Numerical Simulation of Winter Katabatic Winds from West Antarctica Crossing Siple Coast and the Ross Ice Shelf*

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ABSTRACT

Twenty-four-hour numerical simulations of wintertime surface winds under clear sky conditions over the West Antarctic ice sheet and its vicinity are performed using a hydrostatic, three-dimensional primitive equation model. Two initial states are examined: a state of rest, and a prescribed pressure field associated with katabatic winds from West Antarctica propagating across the Ross Ice Shelf. The Antarctic katabatic winds are mainly due to the strong radiative cooling of the ice slopes. The West Antarctic terrain is different from that of East Antarctica in two respects: its mean elevation is much lower, and the slope in the interior is steeper than near the margin at Siple Coast.

The simulated surface wind regime reveals confluence zones just inland from the coast and diffuence zones around the crest of the terrain. The model results suggest that the continuation of katabatic winds beyond coastal confluence zones, which are sustained by cold-air drainage in the interior, has an important impact on airflow over the flat Ross Ice Shelf adjacent to the Transantarctic Mountains. The prescribed pressure disturbance has little impact on the surface winds in the interior but markedly impacts those over and beyond the gently sloping coastal areas. Discussion of the impact of the surface wind on the polynya northwest of the Ross Ice Shelf is also provided. It is shown that the simulated surface wind regime is consistent with the available, mostly surface observational data.

1. Introduction

Katabatic winds are a common climatic feature of the lower atmosphere over Antarctica, especially at the coastal margins where strong and persistent katabatic outflows are often observed. The driving force of this unique wind regime is the strong radiative cooling of the ice slopes. Complexity of the ice topography is another crucial factor in shaping the surface wind field. Ball (1960) pointed out that the magnitude of the terrain-induced pressure gradient force is directly proportional to both the steepness of the terrain and the strength of the temperature inversion in the lower atmosphere. Other forces that determine the surface wind field include the Coriolis force and friction. Schwerdtfeger (1970) indicated the dominant influence of Antarctic topography on the surface wind field. A number of numerical studies of Antarctic time-averaged surface winds have been conducted using an accurate topographic map of the continent. Parish and Bromwich (1987) using the simple steady-state model of Ball (1960) diagnosed the wintertime surface streamlines over Antarctica, which showed a highly irregular wind pattern. The winds converge toward several regions near the coast. This provides an enhanced supply of negatively buoyant air to the coastal slopes, allowing the downslope katabatic winds to be stronger and more persistent. These “confluence zones” are thought to dominate the surface wind regime over Antarctica. Continental-scale simulation of Antarctic katabatic winds (Parish and Bromwich 1991) demonstrated their potential impact on the large-scale Southern Hemisphere tropospheric circulation.

Thermal infrared (TIR) satellite images effectively monitor the behavior of katabatic winds over the Ross Ice Shelf under clear sky conditions. Dark (warm) signatures on images of the shelf are often observed coming from the glaciers along the Transantarctic Mountains, such as Reeves (Kurtz and Bromwich 1983) and Byrd Glaciers (D’Aguantte 1986), and from West Antarctica. Synoptic analyses show that the latter airflows blow along the regional isobars, roughly parallel to the Transantarctic Mountains (Bromwich et al. 1992). An-
other prominent feature on wintertime TIR satellite images is the polynya (area of open water/thin ice surrounded by sea/land ice) along the northwestern edge of the Ross Ice Shelf (Zwally et al. 1983; Bromwich et al. 1992), which is apparently enhanced as this katabatic air mass blows offshore.

The present work uses primitive equation numerical modeling to explore the behavior of the most prominent confluence zone over West Antarctica and its impacts beyond the end of the terrain slope over the Ross Ice Shelf. The different dynamical setting of the area (outlined herein) contrasts with that characterizing a similar feature (Terra Nova Bay) in East Antarctica, which has been studied extensively (e.g., Parish and Bromwich 1989; Bromwich et al. 1990).

2. Antarctic topography and the model

The sloped topography is one of the key factors determining the surface-wind regime over Antarctica. Figure 1 shows the Antarctic topography and the model domain in relation to the entire continent. The Antarctic is a huge dome with less than 3% of its 1.4 \times 10^7 km$^2$ being free of snow or ice for part of the year (Schwerdtfeger 1984). The average elevation of Antarctica is greater than 2300 m, substantially higher than Asia, the second highest continent at 800 m. The Antarctic ice topography is generally characterized by quite gentle terrain slopes in the interior and much steeper slopes near the coastal margin. By contrast the slopes over West Antarctica that reach Siple Coast are steeper in the interior than at the coast. The flat Ross Ice Shelf with an area of about 7 \times 10^5 km$^2$ (Schwerdtfeger 1984) smoothly connects the Siple Coast part of West Antarctica to the Transantarctic Mountains of East Antarctica. Ross Island is located at the northwest corner of the Ross Ice Shelf. Many meteorologically important topographic features, such as Byrd and Reeves Glaciers, are found along the Transantarctic Mountains.

The Antarctic terrain heights employed in the model were obtained from an accurate topographic map at a spatial resolution of $d\alpha = 20$ km (Drewry 1983). Figure 2 presents an isometric view of the Antarctic terrain used in the model along the 180° meridian. As the computational domain intersects the very steep Transantarctic Mountains, a nine-point terrain filter was employed to ensure numerical stability. The model domain consists of 160 \times 160 rectangular array of grid points covering the areas of West Antarctica and the Ross Ice Shelf and their vicinity. Ten terrain following sigma levels ($\sigma = 0.996, 0.985, 0.97, 0.95, 0.93, 0.90, 0.80, 0.60, 0.35, 0.10$) are used in the model where the modified vertical coordinate is defined as

$$\sigma = \frac{p - p_s}{p - p_t},$$

where $p$ stands for the pressure of the free surface at the top of the model, which is set to 25 kPa, and $p_t$ is the surface pressure. The first sigma level corresponds to a height of approximately 20 m above the surface; the remaining vertical levels are distributed such that the highest resolution is found in the lower portion of the atmosphere. This enables the boundary layer to be reasonably well resolved; for example, the sigma level 1 winds used here are close in magnitude and direction to those that would be simulated for the standard height of 10 m.

The present model is adapted from Anthes and Warner (1978) and is described in detail by Parish and Waight (1987). It is a hydrostatic, three-dimensional primitive equation system, in which the prognostic equations include the equations of horizontal motion, energy conservation, and continuity. Longwave radiative cooling is treated explicitly in the model following Cerni and Parish (1984). The turbulent flux of heat in the surface boundary layer is modeled using the similarity profiles suggested in Businger et al. (1971), while the surface friction constant is set to $1.67 \times 10^{-5}$ m$^{-1}$ for the Antarctic ice slopes (from Ball 1960) and to $0.42 \times 10^{-5}$ m$^{-1}$ beyond the foot of the ice slope over the smooth Ross Ice Shelf. Fluxes of heat and momentum in the remainder of the planetary boundary layer are computed based on the work of Brot and Wynaard (1978). No allowance is made for the effects of shortwave radiation and clouds, and thus, the simulations represent Antarctic wintertime surface flow under clear sky conditions. The ground temperature $T_0$ is predicted from the force–restore approach of Blackadar (1978). The surface energy budget is simulated by

$$\frac{\partial T_s}{\partial t} = - \frac{1}{c_b} (R_o - R_e - H_o) - K_c (T_s - T_m),$$

where $R_o$ is the downward longwave radiation flux from the atmosphere, $R_e$ is the upward radiation flux from the ice surface, and $H_o$ is the turbulent heat flux to the surface boundary layer. In our study, the heat capacity for packed snow $C_s$ is set equal to $5.74 \times 10^4$ J m$^{-2}$ K$^{-1}$, and the relaxation coefficient $K_c$ is set to $1.82 \times 10^{-5}$ s$^{-1}$, corresponding to heat addition from synoptic forcing at 5-day intervals. Model experiments (Parish and Waight 1987) show that, for Antarctic slope flows, the time evolution of the temperature at the ground is not particularly sensitive to the exact value of $K_c$. The mean temperature of the lowest layer $T_m$ is estimated from climatology.

Two initial states are examined. One is where there are no large-scale pressure gradients, and the other prescribes the broadscale pressure field associated with katabatic surges from West Antarctica across the Ross Ice Shelf (Bromwich et al. 1992). A pseudoanalytic initialization technique is used in which temperature is related to pressure by means of a power law, but this yields an atmospheric baroclinicity for the second run that is somewhat weaker than in Bromwich et al.
Fig. 1. The Antarctic continent with selected geographic features and scientific stations. The larger square is the model domain, and the smaller square is the area shown in Fig. 7. Height contours are in meters (after Parish and Bromwich 1989).

(1992). Model integrations showed that the low-level winds developed rapidly, and by 24 h the katabatic winds had become well established and reached approximately steady conditions. The following section describes 1) the West Antarctic surface wind characteristics simulated by the above experiments, 2) their impact on airflow over the Ross Ice Shelf, and 3) their modulation of the polynya along the northwestern edge of the shelf.

3. The model results

a. Airflow over West Antarctica

In the first experiment the model simulation started from a state of rest. Figure 3a presents the surface streamlines at $\sigma$ level 1 after the 24-h integration over the model domain for this experiment. The gravity-driven slope flows are directed from the interior to the coastal areas at angles generally some 40°–60° from the fall line. The surface airflow converges into several zones around the margin of the ice sheet, such as the Siple Coast area and Byrd Glacier, while diffuence zones are found at the crest of the terrain. This is consistent with the continental-scale streamline simulation for Antarctica (Parish and Bromwich 1987) shown in Fig. 4. The surface wind field over the continent is characterized by areas of confluence just inland from the coast. The confluence zones are linked to intense and persistent katabatic winds such as at Cape Denison, which has yearly averaged winds of 19 m s$^{-1}$. The confluent airflow tends to enhance the supply of the cold negatively buoyant air available to the downslope coastal slopes. This emphasizes the importance of the simulated irregular drainage pattern in understanding
the katabatic wind regime over Antarctica (Parish and Bromwich 1987).

Surface wind observations over West Antarctica are understandably sparse and discontinuous. All available summer field meteorological observations collected during the Siple Coast Glaciology Project were analyzed to check the model simulation. These included data collected by Naval Support Force Antarctica meteorologists and by parties from the Byrd Polar Research Center. All these data (from the 1984/85, 1985/86, 1987/88, and 1988/89 austral summers) are summarized in Fig. 5, in which streamlines from the first run are superimposed. The plotted wind vector for each temporary site is the vector-average wind for the entire period of record. The number in parentheses is the ratio of the mean speed at the camp to the mean speed measured over the same interval by the Byrd automatic weather station (AWS); this normalization is designed to eliminate the impact of varying observational intervals so that the speed results can be viewed as resulting from simultaneous measurements.

The simulated winds are generally consistent with the pattern of the resultant wind vectors. The resultant direction at Catchment BC (CBC) is more contour parallel than other camps near Siple Coast. At several higher elevation sites in East Antarctica, resultant directions for the summer months have been found to be considerably more contour parallel than during the winter (van Meurs and Allison 1986; Sorbjan et al. 1986; Kikuchi et al. 1988). This appears to arise due to vertical momentum mixing associated with superadiabatic lapse rates that can develop in the lowest few hundred meters (Kikuchi et al. 1988) around noon during the summer. The speed ratios at the four Siple Coast stations [Upstream C (UPC), North Camp (N), Upstream B (U), and South Camp (S)] increase steadily toward the south. The resultant speed at UPC is significantly larger than at N, but the average speeds are almost identical. This means that the directional constancy at UPC is significantly larger than at N. This result probably arises because N is situated on a ridge where it is above the drainage airflow (Bromwich 1986), while UPC is located on the exposed ice slopes. Unfortunately, this smaller-scale feature is not resolved by the present simulation and requires a finer spatial resolution to confirm this conjecture. The simulation of the wind field over Siple Coast along stations N, U, and S will be discussed in detail in the next section.

The second experiment examines the role of synoptic-scale forcing in shaping the surface-wind regime over West Antarctica and its vicinity. It started initially with the pressure field accompanying katabatic surges across the Ross Ice Shelf (Bromwich et al. 1992). The sea level pressure field was initially in geostrophic balance with the wind field. Figure 3c shows the results of the wind vectors at σ level 1 after the 24-h integration for the second experiment. Comparing Figs. 3b and 3c, it can be seen that the basic wind patterns in these two experiments are similar. Broadscale low-level diffused motion in the interior is present in both simulations. Confluence zones near the coast are in evidence in both simulations as well. It can also be seen that the wind directions over the continental interior remain almost unchanged, but the wind directions for some coastal areas, such as Siple Coast, are significantly influenced by the cyclonic synoptic pressure pattern. Wind patterns over isolated obstacles, such as Nilsen Plateau and Siple Dome near Siple Coast area, are also markedly affected, as discussed in more detail below. The terrain remains the dominant forcing mechanism in shaping the surface confluence and diffuseness, though synoptic forcing can change the magnitude of the near-surface winds. Synoptic forcing in general plays only a small role in the surface wind field over the high-elevation portions of the ice sheet (Parish 1984). It should be pointed out that the wind patterns over glaciers along the Transantarctic Mountains—the Byrd and Reeves Glaciers, for example—are hardly affected. Channeling of the downslope winds from broad interior areas of the ice sheet down these glaciers is manifest.

b. Surface winds over Siple Coast, West Antarctica

As mentioned earlier, the Siple Coast area has a different dynamic setting as compared to East Antarctica in that the terrain slopes are steeper in the interior than near the coast. This combined with the lower elevations of West Antarctica potentially allows synoptic-scale
disturbances to penetrate deeper into the ice sheet, and thus have a significant impact on the surface winds. The winds crossing Siple Coast are frequently influenced by synoptic disturbances centered in the southern Amundsen Sea (Bromwich et al. 1992). Diagnostic analysis of the vorticity at the first sigma-level over the Siple Coast area supports this idea. The average negative vorticity increases from $2.8 \times 10^{-4}$ s$^{-1}$ for the first run to $3.5 \times 10^{-4}$ s$^{-1}$ for the synoptic cyclone run. This suggests that cyclones in the southern Amundsen Sea contribute to cyclonic rotation of the surface winds over the Siple Coast area. Satellite observations (Bromwich et al. 1992) have shown that the katabatic wind signatures in TIR images have a cyclonic orientation. The weaker effect of cyclones on the interior wind regime is shown in Fig. 6, where the vertical coordinate represents the vorticity difference between the run with and without synoptic-scale forcing, and the horizontal coordinate spans the section from the interior to the Siple Coast area (shown as 4–3 in Fig. 7a). It is clear that the cyclone does not significantly influence the winds in the interior of the continent. The average increase in negative vorticity in the interior is about $1.0 \times 10^{-4}$ s$^{-1}$. Near the coast, by contrast, where the ice slope is much gentler, the increase in negative vorticity is about $1.8 \times 10^{-4}$ s$^{-1}$. The low elevation and gently sloping coastal area are more readily influenced by passing cyclones, and the katabatic outflows concentrated over this area are enhanced.

For a better understanding of the dynamics of Antarctic surface winds, the pressure gradient force in the Antarctic inversion layer can be expanded (Ball 1960). As the model atmosphere is in hydrostatic equilibrium, the horizontal pressure gradient within the inversion layer $-\rho^{-1}\nabla_h P$ can be decomposed as follows:

$$
-\frac{1}{\rho} \nabla_h P = -\frac{1}{\rho} \nabla_h P - g^*\alpha + g^*\nabla_h h,
$$

(1) (II) (III)

where $g^*$ is the reduced gravity given by $g \Delta\theta/\theta$, $g$ the acceleration due to gravity, $\Delta\theta$ the inversion strength, $\theta$ the mean potential temperature, $\alpha$ the slope of the ice sheet, and $h$ the depth of the inversion layer. Term I is the pressure gradient above the inversion layer, term II is the downslope buoyancy force, and term III is the force due to the change of the inversion-layer depth. These components can be readily calculated from the model output and the terrain data. A coastal and an interior grid point near Siple Coast have been chosen to evaluate the impact of each force in these locations (marked as A and B, respectively, in Figs. 3b and 3c). Table I presents the results for these two locations from both runs. The values are in scalar form. It is not surprising that the downslope buoyancy force (second term) is important, with the forcing in the interior being larger than at the coast because of the pattern of the inversion strength and the terrain setting outlined above. This force decreases in the second run with more decrease at the coast. This implies that the inversion strength is weakened by intrusion of the cyclone. With the introduction of the synoptic forcing, the pressure gradient above the inversion layer (term I) increases; this term is almost as large as the downslope buoyancy force in the coastal area. This is consistent with the conclusion that the wind near the Siple Coast is more subject to synoptic forcing than in the interior. The third term is small.

Figures 7a,b present an enlargement of the simulated surface-wind vectors over the Siple Coast and its vicinity from the two model runs. The run with the specified initial synoptic pressure field (Fig. 7b) displays stronger winds and more downslope flows; however, the confluence zone near Siple Coast, where the cold airflows from a large interior section of West Antarctica are concentrated, is noticeable in both runs. The average simulated wind speed in the confluence zone increases from 8.6 to near 14 m s$^{-1}$, and the wind direction becomes 20°–30° more downslope. The stronger downslope winds suggest that cold air is more efficiently transported from the interior to the coast and potentially enhances the surface flow over the Ross Ice Shelf where the slope is near zero. Bromwich et al. (1992) indicated that cyclones over the southern Amundsen Sea are often associated with katabatic winds that cross Siple Coast and propagate far across the flat Ross Ice Shelf. The run with the initial state of rest (Fig. 7a) produces nearly stagnant flows downslope of isolated peaks near the coast, such as Siple Dome and the Nilsen Plateau area. The upstream flows are blocked by these obstacles and are forced around them. With the inclusion of the synoptic pressure field, the airflows are able to ascend these mountains. The flow becomes more uniform. The simulated surface winds show significant downslope outflows especially over the Nilsen Plateau area. Studies (Bromwich et al. 1992) show that the Nilsen Plateau area is one place where prominent dark katabatic signatures are often found on wintertime TIR satellite images during katabatic surge events.

Figures 8a,b depict the simulated vertical wind speed cross section near Siple Coast along section 1–2 (see Fig. 7) for the two runs after the 24-h integration period. North Camp, Upstream B, and South Camp were temporary observational stations around Siple Coast aligned in a north–south direction. It is clearly shown that the simulated wind speed increases from North Camp to Upstream B to South Camp in the both runs. A significant increase in wind speed appears between Upstream B and South Camp. The confluence zone is concentrated on the southern part of the Siple Coast. In the run without synoptic pressure gradients (Fig. 8a) the maximum wind axis of about 15 m s$^{-1}$ is located south of South Camp, where the terrain slope is steeper; its northern boundary is between Upstream B and South Camp. This is similar to the analysis by Brom-
wich and Du (1991). Bromwich (1986) summarized simultaneous surface wind observations taken at the three camps during the austral summers of 1984–86 as shown in Table 2. It can be seen that both wind directional constancy and resultant wind speed increase steadily toward the confluence zone, which is generally located to the south of Upstream B (see also Fig. 5). South Camp, the southernmost of the three stations, experiences the strongest wind speeds and the highest directional constancies. South Camp is most likely located in the confluence zone whose northern edge is close to, but south of, Upstream B. It is likely that the summertime winds are weaker than those during winter because of the decreased strength of the surface temperature inversion due to shortwave radiative heating, but the airflow pattern should be similar (compare Parish et al. 1993). Our simulated wind field confirms the conclusions reached earlier by Bromwich (1986).

It can be further inferred from Table 2 that the higher directional constancy observed at Upstream B than at North Camp arises because Upstream B sometimes is embedded within the confluence zone (Bromwich 1986). The simulated wind field strongly supports this inference. In the run with the specified pressure field, the zone of strong winds becomes wider and deeper with larger vertical wind shear (Fig. 8b). A shallower and weaker wind zone with smaller vertical shear extends north of Upstream B. The location of the strongest winds does not change significantly but the speeds are nearly twice as strong. The height of the maximum wind speed increases from about 100 m to more than 200 m. The important role of vertical mixing in determining the wind regime in the boundary layer is emphasized. The strong winds over Siple Coast area due to the influence of synoptic cyclones lead to the enhancement of the vertical exchanges of momentum and sensible heat. This is shown by Fig. 9, which displays the temperature difference at sigma level 1 between the second run and the first run along the same transect as Fig. 8. The simulated near-surface temperature in-

![Streamlines](image-url)
Fig. 3. (Continued)
creases toward the confluence zone from about 2°C to near 9°C in response to the stronger vertical mixing within the lowest portion of the atmosphere. The weakened inversion strength, reflected by a decrease in the downslope buoyancy force, has been seen in Table 1. The stronger katabatic winds within the confluence zone efficiently mix a deeper atmospheric layer.

It seems that the location of the strongest winds to the south of South Camp is barely affected by the synoptic-scale disturbances centered in the southern Amundsen Sea (Fig. 8). Presumably, the shape of the ice terrain is the major factor locating the strongest part of the confluence zone. This is important in understanding the highly irregular wind pattern over the Antarctic continent. Qualitatively, the location of the simulated confluence zone is consistent with the observed wind data (Bromwich 1986), but the strongest wind zone south of South Camp above 200 m in altitude (Fig. 8b) needs observational verification.

c. **Katabatic surges over Ross Ice Shelf**

Parish and Bromwich (1986) suggested that the convergence zone near Siple Coast may continually add large amounts of cold air to the boundary layer over the Ross Ice Shelf. Analyses of satellite imagery and AWS data (Bromwich 1989a; Bromwich et al. 1992; Carrasco and Bromwich 1993) demonstrated the frequent occurrence of dark signatures on the Ross Ice Shelf near Siple Coast; these features probably reflect the presence of cold katabatic airstreams blowing down from Marie Byrd Land. At times, the dark signature can be traced from southern Marie Byrd Land to the northwestern edge of the shelf where a prominent coastal polynya is formed. The dark signatures occur in conjunction with the passage of synoptic cyclones over the southern Amundsen Sea. Isobars of the synoptic cyclones become parallel to the Transantarctic Mountains. The katabatic winds from West Antarctica
Fig. 5. Summer surface-wind measurements over West Antarctica collected during the Siple Coast Glaciology Project. For each site (CBC, UPC, N, U, S, and Downstream B (DNB)), the resultant wind vector over the period of record is plotted in conventional notation; $U$, $N$, and $S$ are discussed in more detail in conjunction with Figs. 7–9. The numbers in parentheses are the ratio of the camp speeds to those measured by the Byrd AWS. Dashed lines are elevation contours in meters. Streamlines from the first run are superimposed.

Fig. 6. Downslope variation of vorticity difference between the two runs along section 4–3 in Fig. 7a.

Blowing across the ice shelf and the glacier winds from East Antarctica are accelerated by the synoptic pressure gradient and generally propagate parallel to Transantarctic Mountains. The occasional presence of dark signatures across the entire ice shelf reveals the great horizontal distance (~1000 km) that the katabatic winds can cover when they have the appropriate synoptic-scale support.

The major sources of katabatic airflow across Ross Ice Shelf, as shown in Fig. 3, are West Antarctica and the glaciers flowing down from East Antarctica through the Transantarctic Mountains; among the latter, Byrd Glacier makes the largest contribution. Many studies have shown (D’Aguanno 1986; Bromwich 1989b;

| Table 1 | Scalar values of pressure gradient components (in $\text{m s}^{-1}$ $\text{h}^{-1}$) [(Eq. (3)] over West Antarctica resulting from the model output |
|---------|-----------------|-----------------|-----------------|-----------------|
|         | Run 1 | Run 2 | Run 2 | Run 2 |
| Term    | Coast | Interior | Coast | Interior |
| I       | 0.5   | 0.2     | 4.8   | 4.9     |
| II      | 6.8   | 8.4     | 5.2   | 7.1     |
| III     | 0.3   | 0.1     | 0.2   | 0.2     |
Bromwich et al. (1992) that dark signatures are commonplace near East Antarctic glaciers on wintertime TIR satellite images of the Ross Ice Shelf. Bromwich (1989b) found that a high frequency of these katabatic wind signatures was characteristic of the Ross Ice Shelf near Byrd Glacier. The Byrd Glacier feature is situated downwind of a confluence zone over the East Antarctic ice sheet, which results in intensified and more persis-
tent katabatic winds at the foot of the glacier. It was inferred that these prominent warm signatures are actually generated by comparatively cold and negatively buoyant airflows.

The basic patterns of simulated wind vectors over the Ross Ice Shelf for the initial states of rest and synoptic-scale disturbance appear to be similar (Figs. 3b,c). The airflow over the shelf parallel to the Transantarctic Mountains is associated with katabatic winds coming from southern Marie Byrd Land and propagating parallel to the mountains to the northwestern edge of the shelf. Fed by the readily available cold air crossing Siple Coast and strengthened by the synoptic pressure field, the speed of airflow increases from 8.8 to 11.9 m s$^{-1}$, and the mountain-parallel airflow becomes wider. The wind data recorded by AWS 08 located on the southeastern part of the Ross Ice Shelf show that the wind speed averaged 3.9 m s$^{-1}$ for five winter months of 1988 and increased to 7.0 m s$^{-1}$ for all signature days (days with at least one satellite image showing the katabatic signature far across the Ross Ice Shelf) in the same period (Bromwich et al. 1992). The cold katabatic air mass from Byrd Glacier merges with the flow from West Antarctica and enhances the wind speed over the northwestern part of the Ross Ice Shelf. As will be seen later in Table 4, wind speed near Byrd Glacier at AWS 15 for signature days increased by 2.2 m s$^{-1}$ compared to the winter mean.

The model results can be used to investigate the dynamics of the simulated airflow across the Ross Ice Shelf. In Fig. 10a the depth of the surface inversion is plotted along with the sigma level 1 wind vectors for the first run. A much deeper inversion is present along the Transantarctic Mountains (~550 m) than out over the Ross Ice Shelf (~280 m). Using (3) again (see Table 3), term II is very small (0.2 m s$^{-1}$ h$^{-1}$) because the ice shelf is nearly flat. Term III at 3.6 m s$^{-1}$ h$^{-1}$ dominates the total pressure gradient force with term I being small (0.4 m s$^{-1}$ h$^{-1}$). These terms were estimated between points inside (A) and outside (B) the cold mountain-parallel airstream; AB is normal to the flow. The force balance indicates that the airflow results primarily from a geostrophic balance between the pressure gradient due to the cold air piled up against the mountains (term III) and the Coriolis force. This implies that this flow is a barrier wind (Schwerdtfeger 1984); however, the cold air is here supplied by katabatic drainage. The following governing dynamics can be envisaged. When a katabatic wind blows down onto the flat ice shelf it loses the downslope buoyancy forcing. The Coriolis force turns the flow to the left over the almost frictionless ice shelf and eventually reaches the mountains where it must pile up because the stratified flow cannot surmount such a vertical wall. Eventually the above geostrophic balance is established, resulting in what might be called a katabatic barrier wind.

Figure 10b shows a more complicated situation for the run with the specified synoptic pressure gradient. Here there is a shallow layer of cold air along the mountains (~160 m) with a much deeper layer over the ice shelf (~500 m). In between there is a narrow zone (~20 km) where much deeper cold air (~1200 m) is simulated; this feature cannot be depicted in Fig. 10b, because of its small width. This feature has some aspects in common with observed hydraulic jumps (Pettré and Andre 1991). Evaluating (3) between points C and D shows that the synoptic pressure gradient force (6.4 m s$^{-1}$ h$^{-1}$) is opposed by the sloping depth of cold air perpendicular to the mountains (~3.0 m s$^{-1}$ h$^{-1}$). Term II is still very small (0.8 m s$^{-1}$ h$^{-1}$). The flow is close to being geostrophic with the total pressure gradient force (4.2 m s$^{-1}$ h$^{-1}$) nearly balancing the Coriolis force (5.9 m s$^{-1}$ h$^{-1}$). Comparison of Figs. 10a and 10b suggests that the entire pattern in Fig. 10a has been shifted northward by the synoptic pressure gradient with the abrupt increase in inversion depth that was present near the foot of the terrain slope now essentially being found near the center of the ice shelf with an orientation that approximately parallels the Transantarctic Mountains. It should be emphasized here that the estimation of the terms reflects the qualitative relationship and is location sensitive.

To evaluate the thermal characteristics of the surface airflow over Ross Ice Shelf where the dark satellite signatures are often found, the simulated sigma level 1

| Table 2. Simultaneous surface wind statistics for Siple Coast stations (after Bromwich 1986). |
|---------------------------------|-----------------|-----------------|
| Temporary location             | Resultant wind | Mean speed      |
|                                | (m s$^{-1}$)    | (m s$^{-1}$)    |
|                                | Directional     | Dir. constancy  |
|                                | constancy       |                 |
| 21 December 1984–1 January 1985
| North Camp                      | 66°, 1.3        | 0.41            | 3.2 |
| Upstream B                      | 92°, 2.6        | 0.74            | 3.5 |
| South Camp                      | 87°, 6.4        | 0.97            | 6.6 |
| 20 December 1985–11 January 1986
| North Camp                      | 13°, 0.4        | 0.25            | 1.6 |
| Upstream B                      | 87°, 1.0        | 0.59            | 1.7 |
| South Camp                      | 99°, 3.9        | 0.90            | 4.3 |

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**Fig. 9.** The temperature changes between the two runs at sigma level 1 along the 1–2 transect of Fig. 7.
temperatures were averaged for the dark signature area and elsewhere on the shelf for both model runs. In the first run the average temperature difference between the dark signature area (southwest part of the shelf) and the other area of the shelf (northeast part) is about −2.7°C. In this case, the air temperature is colder along the Transantarctic Mountains part of the shelf. In the second run, the average temperature difference is about 6.6°C. The temperature over the dark signature area is much higher than that elsewhere on the Ross Ice Shelf. These warmer surface-air temperatures than in adjacent areas are probably associated with the dark signature often seen on wintertime clear-sky TIR satellite images in conjunction with katabatic surges and synoptic-scale cyclones over the southern Amundsen Sea. D’Aguanno (1986) found that the TIR emission temperature of the katabatic signature emerging from Byrd Glacier was about 10°C higher than that outside the katabatic stream. This warm air mass was inferred to be due to adiabatic compression during descent of the cold katabatic air from the Antarctic interior. It is unlikely, however, that the warming of the near-surface air over the nearly flat Ross Ice Shelf is due to the same cause. Based on available observational data, Bromwich (1989b) constructed an illustrative sketch (Fig. 11) to explain the warm (dark) katabatic signatures repeatedly occurring on the Transantarctic Mountains side of the Ross Ice Shelf. It was argued that if the radiation inversion can be substantially weakened by turbulent vertical mixing, then the katabatic temperature profile like Fig. 11 can be generated by bringing much warmer and higher-speed air down to the surface, whereas the airflow outside the katabatic surges is less impacted by the downward turbulent transport of momentum and sensible heat, and continues to be much colder. Observational results from two aircraft missions (Parish and Bromwich 1989) confirmed the above argument. This is also consistent with the model results presented above. The intense vertical mixing in the planetary boundary layer is one of the important processes producing the dark satellite signatures on the Ross Ice Shelf.

To compare the model results more fully with the available observational data, surface weather observations obtained from the network of AWS installed on the Ross Ice Shelf were examined. Three-hour observations of air pressure, temperature, wind speed, and wind direction were taken from Keller et al. (1989).
AWS data from signature days (as in Bromwich et al. 1992) associated with synoptic-scale cyclones over the southern Amundsen Sea were compared with corresponding mean values for April–August 1988. Table 4 summarizes the results in which departures of the pressure, temperature, wind speed, and directional constancy for signature days from the five-month average are provided.

Figure 12 presents geographically the pressure and temperature differences. The directional constancies for signature days are also plotted. Dashed lines are five-month resultant wind directions, and the solid lines with wind speeds plotted in conventional fashion are the vector-average winds for signature days. The terrain is plotted from the digitized Antarctic map used in the simulations. It is clear that the decrease of pressure over the Ross Ice Shelf on signature days is significant.

The larger pressure decreases (5.5–7.5 hPa) over the eastern part of the Ross Ice Shelf are due to the influence of the cyclones around the southern Amundsen Sea. Note the large warming over the southern part of the Ross Ice Shelf (AWS 08 and 11), where the katabatic surges are frequently observed on winter TIR satellite images (Fig. 13). The wind speed also increased on dark signature days.

It is interesting to note that the wind direction at AWS 15 is approximately parallel to the orientation of Byrd Glacier, which is downwind of a major confluence area in the continental interior. However, the wind direction at AWS 24, which is located near AWS 15 but far away from Byrd Glacier, is nearly the same as that at AWS 15. This demonstrates that the wind direction at AWS 24 is still controlled by katabatic outflow from Byrd Glacier. The greater increase in wind directional constancy at AWS 24 (0.14) than at AWS 15 (0.08) suggests that this site is more influenced by katabatic winds from Byrd Glacier during signature days, whereas AWS 15 is usually affected by this wind regime. The wind direction at AWS 11, which is near Beardmore Glacier, is also influenced by katabatic winds from that glacier. AWS 08 is located on the shelf far from the glaciers, and its wind direction is primarily influenced by airflow from southern Marie Byrd Land.

<table>
<thead>
<tr>
<th></th>
<th>Run 1</th>
<th>Run 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Term I</td>
<td>0.4</td>
<td>6.4</td>
</tr>
<tr>
<td>Term II</td>
<td>0.2</td>
<td>0.8</td>
</tr>
<tr>
<td>Term III</td>
<td>3.6</td>
<td>-3.0</td>
</tr>
</tbody>
</table>
The simulated winds are in broad agreement with the observations, given that the modeled convergent flow is not as focused as in the real world due to the smoothed topography.

Figure 14 is the simulated sea level pressure field resulting from the initially specified synoptic-scale disturbance with the sigma level 1 wind vectors superimposed. The high pressure values over the Antarctic interior are known to be spurious due to the high elevations and thus are not shown. The near-surface airflow in the interior is driven mainly by the downslope pressure gradient generated by the strong radiational cooling of the ice slopes and the steep terrain slope. The katabatic winds over East Antarctic glaciers blow down to the coastal margin crossing the isobars at a large angle. As the katabatic airflows move onto the Ross Ice Shelf where the slope is near zero, they turn left and then run more or less parallel to the steep Transantarctic Mountains. This is consistent with TIR satellite observations as in Fig. 13. Surface observations from the AWS over the Ross Sea–Ross Ice Shelf area can be used to construct the sea level pressure field (e.g., Keller et al. 1989). Observational analysis (Bromwich 1992) revealed that for one case a well-defined low pressure trough was positioned over the eastern side of the Ross Ice Shelf. A winter-long study (Bromwich et al. 1992) reached a similar conclusion. The katabatic surge is associated with the synoptic-scale cyclone over the southern Amundsen Sea, and its trough extending southwestward toward the Ross Ice Shelf. This regional atmospheric circulation is reasonably well simulated (Fig. 14).

d. Katabatic surges and polynya formation

Polynyas are common features of the seasonal sea-ice zone surrounding Antarctica. Their geographic distribution and short-term variability are seen in TIR images collected by polar-orbiting meteorological satellites. The area of the polynya is closely related to the surface-wind regime (Knapp 1972; Cavalieri and Martin 1985). Accurate wind data over polynyas are un-

<table>
<thead>
<tr>
<th>AWS</th>
<th>Δ* Wind speed (m s⁻¹)</th>
<th>Δ Pressure (hPa)</th>
<th>Δ Temperature (°C)</th>
<th>Δ Constancy</th>
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<tr>
<td>00</td>
<td>0.8</td>
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</tr>
<tr>
<td>07</td>
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<tr>
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<tr>
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<td>−5.5</td>
<td>5.5</td>
<td>0.01</td>
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<tr>
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</tr>
<tr>
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</tr>
<tr>
<td>24</td>
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</tr>
<tr>
<td>25</td>
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<td>−7.3</td>
<td>1.9</td>
<td>0.15</td>
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<tr>
<td>05</td>
<td>4.2</td>
<td>−2.8</td>
<td>0.1</td>
<td>0.06</td>
</tr>
</tbody>
</table>

* Difference between signature days and the five-month mean.
available because of the lack of surface observations. The Terra Nova Bay polynya is formed by strong katabatic drainage winds and is influenced by the synoptic conditions (Kurtz and Bromwich 1983; Bromwich and Kurtz 1984). The northwestern edge (west of 180°) of the Ross Ice Shelf is a place where a prominent polynya is often observed (Zwally et al. 1983, 1985). Bromwich et al. (1993) indicated that expansions of this polynya are associated with katabatic surges crossing the shelf from West Antarctica with the assistance of the regional pressure field and from the main glaciers through the Transantarctic Mountains. Another important factor for maximum polynya size is air temperature. For a given offshore wind speed, colder air produces a smaller polynya, and vice versa (Pease 1987).

Figure 15a depicts the surface-wind profile between the 180° meridian and Ross Island along the northwestern edge of the Ross Ice Shelf from the model results after 24-h integration. Without the synoptic forcing (solid line), the wind speed is about 12 m s⁻¹ near Ross Island and decreases toward the east. In the second run, the wind speed increases significantly (dashed line). The average increase is more than 2 m s⁻¹ and is most marked near 180°. The surface-wind field over the northwestern edge of the shelf is significantly influenced by the synoptic-scale cyclones over the southern Amundsen Sea.

To estimate the steady-state polynya size north of the edge of the shelf, wind speed and air temperature from the results of the synoptic forcing run are used as input into a wind-driven coastal polynya model (Pease 1987). This model does not consider shortwave radiation and thus can be used to examine winter polynyas at high latitudes. Both the sensible heat flux and latent heat flux are calculated from the bulk aerodynamic formulas. Although the evaporative heat flux is neglected in the Pease (1987) model, a later paper (Smith et al. 1990) noted the importance of latent heat for winter polynya formation. Here, we choose the sensible heat transfer coefficient to be $1.25 \times 10^{-3}$ and the latent
heat coefficient to be $1.33 \times 10^{-3}$ (following Kurtz and Bromwich 1985). The relative humidity of the air is taken to be 30%, because the results are relatively insensitive to the presumed moisture content of the atmosphere. Water surface temperature varies only by several tenths of a degree for any given polynya (Schumacher et al. 1983), so the polynya upward longwave radiation is assumed to be constant. The simple approximation of $\sigma_e a^4 T_a^4$ is used to calculate the downward longwave radiation, where $\sigma_e$ is the Stefan–Boltzmann constant, $a$ is the effective emissivity of the air taken to be 0.95, and $T_a$ is the air temperature. This ignores the effects of clouds and water vapor in the atmosphere. For our case, the wind speed and air temperature are treated as variables and are taken from the lowest model level.

Figure 15b shows the calculated maximum polynya size for the second run along the same transect as Fig. 15a. The size ranges from about 10 km to more than 12 km in the vicinity of Ross Island. The difference exists because near Ross Island the air temperature is higher and the wind speed is stronger. These meteorological conditions favor the formation of a larger polynya than around 180°. When a dark satellite signature was observed over the shelf, a larger polynya was found along the northern edge of the ice shelf, especially along the western side (Bromwich et al. 1992). It is interesting to compare the calculated polynya width to the satellite-observed polynya width associated with katabatic surges crossing the Ross Ice Shelf (Bromwich et al. 1993). The polynya width can be estimated from TIR satellite images by counting the number of polynya pixels. It should be pointed out that some uncertainties exist in determining such pixels. A polynya is an area of both open water and thin ice and has much higher temperatures than its surrounding ice.
cover. It appears to be dark on TIR satellite images. Practically, the dark pixels are counted. Errors may occur near the transition from dark to brighter tones—whatever digital criterion is used to define a polynya pixel. The typical satellite-observed polynya width during the katabatic surges was about 16 km (Bromwich et al. 1993). The calculated width roughly agrees with the satellite observations, though it is underestimated. This demonstrates the ability of the model to produce qualitatively realistic meteorological conditions over the area where prominent polynyas are often observed.

4. Conclusions

Numerical simulations of katabatic winds over and beyond West Antarctica have been conducted using a three-dimensional, primitive equation model. Antarctic ice topography at 20-km resolution has been used and resolves all the topographically important terrain features in the model domain, such as Siple Coast, Nilsen Plateau, and Byrd Glacier. The terrain-induced katabatic winds are well simulated. The model results suggest that the surface wind pattern over the Antarctic interior is little affected by synoptic-scale disturbances beyond the ice-sheet margin. This is illustrated by the model diagnostic analysis of vorticity and pressure gradient forcing upslope from Siple Coast. The Siple Coast area, having lower elevations and gentler slopes than the interior, is subject much more to the effects of synoptic-scale cyclones. The enhanced katabatic winds result in stronger vertical mixing within the planetary boundary layer. This tends to weaken the surface temperature inversion, allowing the synoptic pressure gradient force to become of comparable importance to the downslope buoyancy force. Strong and persistent katabatic winds downwind of confluence zones are sustained by cold airflow from the broad interior section. The model produces the well-known confluence zones near the Antarctic coast.

Abundant observational evidence (Stearns and Wendler 1988; Bromwich 1989a; Bromwich et al. 1992) shows that katabatic wind signatures are common features over the Ross Ice Shelf on wintertime TIR satellite images. The simulated katabatic surges are in good agreement with the satellite analyses that show the katabatic winds blowing across the flat Ross Ice Shelf in a geostrophic fashion. The model suggests that the katabatic surges are fed mainly by cold airmasses...
originating from the Siple Coast area of West Antarctica and from the Byrd Glacier section of East Antarctica. This is in agreement with satellite imagery interpretation (Bromwich 1989a; Bromwich et al. 1992). The model outputs are treated as inputs into a wind-driven polynya model to estimate the polynya width for those areas where the prominent polynyas are often observed in TIR images. It is shown that the calculated polynya width is consistent with the observations.

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