

A Numerical Study of Strong Katabatic Winds over Antarctica

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(Manuscript received 17 September 1983, in final form 25 November 1983)

ABSTRACT

Certain coastal sections of Antarctica, most notably Adelie Land and Terra Nova Bay, experience anomalously intense, persistent katabatic winds. The forcing of such katabatic outflow is believed to originate several hundred kilometers upslope in the interior of the continent where cold air drainage currents from a large area converge into a relatively narrow zone focused on the steeply-sloping ice terrain near the coastline. Numerical simulations with a three-dimensional hydrostatic model incorporating terrain features representative of Adelie Land reveal a significant topographical channeling of the surface airflow. Katabatic wind speeds as depicted by the model are greatly enhanced downslope of the convergence channel. These results emphasize the importance of topography in the continental interior in shaping the character of coastal katabatic flow.

1. Introduction

Katabatic winds are commonplace over the ice slopes along the coastal rim of the Antarctic continent. Some of the most persistent and most intense surface winds have been recorded at coastal stations in East Antarctica which lie at the terminus of the steep ice escarpment fully exposed to the cold air drainage from the continental interior. Without doubt, the primary forcing mechanism for katabatic winds is the pronounced radiational cooling of air adjacent to the sloping ice surface. Theory suggests and observations show that katabatic flows display a marked directional preference (see Table 1). The extremely high directional constancy values are characteristic of topographically-forced winds. Furthermore, resultant wind directions and/or sastrugi orientation indicate a consistent relationship between terrain slope and deviation angle from the fall line as modified by the low-level stability and the Coriolis parameter. Near the steep coastal slopes, winds are directed nearly down the fall line; ice slopes in the interior of Antarctica are more gentle and consequently winds are directed at significantly larger angles from the fall line in response to the increased importance of Coriolis turning (Parish, 1982). Synoptic features, such as migratory cyclones often situated in the intense baroclinic zone near the Antarctic coast, can be of local importance as well. Ball (1960) notes that katabatic wind speeds increase as east-to-west moving cyclones approach, although changes in wind direction are small.

Certain sections of the Antarctic coast experience anomalously strong katabatic winds. The Australasian Antarctic Expedition of 1911–1914 under the leadership of Douglas Mawson established a base camp, Cape Denison, along the coast of Adelie Land (Fig.

1). After collecting wind data for two years, the party returned to Australia to report the exceptionally persistent, intense katabatic winds encountered at Cape Denison. Similar conditions were experienced nearly 40 years later by a French expedition which wintered two years at Port Martin, some 60 km west of Cape Denison along the Adelie Land coast. Resultant wind statistics for both stations are listed in Table 1. The mean yearly katabatic wind speeds are nearly 75% greater than those recorded at other stations located at the foot of the steep terminal ice slopes. Another coastal region which experiences abnormally intense katabatic flow is situated near Terra Nova Bay, the site where Scott's Northern Party was forced to spend the winter of 1912. The journals maintained by party members remain the only wintertime observations. Bromwich and Kurtz (1982) and Kurtz and Bromwich (1983) have re-examined these observations and propose that the katabatic nature of Terra Nova Bay is similar to the Cape Denison–Port Martin region. They also conclude the strong winds are responsible for the maintenance of open water in Terra Nova Bay throughout even the coldest winter months. The unique character of the katabatic wind at such locations has long been of interest to Antarctic meteorologists. In a summary article dealing with possible explanations for the extreme katabatic winds of the Cape Denison–Port Martin area, Loewe (1974) noted that terrain slopes of Adelie Land are not significantly steeper than other coastal sites, and that the strong winds did not seem related to the passage of synoptic cyclones.

2. General description of strong katabatic winds

According to Lettau and Schwerdtfeger (1967), one serious limitation to the occurrence of katabatic winds

TABLE 1. Mean yearly resultant wind (m s^{-1}) statistics for stations in the interior and coastal regions of Antarctica. (N refers to the length of the data record in years; the deviation angle is the angle between the resultant wind and fall line of the ice terrain.)

	N	Resultant wind	Directional constancy	Deviation angle
Interior stations				
Pionerskaya	1	9.3, 131°	0.92	~55°
Charcot	1	8.6, 163°	0.91	~50°
Vostok	16	4.1, 243°	0.81	~60°
Amundsen-Scott	16	4.6, 039°	0.79	~50°
Byrd	14	6.6, 013°	0.86	~60°
Coastal stations				
Cape Denison	2	19.0, 161°	0.97	~20°
Port Martin	2	16.9, 146°	0.94	~5°
Mawson	15	9.8, 130°	0.93	~10°
Mirny	17	9.7, 127°	0.90	~50°
Molodezhnaya	11	8.4, 126°	0.85	~10°

is the supply of cold air upslope. They note even moderate katabatic winds rapidly drain an existing source of negatively-buoyant air. Therefore, persistent katabatic flow requires replenishment of cold air either by convergence of drainage currents upwind in the continental interior or some localized intense cooling. Without some means of providing a continual supply of cold air, katabatic flows must be restricted to a short-lived, high-intensity nature. Further, this suggests that the katabatic potential of coastal stretches is largely determined by surface radiative characteristics and drainage patterns in the interior of the continent. It is impossible to account for observed variations in the

nature of katabatic winds by simply focusing on the local terrain features. Consideration must be given to the entire katabatic setting, including the dynamic and thermodynamic processes upslope.

Parish (1981) attributed the strong, persistent katabatic winds along Adelie Land to the convergence of surface drainage currents over a broad section of the interior. As revealed in the time-averaged pattern of surface streamlines from Parish (1982) shown in Fig. 2, the topography south of the Cape Denison-Port Martin area tends to disrupt the radial drainage of cold air off the interior ice dome, resulting in a confluence of negatively-buoyant air from a wide area into a narrow zone directed towards the coast of Adelie Land. A similar convergence feature can be located upslope of Terra Nova Bay. In general, drainage currents converge east of ridge features or west of valleys. The cold air concentration provides a large reservoir from which coastal katabatic storms can feed.

3. Numerical simulation of katabatic winds

a. Model equations

The picture set forth in Section 2 is that katabatic flow at locations along the Antarctic coast is profoundly influenced by the supply of cold air made available through radiational cooling and surface drainage patterns in the Antarctic hinterland. In order to test this conceptual hypothesis, numerical simulations of katabatic winds have been carried out using a three-dimensional, hydrostatic, primitive equation model. Patterned after Anthes and Warner (1978; hereafter referred to as AW), the model is written in terrain

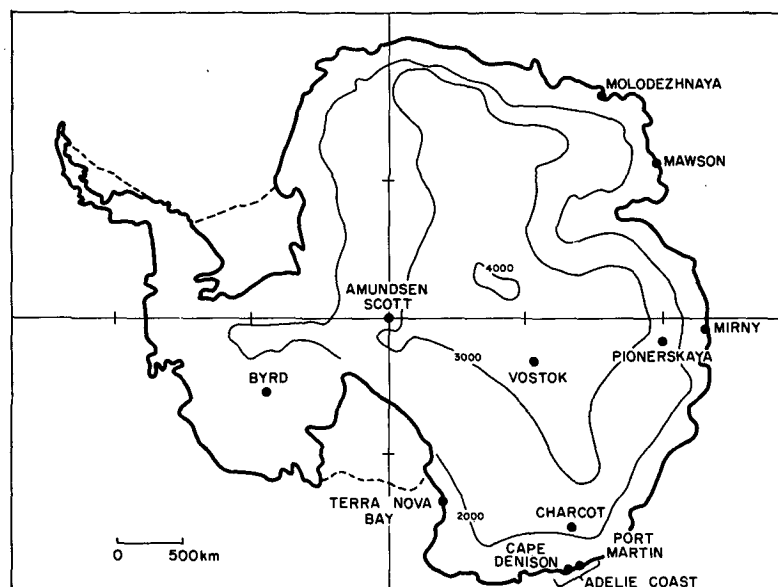


FIG. 1. The Antarctic Continent including positions of stations mentioned in text. Contour lines drawn for 2000, 3000 and 4000 m only.

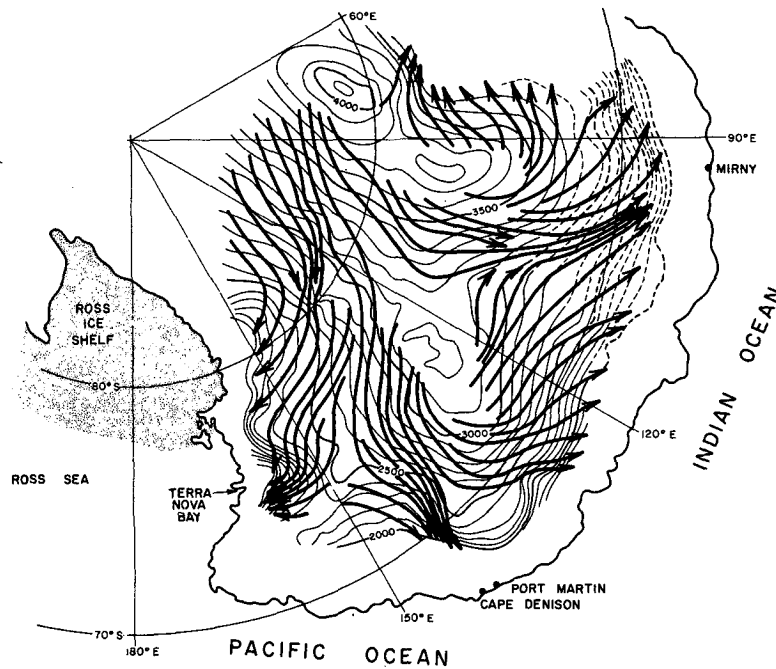


FIG. 2. Time-averaged winter flow pattern over the surface of the Antarctic based on model wind calculations of Parish (1982).

following σ -coordinates to allow for irregular ice terrain. The vertical coordinate σ is defined by

$$\sigma = \frac{p - p_t}{p_s - p_t} = \frac{p - p_t}{p^*}, \quad (1)$$

where p is pressure, p_t the pressure at the top of the model (set to 25 kPa), and p_s the surface pressure.

The flux forms of the horizontal scalar equations of motion, written in σ -coordinates, are:

$$\frac{\partial p^*u}{\partial t} = - \frac{\partial p^*uu}{\partial x} - \frac{\partial p^*vu}{\partial y} - \frac{\partial p^*u\sigma}{\partial \sigma} - p^* \left[\frac{RT}{(p^* + p_t/\sigma)} \frac{\partial p^*}{\partial x} + \frac{\partial \phi}{\partial x} \right] + p^*vf + FU, \quad (2)$$

$$\frac{\partial p^*v}{\partial t} = - \frac{\partial p^*uv}{\partial x} - \frac{\partial p^*vv}{\partial y} - \frac{\partial p^*v\sigma}{\partial \sigma} - p^* \left[\frac{RT}{(p^* + p_t/\sigma)} \frac{\partial p^*}{\partial y} + \frac{\partial \phi}{\partial y} \right] - p^*uf + FV. \quad (3)$$

The terms FU and FV refer to the horizontal and vertical diffusion of momentum and will be discussed later. Other symbols have their usual meteorological meaning. Since the region of interest for modeling katabatic winds covers an area approximately 400 km \times 400 km, the variation in the Coriolis parameter with latitude is ignored as are the earth curvature terms in the preceding equations.

The complete advective form of the thermodynamic equation used in the model is

$$\begin{aligned} \frac{\partial T}{\partial t} = & \left(\frac{RT}{p^*\sigma + p_t} \right) \frac{\sigma}{c_p} \frac{\partial p^*}{\partial t} + \dot{\sigma} \frac{\partial T}{\partial \sigma} - \left(\frac{RTp^*}{p^*\sigma + p_t} \right) \frac{\dot{\sigma}}{c_p} \\ & - \left(u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right) + \frac{1}{c_p} \left(\frac{RT\sigma}{p^*\sigma + p_t} \right) \\ & \times \left(u \frac{\partial p^*}{\partial x} + v \frac{\partial p^*}{\partial y} \right) + \frac{Q}{c_p} + FT. \quad (4) \end{aligned}$$

The diabatic heating term is represented by Q . The primary diabatic influence to be considered for katabatic wind modeling is the radiational cooling. Latent heating effects such as condensation and sublimation are neglected owing to the extreme cold and consequently low water vapor content of the Antarctic atmosphere. The FT term again represents effects of horizontal and vertical diffusion. Included in this term is the turbulent transfer of heat. Seaman and Anthes (1981) have performed experiments with various finite difference forms of the thermodynamic equation; they note that the advective form possesses superior stability characteristics compared to flux forms such as that used by AW.

The fourth prognostic equation is the continuity equation,

$$\frac{\partial p^*}{\partial t} = - \frac{\partial p^*u}{\partial x} - \frac{\partial p^*v}{\partial y} - \frac{\partial p^*\dot{\sigma}}{\partial \sigma}. \quad (5)$$

The hydrostatic equation is required to complete the system. In σ -coordinates it is written

$$\frac{\partial \phi}{\partial \ln(\sigma + p_0/p^*)} = -RT. \quad (6)$$

b. Finite difference techniques

The horizontal grid used in the katabatic wind study consists of a 20×20 array with a grid spacing of 20 km. Horizontal motion components are staggered midway between the mass variables T , p^* and ϕ . The hydrostatic equation is integrated directly to the σ -levels, thereby eliminating any errors from interpolating between σ -surfaces. The S1 finite difference forms from Anthes (1972) are used to estimate the horizontal and vertical derivatives. As shown in AW, the finite difference forms conserve mass and momentum exactly and approximately conserve energy. A total of ten vertical levels are incorporated in the model; the spacing between levels is variable with the highest resolution in the lowest portion of the atmosphere in order to permit a more detailed katabatic wind structure.

Model equations have been integrated using the method of Brown and Campana (1978), which allows a time step nearly twice as large as the conventional centered-in-time (leapfrog) approach. A series of experiments have been conducted in order to determine proper lateral boundary conditions. As in AW, the best results were obtained by specifying mass variables on the boundaries and extrapolating the normal flux terms p^*u and p^*v . To help suppress spurious boundary reflections, horizontal diffusion coefficients were increased in the vicinity of the end grid points.

c. Planetary boundary layer parameterization

The planetary boundary layer (PBL) represents that part of the atmosphere where turbulent transfer of heat and momentum occur. In the unstable PBL a strong positive heat flux arising from surface heating often leads to a well-mixed boundary layer capped by an inversion. The height of such a boundary layer is therefore well-defined. However, under stable conditions the PBL depth is not nearly as well marked. Theoretical and observational studies often show a relationship between stable PBL depth and the ratio u_*/f . From observations of Clarke (1970), Deardorff (1972) estimated a PBL depth of $0.23 u_*/f$. For katabatic flow, a value of $0.2 u_*/f$ was assigned. The friction velocity u_* in the model is related to the magnitude of the surface wind (wind at the lowest σ -level) by the empirical expression

$$u_* = 0.055V_s. \quad (7)$$

The formula was obtained by a linear fit to the similarity functions of Businger *et al.* (1971) over a range of cooling rates and valid for the 10 m level. A roughness parameter of 0.01 m was assumed. The above relationship gives u_* values within 10% of those from

similarity theory without using the time-consuming iterative schemes required to find the similarity-derived values.

The lowest σ -level is assumed to be at the top of the surface boundary layer (SBL). The stress in the SBL is expressed in vector form by

$$\tau = \rho u_*^2 \frac{V_s}{V_s}. \quad (8)$$

A bulk parameterized form of the surface heat flux in the SBL, similar to that of Deardorff (1972), is used. Mathematically, the surface heat flux is expressed by

$$H_0 = \rho c_p C_h (T_g - T_s) V_s,$$

where C_h is assigned a value of 0.002. The ground temperature T_g is obtained using the surface energy budget equation of Blackadar (1978):

$$\frac{\partial T_g}{\partial t} = \frac{1}{C_g} (I_A - \sigma T_g^4 - H_0) - K_r (T_g - T_m). \quad (9)$$

Here C_g is the heat capacity of the ground (snow density of 300 kg m^{-3}), I_A the downward longwave radiation from the atmosphere (assumed constant), K_r a relaxation constant of $8.58 \times 10^{-5} \text{ s}^{-1}$ and T_m the mean surface air temperature set to 240 K.

Vertical fluxes of heat and momentum in the PBL are given by

$$-\overline{w'v'} = K_m \frac{\partial V}{\partial z}, \quad (10)$$

$$-\overline{w'T'} = K_h \frac{\partial T}{\partial z}. \quad (11)$$

Values of the eddy diffusivities are computed by the scheme of Agee *et al.* (1973). Although the method is very simple, the resulting shape of the K profiles agrees qualitatively with those of Busch *et al.* (1976) and Brost and Wyngaard (1978). In fact, the steady-state profiles for the stable boundary layer obtained by Brost and Wyngaard (1978) were used as a guide in determining the proper coefficients in the Agee *et al.* (1973) formula. The ratio of diffusivities of heat and momentum in the stable Antarctic atmosphere is assumed to be 1.0. Subgrid-scale horizontal diffusion is modeled following AW.

d. Model form of the Antarctic Continent

An idealized parabolic form of the Antarctic ice slope has been used in the model experiments. Mathematically, the height of the ice surface can be expressed by

$$z(x, y) = 2800.0 \text{ m} \left(1 - \frac{y}{3.5 \times 10^5 \text{ m}} \right)^{1/2}, \quad (12)$$

where y is the distance component (in meters) directed down the slope and x is the distance along the ice contours. Comparison of the mathematical profile with

an actual ice profile, obtained by taking height values spaced 20 km apart, starting at the coastal station Mirny from the American Geographical Society 1:5 000 000 scale map of Antarctica, is illustrated in Fig. 3. Without question, the idealized form is representative of the actual profile. There are two reasons why an idealized profile is used. First, it is hoped that such a smooth representation will help reduce the inertia-gravity waves generated during the model integration. Second, such a profile provides a convenient means for assuming periodic boundary conditions across the slope, thereby simplifying boundary conditions on two sides of the domain.

4. Model results

During the course of model development, a number of experiments were conducted to test the ability of the model to simulate certain gross features of katabatic wind and temperature profiles, as well as to infer the sensitivity of katabatic wind speeds to variations in initial model conditions. This section presents results from such model experiments for a variety of ice terrain configurations and synoptic forcing conditions. The primary purpose is to identify key mechanisms in driving strong katabatic flow such as displayed in the Cape Denison vicinity. In all experiments, an initial isothermal atmosphere of 250 K is assumed. While this may be somewhat unrealistic for the upper troposphere, observations (see Schwerdtfeger, 1970) show that such isothermal layers are common above the inversion layer in the interior of the continent, extending at times 1000 m or more above the inversion top. Near-isothermal layers in the lower atmosphere are also frequent near the Antarctic coast where strong mixing tends to prevent large temperature inversions from forming.

a. Simple katabatic flow

The first task is to ensure that the numerical model is capable of simulating basic features of Antarctic katabatic winds such as the low-level maximum in the wind profile. The first test incorporates the idealized ice terrain expressed in (12) and excludes effects of synoptic scale forcing. The isothermal atmosphere is initialized about a basic state of rest, thereby focusing directly on the terrain-induced drainage flows. The model equations are integrated for a period of 10 h by which time the slope flows should be well-defined.

During the first hour of the model run, a low-level inversion begins to form because of the radiational cooling of the lower atmosphere. Katabatic winds develop in response to the downslope-directed pressure gradient force resulting from the presence of a temperature inversion over sloping terrain. The variations in streamline orientation, wind speed and temperature for the second σ -level (~ 50 m above the ice surface) after the 10 h model integration period are shown in Fig. 4. Owing to the lack of cross-slope variations in

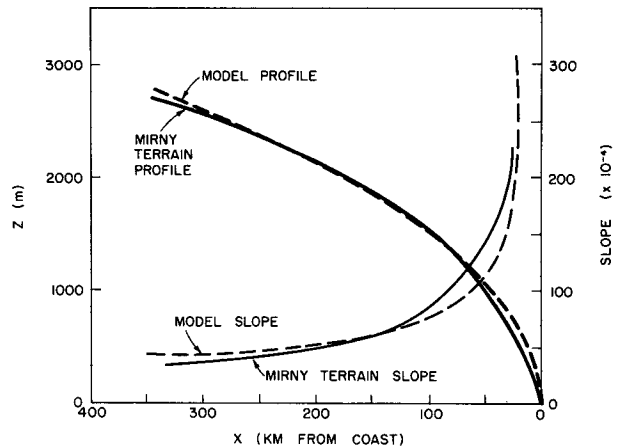


FIG. 3. Comparison between model ice profile and terrain slope with near-coastal ice topography south of Mirny.

the ice profile, the streamlines shown in Fig. 4a display a uniform character and are directed at angles across the fall line between 30 and 50°. Fig. 4b depicts the katabatic wind speeds at the 50 m level where the winds approach their maximum value. As expected, katabatic wind speeds are greatest near the model coast where the ice slopes are steepest. The temperature field is illustrated in Fig. 4c. Since the initial temperature profile is assumed isothermal, temperatures give an indication as to inversion strength. The cold temperatures (strong inversions) found over the interior of the model ice sheet are consistent with the relatively slow wind speeds and reduced mixing tendencies.

Profiles of wind and temperature after the 10-h integration period for representative interior and coastal sites are shown in Fig. 5. Wind profiles in each case are quite representative of well-developed katabatic winds, displaying the low-level maximum of wind between 50 and 250 m above ground level. Indicated wind speeds at the surface (10 m level) of ~ 12 m s⁻¹ are reasonable estimates for average katabatic-type flows and in good agreement with resultant winds at the "normal" katabatic-prone stations listed in Table 1. The location of the interior site corresponds to a height of 2200 m on the ice terrain and a slope of about 5×10^{-3} which is significantly steeper than the high Antarctic Plateau (10^{-4} or less). Charcot and Pionerskaya are two stations which may be used in comparison with model output. Both stations lie at about 2500 m and terrain slopes are probably only 60–70% of the interior model site. Resultant winds for these stations are between 8–9 m s⁻¹, comparable with the indicated surface winds of 9 m s⁻¹ from Fig. 5.

The temperature patterns in Fig. 5 show reasonable agreement with observations as well. At the coast, near-surface winds are quite strong, indicating a fair amount of turbulent mixing. Furthermore, using continuity arguments, the downslope increase of wind speeds implies a transport of warmer air from above the inversion

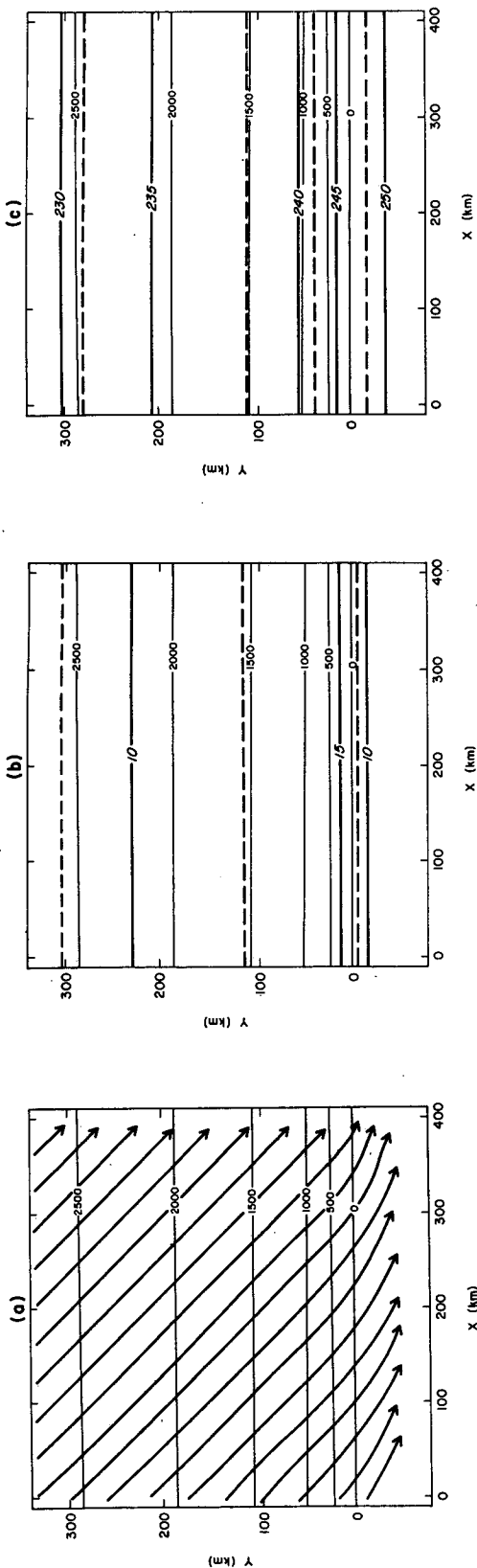


FIG. 4. Model streamline: (a) wind speed; (b) temperature (K) fields at the second σ -level (about 50 m above ground) after a 10 h time integration for the uniform ice profile. South is directed toward the top of each figure; east is to the left. Ice terrain height contours are denoted by thin solid lines.

layer into the katabatic layer. These processes tend to prevent significant temperature inversions from frequently developing in the coastal environment. This is in contrast to the interior site where mixing processes are not nearly as strong. As shown in Fig. 5, the temperature inversion in the interior is quite strong. The shape of the profile suggests that the model interior site is transitional between the observed weak inversion profiles of the coastal environment and the well-documented exponential-shaped inversion profiles (Schwerdtfeger, 1970; Mahrt and Schwerdtfeger, 1970) found on the very gentle slopes of the Antarctic Plateau. Model simulations of temperature inversions also appear to be in qualitative agreement with mean wintertime inversion strengths over Antarctica as inferred by Phillipot and Zillman (1970).

b. Effect of lateral terrain variations on katabatic flow

In Section 2 it was argued that the intensity and persistence of katabatic winds at the coast depend on the cold air supply in the interior of the continent. The size of the cold air reservoir, in turn, is determined largely by the degree of confluence of the terrain-forced near-surface motion field. Regions of streamline confluence discussed by Parish (1982) are generally located east of major ice ridges in the Antarctic interior. In order to simulate such features, sinusoidal height variations having a wave-length about equal to the width of the horizontal domain have been superimposed on the parabolic ice dome. Mathematically, the modified model ice terrain can be expressed as

$$z(x, y) = 2800.0 \text{ m} \left(1 - \frac{y}{3.5 \times 10^5 \text{ m}} \right)^{1/2} + A \cos\left(\frac{2\pi}{L} x\right); \quad (13)$$

where A is the amplitude of the ice height variations and L the wavelength. Representative amplitudes can be inferred from topographic maps of Antarctica. Two cases will be presented in this section.

An amplitude factor of 250 m is used for the first test. The resulting model ice topography displays a significant wave-like feature, tending to disrupt the uniform drainage pattern displayed in Fig. 4a. As in the previous case of simple katabatic flow, the isothermal model atmosphere is initialized about a basic state of no motion. Equations were then integrated for a 10 h period. Differences in the nature of the slope flows resulting from terrain variations become obvious soon into the model run. The ice topography forces a marked near-surface flow confluence east of the ridge. In this area the cold air accumulates rapidly, forming a large supply pool of negatively-buoyant air available to feed downslope coastal regions.

The streamline, wind speed and temperature fields for the second σ -level after the 10 h integration period are shown in Fig. 6. All three fields attest to the fact that lateral ice terrain variations profoundly influence

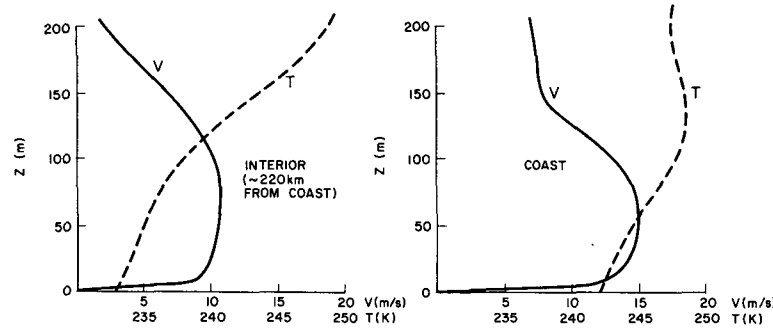


FIG. 5. Wind and temperature profiles over model ice terrain interior and coastal sections after 10 h time integration.

the cold air drainage patterns in the interior of the continent. Fig. 6a reveals the strong confluence of the streamlines east of the ridge and diffuence west of the ridge. Fig. 6b shows that the core of enhanced katabatic winds is found along the axis of the confluence zone. Furthermore, relatively weak katabatic flow is associated with the diffluent regions depicted in Fig. 6a, thus emphasizing the importance of the cold air supply. Compared to the simple katabatic wind example in the previous section, the coastal katabatic winds downslope of the confluence zone are nearly 60% stronger than corresponding katabatic winds over the uniform ice topography. Coastal katabatic wind speeds at the first σ -level (10 m) downslope of the confluence channel are nearly 18 m s^{-1} . This is in good agreement with the resultant winds along the Cape Denison-Port Martin coastal stretch.

The temperature field illustrated in Fig. 6c shows that the confluence zone is characterized by markedly colder temperatures than the surrounding environment. The collection of cold air in the region east of the ice ridge provides an ample source of negatively-buoyant air necessary for the strong coastal katabatic flow. Near the coast, however, the temperature contrast between the fast katabatic stream and other more tranquil sections of the model coastline is not nearly as well defined. Observations show this result is probably realistic. Loewe (1974) notes that the average temperatures at Cape Denison and Port Martin are not significantly different from other katabatic-prone stations about the rim of Antarctica.

The second test which incorporates lateral ice terrain variations is identical to the first test, except that the amplitude factor has been reduced to 125 m. The model terrain displays a less marked wave-like feature. However, the resulting katabatic winds after the 10 h model integration period are very similar to those from the first test. Final streamline, wind speed and temperature fields for the second σ -level are shown in Fig. 7. The terrain irregularities again significantly disrupt the drainage pattern in the lowest levels of the atmosphere. Fig. 7a shows a strong confluence of the streamlines east of the ridge, although the entire channel is oriented somewhat west of the position in Fig.

6a in response to the different terrain slope configuration. Wind speeds in Fig. 7b again show the enhanced katabatic winds along the core of the confluence channel. The magnitudes are only slightly less than those from the first test (Fig. 6b) and considerably stronger than those produced over the uniform terrain of the previous section. The temperature field in Fig. 7c again illustrates the accumulation of radiatively cooled air in the confluence zone which, in turn, provides the large supply necessary for flow continuity. These results suggest that even minor wave-like features in the interior ice terrain of Antarctica may lead to significant variations in katabatic winds.

c. Effect of synoptic forcing

Resultant wind directions leave no doubt as to the central importance of terrain slopes on the generation of katabatic winds. However, Stretten (1968) notes that the synoptic component can be of local importance in long duration katabatic events near the coast of Antarctica. To infer the role of pressure gradients in the free atmosphere on katabatic winds, a series of numerical experiments have been performed under a variety of upper-level conditions. In this section the results of one such experiment are presented although the discussion is generalized to incorporate findings from other experiments.

To start the model integration, a 10 m s^{-1} easterly geostrophic wind has been initialized at each level. Friction has not been accounted for in the initialization process, resulting in some inertia-gravity wave generation at the start of the model run. However, this noise becomes damped out considerably after the first few hours of model integration time. The same isothermal atmosphere is used as in all previous experiments. The nonuniform ice terrain profile expressed in (13) is used with a 250 m amplitude factor. Results of the model run at the 50 m level above ground after the 10-h integration period are shown in Fig. 8. The streamline pattern (Fig. 8a) is qualitatively very similar to those shown previously. The terrain remains the dominant forcing mechanism in shaping the near-surface confluence east of the ridge. However, the confluence zone is positioned slightly westward compared

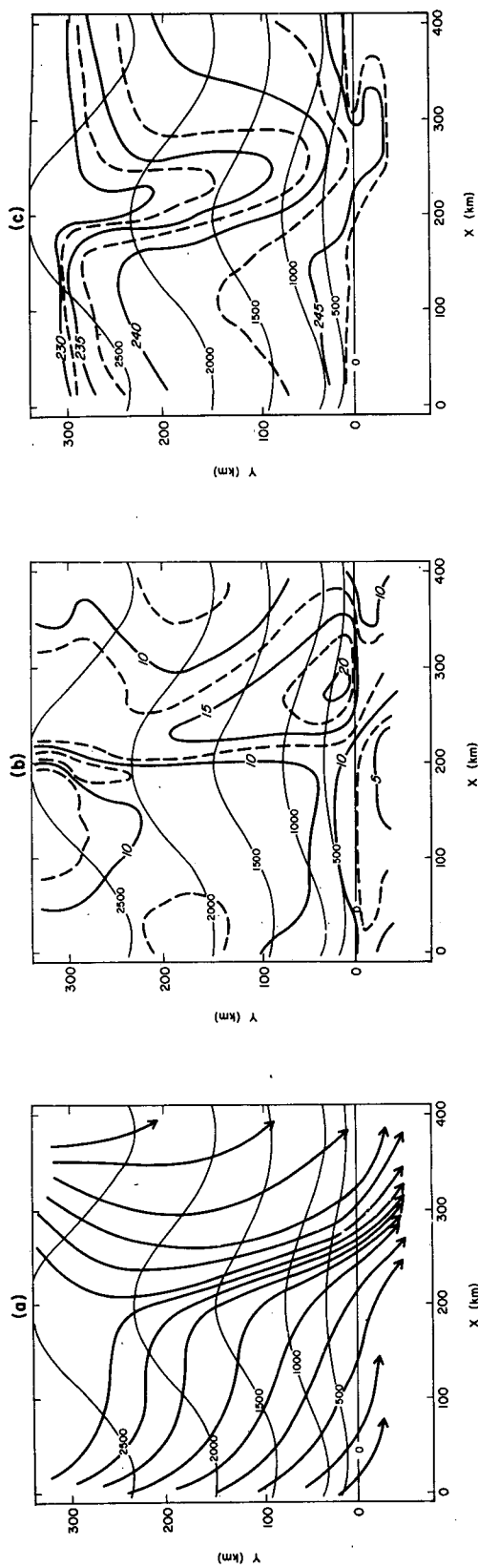


FIG. 6. As in Fig. 4, except for ice profile with 250 m amplitude lateral ice contour variation.

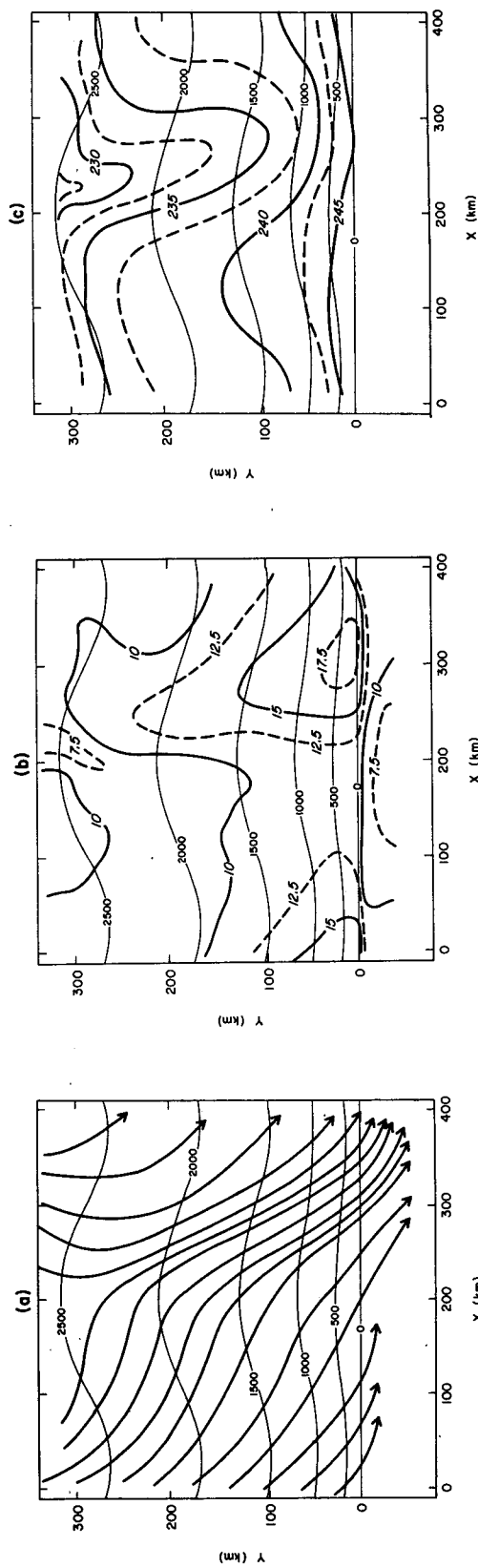


FIG. 7. As in Fig. 4, except for ice profile with 125 m amplitude lateral ice contour variation.

with the pattern in Fig. 6a. It appears that the synoptic pressure gradient acts to shift the entire pattern without significantly disrupting the terrain-forced cold air drainage pattern. This feature is reflected in the katabatic wind speeds of Fig. 8b. The core of strongest winds again is oriented along the confluence channel and is situated west of the position shown in Fig. 6b. Introduction of an upper level pressure gradient has not significantly altered the katabatic wind magnitudes although slightly greater wind speeds occur in Fig. 8b, possibly in response to the lower pressure over the ocean. The temperature field in Fig. 8c is similar to earlier results as well. The coldest temperatures are again found along the confluence zone, shifted from Fig. 6c in response to the upper level forcing.

Other experiments have been conducted under different magnitudes and directions of the synoptic pressure gradient force. With the exception of a few cases, the results of such experiments are consistent and show synoptic-scale forcing to be usually only of secondary importance. The synoptic pressure gradients supporting easterly or westerly geostrophic winds act mostly to shift the terrain-forced confluence zone a certain distance, depending on the magnitude of the upper-level forcing. Changes in katabatic wind speeds under such conditions are generally small. For cases in which the synoptic pressure gradient supports northerly or southerly geostrophic winds, the confluence patterns are not significantly shifted. However, katabatic wind speeds are modified, enhancement of katabatic flow occurring with southerly geostrophic flow. Such results are in qualitative agreement with observations presented in Ball (1960). The only exception occurs with anomalously strong pressure gradients in the free atmosphere which overwhelm even the terrain-induced accelerations. Such cases must occur infrequently in Antarctica since the mean-yearly directional wind constancy values for katabatic-prone coastal stations are usually 0.8–0.9.

5. Conclusions

Ample observational evidence exists to show that katabatic winds of the coastal environment of Antarctica are forced primarily by radiational cooling of the sloping ice terrain. Although katabatic winds occur frequently at most coastal stations exposed to drainage off the continental interior, the intensity and duration of such flows vary considerably from site to site. Regions such as the Adelie Land section, encompassing Cape Denison and Port Martin, and Terra Nova Bay experience katabatic winds unequaled in strength and persistence anywhere along the explored coastal perimeter of Antarctica. The most important factor governing the occurrence of such anomalous katabatic episodes appears to be the supply of radiatively cooled air in the interior of the continent. Large pools of negatively-buoyant air are required to maintain continuity of the downslope component of the wind. Over most of the coastal rim, katabatic flow tends to be of

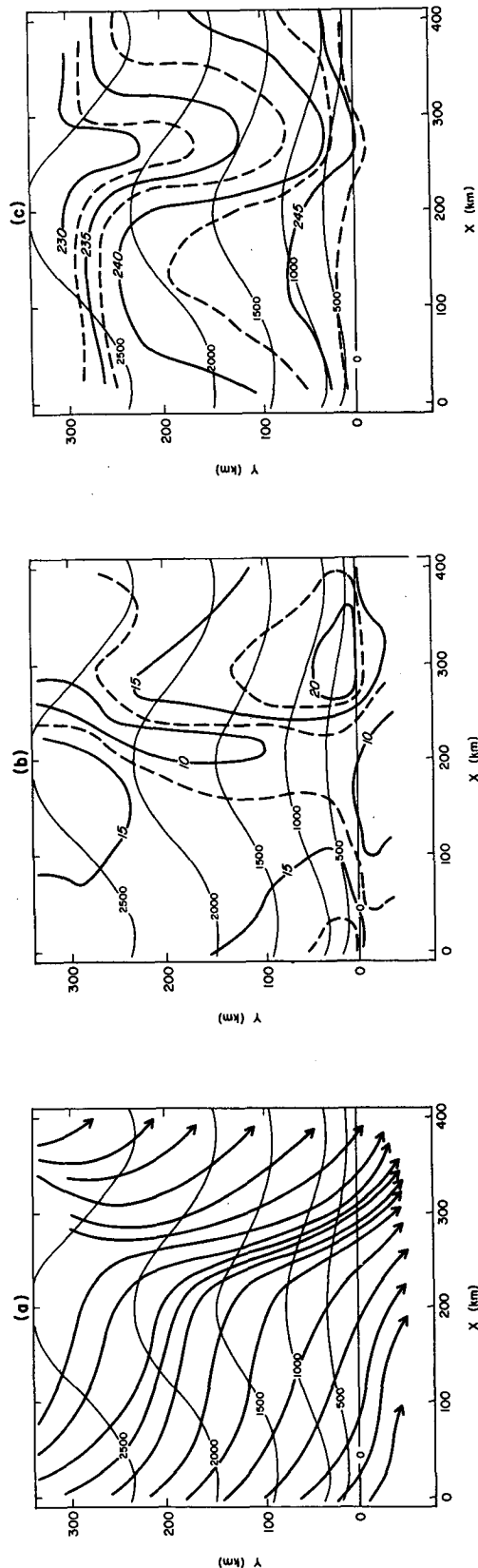


FIG. 8. As in Fig. 4, except for uniform 10 m s^{-1} easterly geostrophic wind initialized at all levels over ice profile with 250 m amplitude contour variation.

a sporadic nature owing to limited cold air supplies in the upslope interior. However, large supply pools form where near-surface drainage currents from an extended area over the interior pass through a constricted section of coastline. The cold air confluence is forced by wave-like irregularities in the ice terrain over the Antarctic hinterland and usually is located east of the ridge features. Clear examples can be seen upslope of the Cape Denison-Port Martin and Terra Nova Bay regions in Fig. 2.

Numerical simulations of katabatic winds have been carried out using a three-dimensional, primitive equation model. The model Antarctic ice topography has been specified by a parabolic equation; it is closely representative of the actual terrain of the coastal environment. A simple surface energy budget has been incorporated into the planetary boundary layer parameterization scheme to account for the radiational cooling of the air over the ice surface. The model is able to resolve the general shape of the katabatic wind and temperature profiles at locations near the coast as well as higher up on the more gently sloping interior terrain. Furthermore, katabatic wind speeds are in fair agreement with mean resultant winds from stations in coastal Antarctica.

Introduction of wave-like features in the model ice terrain acts to disrupt significantly the uniform near-surface cold air drainage pattern, resulting in a confluence of streamlines east of the ridge. The enhanced cold air reservoir provides a large supply of negatively-buoyant air, enabling katabatic winds to achieve substantially higher speeds downslope of the confluence zone and comparable to observed mean katabatic winds at Cape Denison and Port Martin. Numerical results suggest that even relatively minor terrain irregularities may have a significant impact on the katabatic wind regime.

Model results also suggest that the role of synoptic forcing is probably only of secondary importance in shaping the drainage pattern in the continental interior. Pressure gradients in the free atmosphere tend to shift the position of the confluence zone and/or modify katabatic wind magnitudes. This may be of some local importance along the fringes of the confluence zone, since the zone of strong katabatic winds may change by 50 km or so depending on the intensity and direction of the large-scale forcing. However, the terrain-forced accelerations which shape the streamline confluence patterns remain the primary forcing mechanisms for katabatic flow.

Such numerical results support the conceptual framework that katabatic winds over the coastal environment of Antarctica are profoundly influenced by the availability of negatively-buoyant air in the interior of the continent. Without a large supply of cold air from which to feed, katabatic winds must be restricted to the short-lived variety. The one fundamental reason for the existence of anomalously strong katabatic events at stations such as Cape Denison and Port Martin

appears to be the enhanced cold air supply upslope from the coast made possible by the topographically-forced confluence of mean surface drainage currents over the interior of Antarctica.

Acknowledgments. The author wishes to thank W. Schwerdtfeger for the numerous discussions regarding Antarctic katabatic winds. This work was supported by the National Science Foundation, under Grant DPP-8115976.

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