ocean, which is 800 km or more away. It seems that the mesoscale horizontal temperature gradient due to ice slope–ocean contrast cannot be applied here. This difference leads to the search for the explanation of the observed geostrophic wind profiles.

c. Katabatic wind circulation

One could think that the pressure gradient force arising from the continent–ocean contrast is the driving force and the flow is not katabatic. As mentioned earlier, Parish (1992a,b) conducted several numerical experiments. The result shows the strong easterly winds appear along the coastal margin and quickly disappear toward the interior of the ice sheet. Applying this result to the current situation, it seems that the continent–ocean scale contrast is not important because the vast Ross Ice Shelf (more than 800 km in extent) lies between Siple Coast and the ocean (Fig. 1).

Based on the field observations (Bromwich and Liu 1996), it appears that the katabatic winds induced by the weak inversion layer over sloping terrain are the likely contributing factor to the katabatic wind circulation. Numerical simulations (e.g., Egger 1985; James 1986, 1988, 1989) suggest that upper-level westerlies develop in response to the Antarctic katabatic wind circulation. An upper-level vortex above the ice sheet was produced within a timescale of a few days by the secondary circulations forced by the boundary layer processes. This continental-scale meridional circulation (Parish and Bromwich 1991) consists of the downslope-directed drainage flow, strong convergence just offshore with rising motion, a return branch at middle to high levels, and sinking motion over Antarctica. Here, the importance of the Antarctic ice topography needs to be emphasized. Parish et al. (1994) conducted simulations using the NCAR CCM1 to assess the impact of the continental orography on the high southern latitude atmospheric circulations. The results suggest that the Antarctic topography and attendant katabatic wind regime play a key role in the climate of the high southern latitudes. For example, zonally averaged meridional wind components show that the thermally direct mean winter-time circulation is far more pronounced in the run with the standard Antarctic topography than without (represented only by a flat ice sheet); in addition, the sea level easterly wind pattern is much stronger with the standard Antarctic topography.

Recent numerical simulations over West Antarctica were carried out by Du and Bromwich (1993) and Du
et al. (1995) using a three-dimensional mesoscale primitive equation model (Parish and Waight 1987). The model consists of a total of 11 irregularly distributed sigma levels with the highest resolution near the surface. The model uses a graybody longwave radiation scheme developed by Cerni and Parish (1984). A surface energy budget equation based on the force–restore model of Blackadar (1978) is used to depict changes in snow surface temperature. The simulations started from tranquil conditions and the katabatic wind regime was quickly established. Observational data from the 1992–93 field program (Bromwich and Liu 1996) were used for comparison. The general features of the simulated wind and temperature fields near the surface are in good agreement with those observed (Du and Bromwich 1993; Du et al. 1995). These simulations confirm the importance of the slope-induced thermal contrast and the katabatic wind circulation induced by the weak inversion layer.

### d. Easterly wind depth difference

One remarkable feature in the observational study (Bromwich and Liu 1996) is that the easterly wind depth significantly increases from Upstream B to South Camp (Fig. 8a). The cross section of wind speed over Upstream B and South Camp during the winter katabatic wind simulation in West Antarctica (Fig. 8a in Bromwich et al. 1994) shows that the blocking is almost nonexistent and the katabatic flow is confined below 500 m AGL. A force balance analysis performed by Bromwich et al. (1994) demonstrates that the downslope buoyancy force dominates. This occurs because the winter katabatic flow blows more toward the fall line than toward the mountains, resulting in less blocking. In summer, the inversion strength in Siple Coast is markedly

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**FIG. 7.** Same as Fig. 6 but for South Camp.

**FIG. 8.** Schematic illustrations of cross sections of the Siple Coast confluence in (a) the north–south (y) and (b) the east–west (x) directions. Vertical axis is height above sea level.
reduced. The field measurements (Bromwich and Liu 1996) reveal that the inversion strength in Siple Coast is reduced from about 15°C in winter [estimated from Fig. 5 in Schwerdtfeger (1984)] to less than 4°C. Thus, the downslope buoyancy force in the area is significantly reduced in summer. Parish and Bromwich (1986) conducted a simple diagnostic study of the surface wind pattern over West Antarctica. A comparison between the winter and summer runs showed that the drainage flows blow more parallel to the elevation contours in the summer than in the winter. This means that the West Antarctic katabatic flows blow more toward the Transantarctic Mountains in summer than in winter, which explains the observed blocking in the Siple Coast area during summer. It is necessary to point out that the inversion strength in the summer run, performed by Parish and Bromwich (1986), used half the strength of the winter run, whereas a quarter would be more realistic. This suggests that the summer West Antarctic katabatic flows in the interior are more parallel to the elevation contours than the model suggests. Thus, the piling up of the West Antarctic katabatic flow against the mountains is expected to be much more pronounced in summer than in winter.

Additional supporting evidence for the blocking influence comes from a recent numerical simulation (Du et al. 1995) in which the springtime windfield in the Siple Coast confluence zone has been examined with a three-dimensional hydrostatic mesoscale primitive equation model. Results from the numerical simulation were compared with data from this field program and generally good agreement was found. Figure 9a shows the cross section of wind speed along Upstream B and South Camp based on the model simulation. It can be seen that strong winds are found at low levels near the Transantarctic Mountains because of the blocking influence, and that the wind speed and the depth of the boundary layer decrease away from the mountains, which are as described previously from the observations.

From Fig. 8a, it is seen that the easterly wind depth increase is not only associated with the airflow being blocked against the Transantarctic Mountains because the top of the easterly wind at South Camp is well above the Transantarctic Mountains, which are approximately 100 km to the south. In fact, during the winter simulation (Bromwich et al. 1994), when the initially specified pressure field, which is parallel to the elevation contours, is applied, the blocking appears (Fig. 8b in Bromwich et al. 1994). This is also confirmed by the strong wind case simulated for summer (Fig. 9b).

Based on the numerical simulations, it appears that the easterly wind depth difference between the two sites is partially due to blocking associated with the katabatic and synoptic winds, which are contour parallel and blow toward the Transantarctic Mountains. To investigate the synoptic influence, the ABM digital data were averaged twice per day during the field program. Figure 10 shows
Fig. 10. Australian Bureau of Meteorology synoptic-scale analyses for mean situation. The contour intervals are 2.5 hPa for (a); 30 geopotential meters for (b), (c), and (d); and 60 geopotential meters for (e).
the analyses ranging from sea level up to 300 hPa. From the sea level pressure analysis (Fig. 10a), it is seen that a time-averaged synoptic low is located over the South Pacific Ocean, north of Marie Byrd Land (Fig. 1). This low extends from sea level up to 500 hPa. At 300 hPa, the low becomes a trough. Below 700 hPa, it is seen that Siple Coast is within the circulation of this low and wind direction is primarily easterly. At 500 hPa, the pressure gradient in Siple Coast is weak, indicating a transition zone where the wind direction switches from easterly to westerly. Comparing this height with that in Fig. 8a, it is seen that the ABM analysis greatly overestimates the transition height. At 300 hPa, the wind direction is westerly, a part of the circumpolar vortex circulation.

The contribution to the blocking from the katabatic and synoptic winds (which are contour parallel) can only partially explain the depth difference between the two sites. As indicated earlier, the depth of the easterly wind at South Camp is well beyond the height of the Transantarctic Mountains. According to the ABM analysis, it appears that the structure of the time-averaged synoptic low also contributes to this difference. This is illustrated by the idealized sketch given in Fig. 11. The tilted constant pressure surface $P_L$ in Fig. 11 is due to the inversion layer overlying sloping terrain, blocking of the katabatic and synoptic winds, and the synoptic pressure gradient (the low is located in the South Pacific Ocean). Above the height of the obstacle (the Transantarctic Mountains), the blocking effect disappears and the synoptic pressure gradient (the upper-level pressure surface $P_U$) remains. Because the slope of $P_L$ is steeper than $P_U$, the easterly geostrophic wind speed (the circle) on pressure surface $P_L$ is stronger than that on $P_U$.

e. Force balance analysis

Because $V$ is much smaller than $U$ (Fig. 4), emphasis is on the $U$ component. Based on the above analysis, we divide the easterly layer into three sublayers: the easterly wind above the blocking height, the easterly wind between the blocking height and the top of inversion layer (approximately 500 m AGL), and the easterly wind in the inversion layer.

In the easterly wind layer above the blocking height, it is relatively clear that only the synoptic pressure gradient ($F_y^s$) associated with the time-averaged low in the South Pacific Ocean is present. However, the influence of the West Antarctic terrain on this low is unknown. In other words, whether or not the West Antarctic terrain has an influence on the slope of $P_U$ in Fig. 11 is not clear.

In the easterly wind layer between the blocking height and the top of the inversion layer, there are three forces in the $y$ (north–south) direction: the restoring pressure gradient force ($F_y^t$) associated with blocking of the katabatic and synoptic winds, the downslope buoyancy force ($F_y^b$) due to the cool katabatic air overlying the sloping inversion layer (Fig. 8a), and the synoptic pressure gradient force ($F_y^s$) associated with the time-averaged low over the South Pacific Ocean. These forces contribute to the $U$ component of the easterly winds. In the $x$ (west–east) direction, only the downslope buoyancy force (which contributes to the $V$ component) is important assuming there is no depth change in the $x$ direction (Fig. 8b). In summary, the $U$, $V$ components of the easterly katabatic wind above the inversion layer are given below based on Ball (1960):

$$U_{y_1} = \int \frac{1}{f} (F_{y_1} + F_{y_2} + F_{y_3}) = -\frac{g|\delta\theta|_y}{f\theta} \left( \frac{\partial h}{\partial y} + \alpha_z \right) + \frac{1}{f} F_{y_3},$$

(5)

$$V_{y_1} = \int \frac{1}{f} F_{y_1} = \frac{g|\delta\theta|_y}{f\theta} \alpha_y,$$

(6)

where $U_{y_1}, V_{y_1}$ are the $U_y, V_y$ components of the easterly
katabatic wind; \( \theta \) the mean potential temperature of the layer; \( [\delta \theta]_I \) and \( [\delta \theta]_S \) the potential temperature contrasts in the \( x, y \) directions of the easterly katabatic flow with the environment (in the \( x \) direction: the air to the west outside the katabatic wind zone; in the \( y \) direction: the air to the north outside the blocking zone); \( h \) the depth of the easterly katabatic flow (measured from the snow surface); and \( \alpha_x, \alpha_y \) the terrain slopes in the \( x, y \) directions. Here, \( [\delta \theta]_I \) and \( [\delta \theta]_S \) are functions of height and they cannot be estimated in this study because the maximum height of the radiosonde data at South Camp (Fig. 12) is around 2500 AGL, a level still inside the easterly wind layer (see Fig. 8a). This situation also occurs with the estimation of \( \delta h/\delta y \) in Eq. (5). The top of the easterly wind layer in Fig. 8a is above the Transantarctic Mountains and associated with the synoptic pressure gradient (\( F_y \)). Thus, it probably contains mixed information about the blocking and the impact of sloping terrain upon the southward penetration of the synoptic low. In conclusion, the lack of data limits the demonstration of the geostrophy of the easterly winds.

The force balances inside the easterly winds of the inversion layer are slightly more complicated because they are involved with the blocked katabatic flow aloft and downslope buoyancy forces (\( F_y \) in the \( x \) direction and \( F_x \) in the \( y \) direction) due to the inversion layer over the sloping terrain (Fig. 8). Considering all these factors, the \( U_x \) and \( V_y \) can be written as follows:

\[
U_x = \frac{1}{f} (F_{x1} + F_{x2} + F_{y3}) = U_{x1} - \frac{g[\delta \theta]_{\text{inv}}}{f \theta} \alpha_y
\]

\[
V_y = \frac{1}{f} (F_{x1} + F_{x2}) = V_{x1} + \frac{g[\delta \theta]_{\text{inv}}}{f \theta} \alpha_y,
\]

where \( U_{x1}, V_{x1} \) are the \( U_x, V_y \) components of the easterly katabatic flow below the inversion layer, \( \theta \) the mean potential temperature of the layer, and \( [\delta \theta]_{\text{inv}} \) the potential temperature difference (calculated based on the inversion layer) due to the inversion layer over the sloping terrain (Fig. 8). The detailed calculating scheme for \( [\delta \theta]_{\text{inv}} \) can be found in Kodama et al. (1989). Equation (7) can be evaluated at both sites using the field observations. To simplify this, only the \( U_{x1} \) is averaged throughout the inversion layer using Eq. (7) and this average \( U_{x2} \) is compared with the observations. Because the katabatic wind in this inversion layer originates from East Antarctica and is not blocked by the Transantarctic Mountains (Bromwich and Liu 1996), the first two terms (\( F_{x1}, F_{x2} \)) do not change in the inversion layer. Assuming the synoptic pressure gradient force \( F_x \) is independent of height, these three forces (\( F_{x1}, F_{x2}, F_{y3} \)) can be estimated by using the balloon-measured wind speed (\( U_{x1} \)) at 500 m AGL (the top of the inversion layer). Adding this wind speed to the layer-averaged wind speed associated with \( F_{y3} \), the average \( U_{x2} \) wind speed of the inversion layer is obtained. At South Camp (Upstream B), \( U_{x1} \) is around \(-4.9\) (\(-2.2\)) m s\(^{-1}\), estimated from Fig. 4. The second term on the right-hand side of Eq. (7) is approximately \(-1.7\) (\(-0.8\)) m s\(^{-1}\) for the average conditions: \( \theta = 270\) (266) K, \( \delta \theta = 2.0\) (3.6), and \( \alpha = -3.35 \times 10^{-3} \) (\(-8.3 \times 10^{-4}\)). The calculated \( U_{x2} \) is \(-6.6\) (\(-3.0\)) m s\(^{-1}\). At Upstream B, these average temperature conditions are obtained from RASS measurements, as described in detail by Bromwich and Liu (1996). The average \( U \) component over the same depth, estimated from Fig. 4, is about \(-6.5\) (\(-2.6\)) m s\(^{-1}\). Thus, the calculated \( U \) components at both sites based on Eq. (7) are close to the measured components. This evaluation demonstrates the geostrophy in the katabatic wind zone below the inversion.

4. Results for extreme cases

a. Light winds

As mentioned in the introduction, the Siple Coast confluence zone offers a very different dynamical setting in comparison to its counterparts in East Antarctica. The terrain slopes, which are steeper in the interior than near the margins, allow the effects of transient synoptic disturbances to penetrate deep into the ice sheet interior, implying that the winds in the Siple Coast area are more affected by synoptic forcing than those in East Antarctica.

Starting from 0000 UTC 24 November 1992 (Fig. 13), there was a light wind period at Upstream B and South Camp. The light winds lasted around 60 h at Upstream B and 48 h at South Camp. From Fig. 13, it can be seen that wind speed at Upstream B was much below average (3.0 m s\(^{-1}\)). A similar situation (Fig. 13) was also found at South Camp (average 8.1 m s\(^{-1}\)). During this period, there were balloon wind profiles available from 24 and 25 November (listed in Table 1). Figures 14a, 14b, and 14c, respectively, show the horizontal resultant wind speed, resultant wind direction, and wind directional constant profiles at Upstream B and South Camp based on Table 1. Remarkable changes in the wind profiles are found at both sites in comparison with the mean wind situation. At Upstream B, the wind speed is the reverse of its mean pattern which decreases with height. Instead, the wind speed increases with height (Fig. 14a). The uniform easterly wind pattern found in the mean situation has disappeared. The wind direction rotates clockwise with height (Fig. 14b). The wind directional constancy profile (Fig. 14c) shows relatively high values, which indicates that the variations of the wind direction profiles measured during the light wind period are small. At South Camp, a similar situation is found. The clockwise rotation of wind direction found at both sites signals cold-air advection into the area. As a result, the entire layer below 2500 m AGL is cooler by several degrees in comparison to the mean profile (Fig. 12). It is necessary to point out that direct
wind profile comparisons between the two sites are not intended. This is because there are biases in the time averages: wind profiles at South Camp started and ended earlier than those at Upstream B (see Table 1).

The linear regression analysis [Eqs. (3) and (4)] was applied to the wind profiles in Fig. 14 and yielded the mean vertical wind shears \((\Lambda_x, \Lambda_y)\). These mean vertical wind shears were applied to the rest of the atmosphere. The calculated wind speed and direction profiles versus the observed are plotted in Figs. 15 and 16 and show that the calculated profiles match well with the observed. This means the observed wind profiles are in geostrophic balance. Due to the absence of accurate large-scale pressure gradient information and the incomplete observations, it is difficult to estimate the individual terms in Eqs. (5)–(8) as was done for the mean situation. Based on the similarity found in the mean situation, the above geostrophic wind balance is inferred.

Based on Eqs. (1) and (2), the horizontal temperature gradients for both sites can be derived from the mean vertical wind shears and they are plotted in Fig. 5. It can be seen that the signs of the temperature gradients in both directions remain the same as the mean wind situation. The temperature gradient in the \(x\) direction \((-\partial T/\partial x)\) markedly (slightly) increases at Upstream B (South Camp) in comparison to the mean; by contrast, the decrease in the \(y\) direction is relatively small. The increase of the temperature gradient in the \(x\) direction means an increase of vertical wind shear in the \(V\) component, based on Eq. (2). This increase allows the \(V\) component to more quickly change from northerly to southerly and the changeover height of the \(V\) component is lower than for the mean situation.

To better understand the circumstances when the light wind case occurs, the ABM analysis at 0000 UTC 25 November is presented in Fig. 17. From the sea level pressure analysis (anomalous over the interior of East Antarctica), it is seen that the isobar pattern in the Siple Coast area (Fig. 17a) is similar to the mean situation (Fig. 10a). However, the center of the synoptic low over the South Pacific Ocean has moved to the Ronne Ice Shelf. As a result, the surface geostrophic winds over West Antarctica have shifted from easterly to southeast, except in the Siple Coast area. The expanse of this synoptic low gradually increases as altitude increases (Figs. 17b–e). At 850 hPa, the wind direction in the Siple Coast shifts from easterly to southeast in response to this expansion. This shift continues, and at 300 hPa the wind direction has changed to southwest. In comparison with the balloon wind direction profiles (Fig. 14b), it is found that the ABM analysis is consistent with the observations. From a dynamical viewpoint, the synoptic setup allows the air to flow away from the mountains and reduces the blocking influence and easterly wind depth.

b. Strong winds

From Fig. 13 it can be seen that a strong wind event started right after the light wind period. The surface winds at both sites greatly exceeded the means (3.0 m s\(^{-1}\) at Upstream B and 8.1 m s\(^{-1}\) at South Camp). The ABM analysis at 0000 UTC 28 November is presented in Fig. 18. Comparing Fig. 18 with the average situation,
it is seen that the center of the synoptic low is still over the South Pacific Ocean, but is intensified. For sea level pressure, the innermost isobar value of the low in the average (strong wind) situation is 977.5 (970) hPa, a difference of 7.5 hPa. At 500 hPa, the difference between the innermost isohypses is 60 geopotential meters. The surface pressure anomaly analysis in Fig. 19 shows the more detailed influence of the low, with a trough extending from northern Marie Byrd Land to the Siple Coast area. In addition to Upstream B and South Camp, strong winds also appeared at Casertz, AWS 08, and AWS 11. The strong winds defined here are those that are at least twice as strong as the monthly average. To better illustrate the influence of the trough associated with the low in the South Pacific Ocean, the surface wind speed averaged from Upstream B and South Camp observations versus the surface pressure anomaly difference (South Camp minus Upstream B) between the two sites is plotted in Fig. 20. From Fig. 20, it can be seen that the wind speed increases/decreases as the pressure anomaly difference increases/decreases. Therefore, the strong wind appearance was linked to the imposed synoptic pressure gradient.

The available pilot balloon wind profiles from both sites are listed in Table 2. Comparing Table 2 with Fig. 13, it can be seen that the balloon launches at Upstream B were during the strong wind period, while the launches at South Camp were near the end of the event. The horizontal resultant wind speed, resultant wind direction, and wind directional constancy profiles during this
period are shown in Fig. 21. The wind speeds at both sites increased (Fig. 21a). The resultant wind directions remained rather uniform below the heights dividing the two flow directions (Fig. 21b). In Fig. 21b, the depths of the easterly boundary layer winds are greater than the averages (Fig. 4b), around 2200 m at Upstream B, and around 2900 m at South Camp. The depth increases were around 300 m at each site. The wind directional constancies at both sites (Fig. 21c) have large values beneath the shear layers, which means that the variations between the wind direction profiles were small. Direct wind profile comparisons between the two sites are not suggested, however, because of the time differences.

The linear regression analysis [Eqs. (3) and (4)] for the light wind case was also applied to the wind profiles in Fig. 21. Figures 22 and 23 show both extrapolated and measured wind profiles at the two sites. The surface vector average wind speeds (from automatic recording anemometers) associated with the wind profiles are 8.1 m s$^{-1}$ at Upstream B and 8.2 m s$^{-1}$ at South Camp. Comparing the wind speeds with the calculated profiles in Figs. 22a and 23a, it can be seen that the surface wind speeds match well with those from the calculated profiles. From Fig. 22, it can be seen that the matching of the two profiles at Upstream B is in good agreement. At South Camp (Fig. 23), the corresponding matching

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**Fig. 15.** Same as Fig. 6 but for light wind situation.

**Fig. 16.** Same as Fig. 7 but for light wind situation.
FIG. 17. Same as Fig. 10 but for 0000 UTC 25 November 1992.
Fig. 18. Same as Fig. 10 but for 0000 UTC 28 November 1992.
of the two profiles is generally good below 3000 m AGL. Above 3000 m AGL, the calculated wind speed and direction profiles overestimate the easterly wind depth by around 500 m. This relates to the sudden increase of vertical wind shear in the measured wind profiles at the interface between the easterlies and the westerlies, which is probably associated with the spatial structure of the trough. With this increase of vertical wind shear, the uniform vertical wind shear assumption fails and results in the above mismatch. Like the light case, it is difficult to evaluate the terms in Eqs. (5) and (7) without knowing the large-scale pressure gradients in detail. However, based on the good agreement between the calculated and measured wind profiles in Figs. 22 and 23, it is concluded (similar to the light wind case) that the winds at both sites were in the geostrophic balance.

The calculated average vertical shears from the linear regression relationship can be converted to the horizontal temperature gradients using Eqs. (1) and (2). Figure 5 shows the temperature gradients in the $x$, $y$ directions. It can be seen that the temperature gradient in the $y$ direction greatly exceeds the mean (almost double) at Upstream B. This implies that the baroclinicity associated with the low in the South Pacific Ocean, superimposed on the existing temperature gradient, resulted in the larger vertical shears in the $U$ component. From Fig. 21, the imposed large-scale pressure gradient accelerated the flow. At the same time, the easterly boundary layer wind depths at both sites increased. Therefore, it can be concluded that the depth increases of the easterly winds were primarily due to the increased easterly wind speeds associated with the imposed large-scale pressure gradient. The strengthened easterly winds acted to increase the height of the directional shear layer despite the increased wind shears.

5. Conclusions

The current study examines the cross-sectional structure of the Siple Coast katabatic confluence zone for the mean situation and for the extremes of light and strong
winds. The study shows that there is a strong cross-sectional change of the katabatic wind and its depth, which means that the confluence zone is highly variable along the terrain contours. By contrast, the East Antarctic counterparts are usually depicted as uniform along the terrain contours even though three-dimensional numerical simulations (e.g., Waight 1987) indicate otherwise.

The results show that the flows at Upstream B and South Camp were in the geostrophic balance for the three situations. In the mean situation, the time-averaged synoptic low located in the South Pacific Ocean further complicates the whole picture of the confluence zone. By adding this time-averaged low impact to the easterly “West Antarctic katabatic wind” component, we have slightly modified the picture presented by Bromwich and Liu (1996). With this modified picture (Fig. 24), major results of this study are summarized as follows.

In the West Antarctic katabatic wind layer (which is above the inversion layer), there are three forces in the $y$ (north–south) direction: the restoring pressure gradient force associated with blocking of the katabatic and synoptic winds, the downslope buoyancy force, and the synoptic pressure gradient force associated with the low in the South Pacific Ocean. Above the West Antarctic katabatic wind layer, the observed easterly wind is due to the synoptic pressure gradient force associated with the low. In the inversion layer where the katabatic wind originates from East Antarctica, in addition to the above three forces, there is the downslope buoyancy force as-
CALCULATED RESULTANT WIND PROFILES VS OBSERVED AT UPSTREAM B (STRONG WIND)

![Graph](image1)

Fig. 22. Same as Fig. 6 but for strong wind situation.

sociated with the inversion. Due to the lack of large-scale data, the demonstration of geostrophy is limited to the inversion layer. The easterly wind depth change along the cross section of the confluence zone suggests the influences of sloping terrain, the blocking of katabatic and synoptic winds, and the synoptic pressure gradient. These influences are mixed in the observations from Siple Coast. Without the aid of detailed numerical simulations, it is difficult to evaluate the separate impact of each influence; further work is needed on this topic. In addition, how the sloping terrain of West Antarctica affects synoptic lows centered over the South Pacific Ocean (e.g., changes their vertical structure over the ice sheet) remains unknown. The influence of large-scale...

CALCULATED RESULTANT WIND PROFILES VS OBSERVED AT SOUTH CAMP (STRONG WIND)

![Graph](image2)

Fig. 23. Same as Fig. 7 but for strong wind situation.
The warm West Antarctic katabatic and synoptic flows overlie the cooler katabatic flow from East Antarctica. Hatched area shows the baroclinic zone where these flows meet. Filled circles are manned stations. At bottom, the north–south cross section from Upstream B to South Camp shows the East Antarctic katabatic wind occupying the inversion layer, the West Antarctic katabatic and synoptic winds (between the tops of inversion and blocking layers) being blocked by the Transantarctic Mountains, the synoptic wind between the top of the blocked layer and the westerlies, and the westerlies. Vertical axis is height above sea level.

transient systems on the katabatic winds in the Siple Coast area has been demonstrated by examining two extremes. Both cases show that the influence can substantially modify the spring wind field in the area.

The temperature gradients derived from the wind profiles are in the same direction for the three situations. This explains the high directional constancies of the surface wind observed at both sites (Bromwich and Liu 1996). From numerical simulations and observations, the study suggests that the katabatic wind circulation dominates the winds in Siple Coast. It needs to be emphasized here that our understanding of the complicated Siple Coast confluence zone is limited and further field work (e.g., aircraft observations with combined ground measurements) and numerical simulations are desirable.

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