Observational Study of the Atmospheric Boundary Layer over Antarctica

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ABSTRACT

During the austral summer of 1982/83, measurements of wind and temperature profiles were made through the atmospheric boundary layer in Adelie Land, East Antarctica, an area known for strong katabatic winds. It was found that a shallow but strong temperature inversion was developed at night, and destroyed during the day, resulting in the development of a well-mixed layer. Wind hodographs were quite regular and spiral-like at night, but irregular during the day. The mean wind direction was about 40° to the left, looking downslope, but more downslope at night and more cross-slope during the day.

The conclusion was derived that during the polar summer the flow over Antarctica is controlled by the gravitational factor (slope-induced baroclinicity), by the thermal stability (turbulent mixing), and also by the synoptic forcing.

1. Introduction

No other single phenomenon has such a strong influence on the climate of a whole continent as the katabatic winds have on Antarctica. Although flows over slopes of Antarctica have been investigated for many years (e.g., Ball, 1957; Mather and Miller, 1967a,b; Schwerdtfeger, 1970; Radok, 1973), the past studies have been restricted by the lack of observational data, especially on the vertical structure of katabatic winds.

The overall picture of katabatic flow, existing in the literature, was mainly obtained from observations performed at a few inland stations in Antarctica. A number of studies employed series of wind soundings at Byrd, South Pole and Vostok (e.g., Lettau and Schwerdtfeger, 1967; Artemyev, 1972). Some investigations used data from Plateau Station with a well-instrumented, 32-m tower (Lettau et al., 1977; Kuhn et al., 1977; Riordan, 1977). In Adelie Land, Delaunay and Poggi (1981) conducted measurements, using a 70-m tower at Dumont d’Urville (an island a few kilometers offshore of the continent), along with 20-m towers on the ice slopes of Adelie Land.

Out of this latter study, with the goal of obtaining a better understanding of the katabatic wind, a joint experiment was initiated between the Geophysical Institute (University of Alaska), the University of Grenoble, and later by Centre de Recherches en Physique de l’Atmosphère (Poggi et al., 1982; Wendler et al., 1983). As a part of the program, in austral summer, January 1983, measurements of the wind and temperature profiles were carried out at different sites along the slope of Antarctica. The observations were supported by simultaneous measurements from five automatic weather stations (AWS), built by Stanford University and serviced by the University of Wisconsin. They were deployed previously in Adelie Land.

Our boundary layer measurements were carried out at three inland sites, where AWS sites D47, D57, and D80 are located. At the same time, the French (Seigneurier et al., 1984) were performing measurements at D10, about 5 km from the ocean (Fig. 1).

The present paper discusses only the boundary layer data collected at the inland sites during the experiment. The French observations (at D10 and Dumont d’Urville) performed in the vicinity of the ocean, where the flow is strongly influenced by local features (Gosink, 1982), are not considered in our study.

2. Area and instrumentation

The geographic and climatic characteristics of the sites can be seen from Table 1. All sites were, of course, steadily covered by snow, and the topography was very uniform. In the area of D57 and D47, where the steepness of the slope increases somewhat relative to D80, well-developed sastrugi fields were found (Wendler and Kodama, 1984).

The measurements were done with an Air-Sonde system, which was developed by Atmospheric Instrumentation Research, Inc. in Boulder, Colorado. Either a kite or a balloon was used as a carrier for the Air-Sonde. The balloons were used when the wind speed was in excess of 12 m s⁻¹, because for higher wind speeds a kite could be lost. The data were telemetered

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to the ground station, and recorded on a magnetic tape. About every fourth dataset was printed out by our HP 97 calculator, so that a quality check could be made of the data while in the field. For 24-hour periods in
January 1983, datasets were taken most of the time in about 6-hour intervals for each of the three stations (D47, D57, D80). Vertical soundings at these sites were not simultaneous.

The most southerly station, D80, had continuous daylight, while there was about 20 hours of daylight at D47, the most northerly station (Table 1). It should be noted that there was a prevalent, marked diurnal variation in temperature, even at D80, due to the sun elevation changes during the course of the day.

The summer (November–January) mean temperatures at the height of 3 m from the snow surface decrease when going up the slope: −17°C for D47, −20°C for D57, and −32°C for D80.

Opposite to this, the mean wind speed at the same level as those of temperatures increases when going down the slope. We found that the mean wind speeds at this level in summer are: 11.3 m s⁻¹ at D47, 9.6 m s⁻¹ at D57, and 5.5 m s⁻¹ at D80.

The wind directional constancy, which is defined as the ratio of the magnitude of the mean wind vector to the mean wind velocity (when wind blows only from one direction, the wind constancy is equal to one; when magnitude and frequency of winds from all directions are the same, the wind constancy is zero), is quite high for all stations, with monthly values of about 0.9.

The increasing wind speed, when traveling down the slope of the Antarctic continent, keeps the equivalent chill factor (Dalrymple and Stroschein, 1966) fairly constant for the three stations, and the values in January were found to be 1500–1700. This corresponds to the chill temperature of about −50°C. These unpleasant environmental conditions, together with the drifting and blowing snow, which was experienced during about half of the time of the expedition, made measurements difficult to perform.

A summary of the collected data is given in Tables
TABLE 2. The summary of the boundary layer measurement at D80, D57, and D47 in January 1983.

<table>
<thead>
<tr>
<th>Run</th>
<th>Location</th>
<th>Day in January</th>
<th>Time</th>
<th>Solar elevation (deg)</th>
<th>Carrier</th>
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<td>12</td>
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</tr>
<tr>
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<td>1111</td>
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<td>13.8</td>
<td>kite</td>
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<tr>
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</table>

2 and 3. Seventeen vertical profiles of the wind and temperatures were obtained (9 profiles for D80, 2 for D57, and 6 for D47). All measured profiles were smoothed using a 3-point averaging operator (sometimes used more than once) to filter obtained scatter, which could not be explained physically.

Table 4 provides the additional information on weather conditions during the experiment.

3. Analysis of the temperature profiles

Our analysis is based on the vertical profiles of the temperature and wind velocity, since the direct measurements of turbulent fluxes were not available. We will start from a discussion of the time variations of the temperature field in the atmospheric boundary layer (ABL).

The observations, performed approximately every six hours at every site, enable us to observe the diurnal cycle of the potential temperature profiles. The longest series of nine profiles was obtained at D80 (Fig. 2). The general picture emerging from Fig. 2 is that the measured potential temperature profiles do not differ from those observed in lower latitudes, where a mixed layer of uniform potential temperature with height occurs during the day and a temperature inversion is observed at night (Arya, 1982). In January at D80 the day (defined as the period with a positive net radiation balance) lasts about 15 hours (see Fig. 6). Figure 2 shows that during daytime hours, a mixed layer with a height of about 200–400 m develops. This indicates a mean entrainment velocity of about \( w_e = 5.6 \times 10^{-3} \) m s\(^{-1}\). This value is one order of magnitude smaller than the typical values observed in lower latitudes and proves that the observed mixed layer is quasi steady state. Based on laboratory measurements and aircraft data, Deardorff and Willis (1974) estimated the entrainment rate \( w_e \) of about 2% of the convective velocity scale \( w_* \). This gives a low value of \( w_* \sim 0.3 \) m s\(^{-1}\) in our case.

Above the mixed layer a 50–100 m thick transition layer borders with the free atmosphere. The potential temperature structure of the free atmosphere is characterized by a gradient of about \( 10^{-2} \) K m\(^{-1}\). The strongly stratified superadiabatic layer underneath, several tens of meters deep, is a transition between the mixed layer and several degrees warmer snow surface.

Figure 2 also shows that at night, a shallow 100–200 m inversion layer rapidly develops above the surface. Its growth does not destroy the mixed layer above, which exists for the whole night and therefore can be called "nocturnal mixed layer." Radiative cooling decreases the temperature near the ground by the order of 10 K at the height of 1.5 m—the lowest level of observations (Table 3). As a result, a negative curvature of the potential temperature profile is usually observed. Only one case of positive curvature of the potential temperature profile was observed in Run 14, which is displayed in Fig. 3. According to Andre and Mahrt (1982), this indicates a relative importance of the turbulent mixing versus radiative cooling in this case.

Analysis of the potential temperature profiles enables us to determine the height of the daytime and nighttime boundary layers. The top of the stable boundary layer \( h \) was defined by the level at which the potential temperature gradient becomes constant with height. The

TABLE 3. General characteristics of the boundary layer in Adelie Land, East Antarctica.

<table>
<thead>
<tr>
<th>Run</th>
<th>Time</th>
<th>( z_i )</th>
<th>( h )</th>
<th>( \theta_0 - \theta_i )</th>
<th>( V_0 )</th>
<th>( V_\phi )</th>
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<td>**</td>
<td>**</td>
<td>**</td>
<td>6.5</td>
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<td>2216</td>
<td>—</td>
<td>—</td>
<td>**</td>
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<td>**</td>
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</tr>
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<td>7.4</td>
</tr>
<tr>
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<td>—</td>
<td>-1.0</td>
<td>7.3</td>
<td>9.5</td>
</tr>
</tbody>
</table>

* \( z_i, h \): heights of daytime and nighttime ABL.
† \( \theta_0 - \theta_i \): temperature difference between top and the bottom (1.5 m) of ABL.
‡ \( V_0, V_\phi \): wind at \( z = 1.5 \) m and at the top of ABL.
**: sounding did not reach to the top of the ABL.
TABLE 4. The information about cloud conditions, radiative balances, and solar elevations for the days of the measurements in Adelie Land, East Antarctica.

<table>
<thead>
<tr>
<th>Day in January</th>
<th>Time</th>
<th>Cloud amount (/10)</th>
<th>Cloud type</th>
<th>Time</th>
<th>Net short (W m⁻²)</th>
<th>Net all (W m⁻²)</th>
<th>Solar elevation (deg)</th>
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<td>2240</td>
<td>—</td>
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<td>94.2</td>
<td>27.7</td>
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<tr>
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<td>1015</td>
<td>10</td>
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<td>81.6</td>
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<td>1730</td>
<td>80.2</td>
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<td>24.0</td>
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</table>

The top of the mixed layer z_i was defined as the height at which the potential temperature starts to increase with height. The values of the boundary layer heights were found "by eye." The definition and determination of the mixed layer height seems reasonably clear. The definition of the stable boundary layer is not unique, but one of many that are possible (Arya, 1981; Wentzel, 1982).

In Figs. 4 and 5, potential temperature profiles from all three sites, are grouped in two stability classes. Figure 4 shows the daytime profiles obtained from observations at D80 and D47. This figure indicates three distinct regions of the daytime ABL: superadiabatic surface layer, mixed layer, and above it, the transition layer to the free atmosphere. In Fig. 5, dimensionless nocturnal profiles are presented. The scaling factor (θ_a - θ_i) is the difference between potential temperature θ_a at the 1.5 m level and potential temperature θ_i at the top of the stable boundary layer. It is important to notice that this difference at night has much higher absolute values than during the day (Table 3).

Figure 5 indicates that dimensionless, nocturnal potential temperature profiles can be expressed in the simple exponential form (1 - z/h)*, where α is a power coefficient. This agrees with the results of Yamada (1979), who obtained the value of α = 3. The cubic model of Yamada approximates our Run 5 profile quite well. However, Fig. 5 shows that the power α changes over a wide range of values. This is caused by nonstationarity due to the short night duration. Results from the radiative model of Cerni and Parish (1984) suggest that steady state in the inversion layer can be reached in a day or so.

A simple exponential formula does not describe the case of the transition profile (Run 7). In this case, turbulence becomes strong enough to occupy the lower part of the inversion layer. The negative curvature of the temperature profile disappears and an isothermal layer is formed above the surface.

Analysis of the Richardson number profiles (not

![Fig. 2. Potential temperature profiles at D80.](image)
shown), which have irregular sharp peaks at different heights, indicates that all nighttime cases belong to the stable-sporadic class of turbulence (Arya, 1982). The layer of continuous turbulence (Ri < 0.25) occurs near the surface only. The stable-continuous regime, defined as Ri < 0.25, in the whole boundary layer, is observed only in Run 17 (shown in Fig. 3), which is a transition between daytime and nighttime conditions. This case can also be classified as neutral, since a mixed layer occupies all the ABL, except for a shallow (25 m) surface inversion sublayer.

4. Analysis of the wind field

Surface winds in Antarctica blow from the high plateau to the edge of the continent, crossing contour lines at an angle of approximately 45° to the left of the fall line (Mather and Miller, 1967). A typical example is shown in Fig. 6 (D80, 13–14 January). The slope direction is shown in the left corner of the figure. From Fig. 6 it follows that the surface wind vector becomes more downslope at night. This effect can also be seen from Fig. 7, which displays a time variation of the mean temperature along with the mean wind speed and direction at D80 (average of the last 17 days of January). The diurnal amplitude of the wind direction in Fig. 6 is 27°, and 16° in Fig. 7. The mean wind in Fig. 6 is oriented about 40° to the left of the fall line. In Fig. 6 the diurnal changes of the potential temperature and net radiation are also shown.

Changes of the wind direction toward the maximum
The total geostrophic wind $\mathbf{G}$ is expressed in the form:

$$ \mathbf{G} = \mathbf{G}_s + [0, v_T], \quad (1) $$

where $\mathbf{G}_s$ is the synoptic geostrophic wind vector. Since $v_T$ is positive during the day and negative at night, the total geostrophic wind vector, at the height $z$, turns to the left or to the right with respect to the vector $\mathbf{G}_s$.

In the case of the strong temperature inversion (and weak turbulent mixing), the negative thermal wind $v_T$ is able to keep the surface wind vector slightly to the left, with respect to the slope vector, for all possible orientations of the flow on the top of the friction layer (Mahrt and Schwerdtfeger, 1970). This explains the persistency of the wind observed during the winter over Antarctica. However, this mechanism does not apply in the summertime.

During the summer, thermal stability and turbulent mixing play more important roles, controlling the angle between the surface and geostrophic wind. In the well-mixed layer during the day, this angle is small, since the wind vector is almost uniform, and close to the geostrophic wind vector. It is easy to find that $v_T \sim \theta'$ (for $\beta = 0.03$, $\psi = 3 \times 10^{-3}$ and $f = 10^{-4}$). Since $\theta' \sim 10$ K during the night and $\theta' \sim 1$ K during the day, it follows that the gravitational factor is important during the night, but is negligible during daytime, since it is only about 0.1 of the geostrophic wind.

The diurnal variation of the wind spirals at D80 are shown in Fig. 8. There, and on all remaining figures, the x-axis of the coordinate system is oriented down slope. The nocturnal hodographs in the figure have shapes of spirals, with wind vectors turning to the left with height. The surface wind at night is more downslope and weaker than during the day. The daytime hodographs of Runs 4 and 9 indicate that the surface wind velocity is turned more cross-slope, as an effect of stability. Close to the top of the ABL, the hodographs unexpectedly turn to the left and decrease the $\mu$-com-
Component. This can be explained as an effect of a synoptic scale baroclinicity or of secondary flows (Brown, 1981).

Characteristic features of the daytime and nighttime wind profiles are shown in Fig. 9. In this figure, the components of the total geostrophic wind, obtained from Eq. (2) are also plotted. We assumed that \(G_r\) is equal to the wind vector on the top of the boundary layer and is constant for the whole layer. The thermal wind is only due to the slope. The synoptic thermal wind components could not be computed. A similar procedure was previously adopted by Kuhn et al. (1977). In Fig. 9, a well-mixed, daytime profile of the wind components (Run 3) is shown to be quite uniform. The components of the nocturnal wind velocity vector increase faster with height and possess characteristic maxima.

In Fig. 10, vertical wind profiles, classified into three stability regions, are presented. Profiles of the stable class (Fig. 10a) have maxima, which, however, are not pronounced and occur at different heights \((z/h)\). The “transition” profiles (Fig. 10b) are quite regular and can be approximated by an exponential formula of the type \((z/h)^n\). The profiles for the convective class are similar to the “transition” profiles.

5. Some theoretical and modeling considerations

In this section several conclusions will be derived by a comparison of our measurements with a simple solution of the momentum equations listed in the Appendix. These equations qualitatively express characteristic features of the ABL over Antarctica. The flow, described by these equations, is based on a balance of the Coriolis, frictional, and pressure forces (with components generated by a slope and by synoptic baroclinicity), resulting in a nonzero wind acceleration. Simple time- and height-dependent parameterizations are developed for the eddy diffusivity and the slope-induced thermal wind \(v_T\). The surface values of \(v_T\) are allowed to vary as a sine function of time, which is positive during the day and negative at night.
The wind hodographs obtained from the 24-hour simulation of the model are shown in Fig. 11. Based on the model results, we are able to derive the following conclusions:

1) The simulated wind vector does not change its direction during the day and is parallel to the geostrophic wind on the top of the ABL. This can explain the uniform shape of the daytime hodographs in Fig. 8.

2) The simulated nighttime hodographs in Fig. 11 have a spiral shape. This also agrees with the shape of the hodographs plotted in Fig. 8. The strongest “low-level jet” is obtained about 21 hours after the beginning of the simulation, equivalent to 0300 LST. This cannot be compared with our data, since the observations at this time were not performed during our experiment. This could explain, though, why the wind maxima at night, presented in Fig. 10a, are not very pronounced. The simulated surface wind at night is further downslope as a simultaneous effect of the gravity (thermal wind $v_T$) and stability factors (increase of the cross-isobar angle). This again is in agreement with Fig. 8.

3) The results of the simulation depicted in Fig. 11 were obtained with the assumption that the synoptic thermal wind is zero. We also performed a numerical experiment (not shown) with the nonzero synoptic thermal wind of the magnitude 5 m s$^{-1}$/km in a 200 m-deep layer (which is equivalent to the horizontal temperature gradient about 1.5 K/100 km). We found that synoptic baroclinicity modifies the wind hodographs very slightly in the shallow ABL. From this we conclude that the irregularities of the daytime hodographs on the top of the ABL, seen in Fig. 8, should be explained as an effect of the secondary circulation, rather than synoptic baroclinicity. The model results presented by Brown (1981) are quite similar to our daytime hodographs in Fig. 8. Brown argues that the presence of the secondary circulation is ensured if the surface velocity exceeds 5 m s$^{-1}$ and the stratification is weakly unstable.

4) The results of the steady-state version of the model, shown in Fig. 12, indicate that the explanation of the wind directional constancy with height by the
mechanism described by Mahrt and Schwerdtfeger (1970) is satisfactory during the winter time. This explanation fails, however, during the summer when more intensive turbulent mixing and positive thermal wind can occur during the day. Therefore, only the height persistency in the direction of the geostrophic flow (synoptic forcing) can explain the constancy of the surface winds during the polar summer.

6. Conclusions

In this paper vertical profiles of the temperature and the wind vector over Antarctica were presented and discussed. The temperature profiles, characteristic for the polar summer, have shown a well-developed mixed layer during the day and a strong inversion layer at night. It was deduced that due to the duration of the day and night, the daytime ABL is quasi steady state, with an entrainment rate of \( \nu_{e} \sim 5.6 \times 10^{-3} \text{ m s}^{-1} \), and the nocturnal ABL is shallow and nonsteady.

During the day, when the surface inversion had vanished the flow was downslope, due to the weak intensity of the surface radiational heating and synoptically driven circulation toward the edge of the continent.

During the night, regular spiral-like wind hodographs were observed. In contrast, the daytime hodographs did not have a spiral shape and were found to be very irregular on the top of the ABL.

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APPENDIX

A Simple Numerical Model of Flows over Antarctica

This simple model is based on the following set of equations, which describes the distribution of the wind velocity in the nonsteady ABL over a slightly inclined terrain (the nonsteady version of the equations derived by Sorbjan, 1983, and written for the Southern Hemisphere).

\[
\begin{align*}
\frac{\partial u}{\partial t} + f(v - v_{e0}) - fz \frac{\partial v}{\partial z} + f w_{T} &= \frac{\partial}{\partial z} \left( k \frac{\partial u}{\partial z} \right) \\
\frac{\partial v}{\partial t} + f(u - u_{e0}) - fz \frac{\partial u}{\partial z} &= - \frac{\partial}{\partial z} \left( k \frac{\partial v}{\partial z} \right)
\end{align*}
\]  \hspace{1cm} (A1)

where

\begin{itemize}
  \item \( u, v \), \( u_{e} \), \( v_{e} \) \( u_{e0}, v_{e0} \) \( f \) \( k \) \( \nu_{T} = \beta \theta' \psi / f \)
\end{itemize}

components of the wind and the geostrophic wind vectors at an arbitrary height and the surface;

\begin{itemize}
  \item Coriolis parameter,
  \item eddy diffusivity,
  \item slope-generated thermal wind.
\end{itemize}

In Eq. (A1), it is assumed that the geostrophic wind can vary with height (synoptic baroclinicity).

Our model is designed to describe qualitatively the time variation of the wind in the ABL. This is achieved by adopting a very simple and somewhat arbitrary closure of Eq. (A1), for the eddy diffusivity \( k \). Generally, the eddy diffusivity can be parameterized as follows:

\[
\begin{align*}
\kappa u_{e} z & (1 - z/h) / (1 + 4.7 \pi z/h) \quad \text{stable} \\
\kappa u_{e} z & (1 - z/h) \quad \text{neutral} \\
4k_{max} z & (1 - z/z_{c}) \quad \text{convective}
\end{align*}
\]  \hspace{1cm} (A2)

(Sorbjan, 1984) (Wyngaard, 1981),
where

- \( h, z_i \) are heights of the stable and convective ABL,
- \( k_{\text{max}} \) is the maximum value of \( k \) in the convective ABL,
- \( \kappa \) is the von Kármán constant,
- \( u_\infty \) is the friction velocity,
- \( \eta = h/L \) is the stability parameter, \( L \) being the Monin-Obukhov length.

We generalize Eq. (A2) assuming:

\[
k(Z, t) = \begin{cases} 
  k_{aw} (1 + 4.7\eta Z) & \text{for nighttime ABL} \\
  k_e & \text{for daytime ABL,} 
\end{cases}
\]  

(A3)

where \( k_{aw} = k_0 z (1 - Z) \) (neutral case, \( k_0 = 100 H \)), \( k_e = 4 k_{aw}/k_0 \), and \( Z = z/H \); \( H \) is the height of ABL. We assumed in Eqs. (A3) that \( H = 200 \text{ m} \), \( k_0 = 20 \text{ m}^2 \text{ s}^{-1} \).

The time forcing in the model is expressed by the function \( F \), which describes a diurnal variation of the eddy diffusivity and the thermal wind. We assumed \( k_{aw}(t) = [1 + cF(t)] \) with \( c = 20 \) and \( \eta(t) = -15 F(t) \). The function \( F(t) \) is defined as:

\[
F(t) = \begin{cases} 
  \sin(2\pi t/T_0) & \text{for } 0 < t < T_1/2, \text{ daytime} \\
  \sin\left(\pi \left(1 + \frac{t - T_1/2}{T_0 - T_1/2}\right)\right) & \text{for } T_1/2 < t < T_0, \text{ nighttime} 
\end{cases}
\]  

(A4)

where \( T_0 = 24 \text{ hours} \), \( T_1/2 = 14 \text{ hours} \).

The form of \( F(t) \) indicates that we assume the day and night to be 14 hours and 10 hours long, respectively. This is in agreement with the net radiation distribution shown in Fig. 8.

In Eq. (A1), we also assume that

\[
v_T = v_T(0)(e^{-Z} - e^{-1})/(1 - e^{-1})
\]  

(A5)

where \( v_T(0) = R\tilde{F}(t) \) and \( R \) is equal to 4.0 at night and 0.4 during the day. Notice that the thermal wind decreases exponentially from the value \( v_T(0) \) at the surface to zero at the top of the ABL. This can be compared with the profile of \( v_T \) shown in Fig. 10. The exponential shape of \( v_T \) was also assumed by Mahrt and Schwerdtfeger (1970). The other parameters in Eq. (A1) are chosen to be \( u_\infty = 5 \text{ m} \text{ s}^{-1} \), \( v_\infty = 5 \text{ m} \text{ s}^{-1} \), \( f = -1.2 \times 10^{-4} \text{ s}^{-1} \).

The boundary conditions are established as:

\[
\begin{align*}
  u &= 0 & & \text{for } z = 0 \\
  v &= 0 \\
  u &= u_\infty + \frac{\partial u}{\partial z} H & & \text{for } z = H, \\
  v &= v_\infty + \frac{\partial v}{\partial z} H 
\end{align*}
\]  

(A6)

As an initial condition, we use the steady state solution of Eq. (A1), with \( k = k_e \) and \( v_T(0) = 0 \). To solve Eq. (A1), we first transform it to the form:

\[
\frac{\partial W}{\partial t} = \frac{\partial}{\partial Z} k \frac{\partial}{\partial Z} W - f(v_T + i\Delta Z)e^{i\theta},
\]  

(A7)

where

\[
W = [(u - iv) - (u_\infty - iv_\infty)]e^{i\theta}
\]

\[
\Delta = \left[ \frac{\partial u}{\partial Z} - j \frac{\partial v}{\partial Z} \right],
\]

and then express Eq. (A7) in the finite difference form. The finite differences algebraic equations were solved by the "tridiagonal algorithm," described in Appendix A of Roache (1972).

REFERENCES


