The early evening boundary layer transition

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SUMMARY

Various mechanisms which lead to nocturnal accelerations and formation of the low-level jet are examined by analysing Wangara data. The stress field is inferred from the wind field and equations of motion. The evolution of the stress divergence, during the evening transition from daytime mixed layer flow to nocturnal boundary layer flow, is found to increase the ageostrophic flow and subsequent nocturnal accelerations by roughly a factor of two. During this transition the stress divergence in the lowest few hundred metres increases due to the fact that the influence of decreasing boundary layer depth exceeds the effect of decreasing surface stress. This leads to temporary deceleration and rotation of the low-level wind vector towards low pressure and thus increases the ageostrophic flow.

Diurnal variation of the geostrophic wind is also found to significantly strengthen the nocturnal flow. This diurnal variation is apparently due to heating and cooling over terrain which slopes gently upward, east of the location of the Wangara experiment.

1. INTRODUCTION

The analyses of Bonner et al. (1968), Wippermann (1973), Paegle and Rasch (1973), Sisterson and Frenzen (1978), Mahrt et al. (1979) and others indicate that the nocturnal low-level jet is a widespread phenomenon becoming a climatic feature in many regions. There appear to be a number of mechanisms which can lead to nocturnal accelerations. The inertial oscillation proposed by Blackadar (1957) and Buajitti and Blackadar (1957) should occur whenever the turbulence intensity and depth of the boundary layer vary diurnally. However, typical values of the daytime frictionally driven ageostrophic flow are far too small to account for the observed amplitude of the nocturnal inertial oscillations. The modelling of Zeman (1979) suggests that evolution of the stress field during the late afternoon decay of the daytime mixed-layer may rapidly increase the low-level ageostrophic flow.

Three additional mechanisms may at times be important or even dominate. First baroclinity associated with surface radiational cooling over sloping terrain can significantly accelerate low-level nocturnal flow (e.g. Lettau 1967). Paegle and Rasch (1973) and Zeman (1979) have suggested that over the Great Plains of the United States this mechanism may be as important or more important than the inertial mechanism. In fact, Caughey et al. (1979) have shown that even very small slopes, on a horizontal scale much smaller than the Great Plains, can significantly influence nocturnal flow.

A second additional mechanism is pre-existing daytime baroclinity. Baroclinity can lead to changes of ageostrophic flow with height particularly in the case of well-mixed flow (Arya and Wyngaard 1975).

A third additional mechanism is ageostrophic flow driven by accelerations due to temporal variations of the synoptic scale geostrophic wind (Young 1973, Mahrt 1974). For the majority of orientations of the geostrophic wind tendency with respect to the existing geostrophic wind, accelerations increase the ageostrophic wind. Advection accelerations may also significantly increase the ageostrophic flow (Paegle and Rasch 1973, Mahrt 1975, Hasse 1976). However, analysis of such effects from actual data requires a special observational analysis (Augstein and Heinricy 1976).

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In actual flows, the resulting ageostrophic flow and nocturnal accelerations depend on the interaction of the above additional mechanisms and the evolution of the turbulence during the transition from daytime mixed-layer flow to nocturnal boundary-layer flow. The change of the stress and ageostrophic wind fields during the transitional period have not been previously studied observationally in any detail. One problem is that turbulent fluxes may be particularly difficult to measure due to non-stationarity during the transitional period. For the same reason, the validity of existing parametrizations of turbulent fluxes is uncertain. Thus, it is not surprising that in modelling studies, the transitional period is normally omitted or replaced with an instantaneous collapse (Thorpe and Guymet 1977). Consequently, there is strong motivation to analyse the observed wind field of the evening transitional period even if adequate observations of turbulent quantities are not available. In the present study we analyse wind profiles, primarily from the Wangara experiment conducted near Hay, New South Wales (Clarke et al. 1971), with the purpose of assessing the importance of the evening transition and other influences on low-level nocturnal accelerations.

2. **Mean flow characteristics**

Frequently, winds in the lowest 50 or 100 m begin to decelerate about two hours before sunset as is evident in averaged conditions during the Wangara experiment (Fig. 1). Sunset ranged from approximately 1745h to 1800h (local time). During this period, winds, first at higher levels, begin to accelerate. Eventually, accelerations occur as low as 10 m above the ground. Wind speeds averaged over a one-year period at the KTVY tower in Oklahoma

![Graph](image)

**Figure 1.** Averaged diurnal variation of windspeed for the Wangara days. Wind levels are labelled in metres. The vertical hatches denote sunset.
Figure 2. Diurnal variation of windspeed during August at Risø, Denmark, averaged over a 10-year period (Petersen 1973). Wind levels are labelled in metres. Sunset ranged from 1940 to 2040 (times are Central European Time).

City (Crawford and Hudson 1970) exhibit similar trends. Such diurnal modulation is even evident in averaged low-level data at high-latitude maritime locations (Fig. 2). On many individual nights at the above locations, the low-level winds attain speeds several times greater than the daytime values.

To analyse Wangara data further, we will use the 'fair weather' Wangara days which were selected by Melgarejo and Deardorff (1974) to evaluate the 1500h surface stress and mixed-layer depth. Their selection procedure required no frontal activity within 500 km of the centre of the Wangara mesoscale network and less than 1/8 low-level cloud cover. However due to some variable high-level cloudiness, such days are characterized by a variety of surface cooling rates. We shall refer to such days as 'fair weather days'. Calculations were also made for the remaining Wangara days. However, since the trends were on average the same, although weaker, we will normally report only averaged results for the nine fair weather days.

In the following discussion we will assume that, on average, the mixed-layer flow at 1500h is oriented roughly in the direction of the geostrophic wind with a lesser perpendicular component usually directed toward low pressure. On Wangara fair weather days the boundary layer winds are observed to rotate significantly toward low pressure beginning two or three hours before sunset as is evident from averaged rotation rates in Fig. 3. The rate of this rotation reaches a maximum around sunset but then vanishes an hour or two after sunset. Computations in section 3 indicate that prior to the evening transition, the cross isobar angle between the actual and geostrophic winds is typically 10–15° due to daytime frictional effects. Figure 3 indicates that over the entire evening transition, this angle increases by more than 30° near the surface and by 10° or more at the usual level of the nocturnal jet (200–300 m).

Shortly after sunset the boundary layer wind vector above 400 m begins to rotate in the opposite direction (toward high pressure) as effects due to the nocturnal oscillation begin to dominate. About two hours later the rotation of the wind vector in the lower layers also begins to reverse direction. We will refer to this time, when the flow is rotating in the direction of the nocturnal inertial oscillation at all levels, as the termination of the evening transition. So defined, the evening transition, attendant rotation of the wind vector
toward low pressure and presumed increase in ageostrophic flow, appear to last 4 or 5 hours. On individual days this wind rotation always began by 1600h, while the termination time for individual nights averaged about 1945h.

The model of Delage (1974) also predicts an early evening maximum in the surface cross-isobar flow angle although the amplitude is small. This early evening maximum occurred on only two of the five nights examined by Caughey et al. (1979) although slope effects appeared to play an important role.

3. INFERENCE OF STRESS FIELD

Since appropriate flux measurements are not available for the evening transitional period, we will attempt to infer the qualitative behaviour of the stress divergence from the wind field and equation of motion. Even the most basic aspects of the evolution of the stress divergence during the evening transition are not known. Therefore we will attempt to infer only the basic features of the stress field and avoid uncertainties and loss of generality which would result from interpreting the data in terms of a specific model.

On the other hand, application of the usual residual method to estimate the stress divergence from the equation of motion will yield uncertain results because of the accumulation of errors resulting from the estimate of the geostrophic wind and the rapid variation of variables with time. For qualitative interpretation, we will avoid these difficulties by
integrating the equation of motion with time after assuming linear temporal variation of the residual terms. Higher order interpolation is not justified for present purposes.

To facilitate easy comparison with other studies, we pose the following interpretation in terms of a Northern Hemispheric framework in which case the equations of motion can be written as:

$$\frac{dw}{dt} = -ifw + i\omega \theta + R(t)$$

(1)

$$w = u + iv$$

$$w_\theta = u_\theta + iv_\theta$$

where $u$ and $v$ are the horizontal wind components in the $x$ and $y$ directions, respectively, $u_\theta$ and $v_\theta$ are the corresponding geostrophic wind components, $t$ is time, $f$ is the Coriolis parameter and $R(t)$ is the advection and stress divergence terms.

Integrating (1) with respect to time, we obtain
\[ w = e^{-if\tau} \int_0^\tau e^{if\tau'} \{ R(t) + ifw_y \} \, dt + w(0)e^{-if\tau} \]  \hspace{1cm} (2)

where \( w(0) \) is the complex horizontal velocity at \( t = 0 \), the suspected beginning of the transitional period. The influence of advections on the composit data, such as in Fig. 3(b), is probably small compared to the large and systematically varying local acceleration and stress divergence during the transitional period. In the following discussion of composit data, we will assume that \( R(t) \) is dominated by the stress divergence.

Assuming that variations of the geostrophic wind are negligible, assuming that the flow is initially in balance (\( dw/dt = 0 \) at \( t = 0 \)) and approximating the stress divergence with linear time dependence

\[ R(0) = if\{w(0) - w_y\} \]

\[ R(t) = R(0) + Bt, \]

the solution for the actual wind vector becomes:

\[ w = w(0) + (Bif^2)[1 - \cos(ft) + i\sin(ft) - ft] \]  \hspace{1cm} (3)

The last term in the brackets is the time-dependent part of the flow which is in balance with the changing stress divergence. The first three terms in the brackets represent inertial effects due to an imbalance of forces generated by the changing stress divergence.

To interpret (3) further, we rotate the coordinate system so that the positive real axis is aligned with the initial wind vector. Since the duration of the transitional period is small compared to the inertial period (\( ft \ll 1 \)), (3) becomes approximately

\[ w \approx w(0) + \frac{1}{2}Bt^2 - i(B/6)f^3 \]  \hspace{1cm} (4)

For purposes of discussion we will assume that the stress divergence vector is roughly opposite to the wind vector. Then increasing magnitude of the stress divergence (\( B \) is real negative) produces a flow modification which is initially directed toward low pressure and rotates away from the initial wind vector. This causes the total wind vector to initially decelerate and rotate toward low pressure as is observed in the lowest few hundred metres during the transitional period of the Wangara experiment (Fig. 3). Thus, the influence of decreasing depth scale on the stress divergence in the lowest few hundred metres may be greater than the influence of decreasing surface stress so that this stress divergence increases.

The rapid decrease of the depth of significant stress may be related to quick extension of cooling above the surface layer into flow with weak shear. As a result, the gradient Richardson number becomes very large around 100–200 m above the ground, implying local reduction or elimination of downward momentum flux at these levels. In fact the profile of the Richardson number computed over 50 m layers at 1800h exhibits a significant maximum occurring somewhere between 50 m to 250 m on all of the fair weather days. This maximum was always greater than 1 and was greater than 3 on all but two evenings. At the same time, the Richardson number in the lowest 50 m was less than 0.25 on all but two evenings and always less than 1. Above 250 m, the Richardson number decreased again to values less than 0.5. Such low Richardson numbers usually extended upward to near the inversion corresponding to the top of the daytime mixed layer. On most of the nights, this vertical structure for the Richardson number continued throughout the night.

Of particular interest here is that below the layer of large Richardson number, the wind decelerates and rotates toward low pressure corresponding to an increase in ageostrophic flow. In the layer of large Richardson number, the wind also rotates toward low pressure between 1500h and 1900h as is evident in Fig. 3(b) for levels between 100 m and 300 m.
However, the wind speed at these levels is at first relatively constant and then after 1700h begins to increase. Flow acceleration and rotation toward low pressure can be induced by rotation of the stress divergence term toward low pressure (B is imaginary positive) as can be seen by examining (4). This corresponds to convergence of momentum in the direction toward low pressure which could result from occasional exchange of momentum with flow near the surface which is rotating toward low pressure even more rapidly.

The rotation of the wind toward low pressure, and implied increase of ageostrophic flow, decreases with height and vanishes in the upper part of the mixed layer. At these levels the wind accelerates and rotates toward high pressure during the transitional period. Such modification of the flow can be induced by decreasing magnitude of the stress divergence (real positive B) and initiation of the nocturnal inertial oscillation. By 1900h, the wind has begun to accelerate and rotate toward high pressure at levels down to 100 m. After 2100h, even the surface flow begins to rotate toward high pressure probably due to some downward mixing of momentum from higher levels. Between 200 m and 300 m a wind maximum usually develops partly due to the decrease of ageostrophic flow with height and frictional retardation of the flow near the surface.

The above arguments are necessarily speculative because the flow was assumed to be initially in balance with the pressure gradient in order to avoid specification of the geostrophic wind. In the next section we attempt more direct estimates of the ageostrophic wind by using several different estimates of the geostrophic wind.

4. Momentum budget based on geostrophic estimates

There is no completely satisfactory method of estimating the geostrophic wind speed with sufficient accuracy for determining ageostrophic winds. Here we will use geostrophic wind estimates constructed by Clarke et al. (1971) from both the synoptic scale pressure field and special 5-station mesoscale network.

A third estimate is constructed by assuming an initial balance at 1500h between the stress divergence, Coriolis and pressure gradient terms and by using recent estimates of the surface stress derived by Hicks (1980) and estimates of the mixed-layer depth (inversion height) constructed by Melgarejo and Deardorff (1974). We assume that the stress vanishes at the top of the boundary layer and that the surface stress is in the same direction as the wind which is vertically averaged across the boundary layer. The change of wind direction with height was normally small in the boundary layer at 1500h so that the angle between the surface stress and the layer averaged wind speed was probably also small. Therefore assuming this angle to be zero is not expected to seriously affect the qualitative interpretation presented below.

We align the x-axis with the geostrophic wind direction. Then the solution for this vertically averaged wind for a Northern Hemispheric framework is:

\[ \bar{u} = \bar{u}_q(1 - F^2/\bar{u}_q^2) \]

\[ \bar{v} = F(1 - F^2/\bar{u}_q^2)^{1/2} \]

\[ F = u_s^2/fh \]

where the overbar refers to vertical averaging, \( u_s \) is the surface friction velocity and \( h \) is the boundary layer depth.

The layer-averaged wind (5) is rotated toward low pressure from the geostrophic wind by the angle

\[ \alpha = \arctan(F/V) \quad V = (\bar{u}^2 + \bar{v}^2)^{1/2} \]
The magnitude of the geostrophic wind can be expressed in terms of the actual flow using (5) and (6), in which case

\[ |V_g| = F \sin \alpha + V \cos \alpha \]  

(7)

Using (6) and (7), the geostrophic winds on Wangara fair weather days were estimated from the 1500h, 300 m observed winds. The low-level jet on fair weather days is normally near or just below 300 m with weak shear above this level. The use of winds at a given level, instead of layer-averaged winds, assumes that the boundary layer is well mixed and barotropic. The latter assumption is reconsidered in section 5.

For the fair weather days, the magnitude of the vector difference between the 1500h 'balanced' geostrophic wind estimated from (6–7) and the 1500h geostrophic wind estimated from the synoptic pressure field averages approximately 2 m s\(^{-1}\). This magnitude is a little more than 30% of the average geostrophic wind speed and roughly as large as the average ageostrophic wind speed after the transition. The agreement between the two geostrophic wind calculations is best in cases of strong geostrophic winds. Values of the geostrophic wind computed from the mesoscale 5-station network disagree with both of the other estimates by an averaged value of almost 4 m s\(^{-1}\) and is therefore considered least reliable.

Part of the difference between the balanced geostrophic wind estimate and the estimate based on the synoptic pressure field is due to baroclinity. For example, comparing the 'balanced' geostrophic winds with the vertically averaged synoptic scale geostrophic wind, computed using synoptic scale thermal winds (Clarke et al. 1971), reduces the averaged difference between the synoptic and balanced geostrophic winds to less than 0.5 m s\(^{-1}\).

We proceed by estimating the ageostrophic flow relative to the geostrophic wind computed from the 1500h balance of forces. We then average results over the nine fair weather days in an attempt to reduce the influences of observational errors, pre-existing mixed-layer accelerations, baroclinity and peculiarities of individual days. The ageostrophic wind, so computed at 300 m, increases from about 1 m s\(^{-1}\) at 1500h to a little more than 3 m s\(^{-1}\) at 1800h (Fig. 4(a)). This ageostrophic flow continues to increase, although at a slower rate, after 1800h. However, computations using the 1500h balanced estimate of the geostrophic wind become increasingly inaccurate due to time variation of the geostrophic wind.

The ageostrophic wind at 300 m, based on the synoptic scale geostrophic wind, increases from about 2 m s\(^{-1}\) at 1500h to about 3 m s\(^{-1}\) at 1800h, and then increases much more slowly after 1800h. Below 100 m, the increase of the ageostrophic flow is a factor of two larger (Fig. 4(b)). This implies that the stress divergence becomes quite large in the lowest 100 m which is consistent with the formation of a high-Richardson number layer somewhere between 50 m and 250 m as discussed in section 3.

These computations indicate that the evolution of the stress divergence during the evening transitional period significantly enhances the ageostrophic flow and thus enhances the strength of the nocturnal inertial oscillation and low-level jet.

5. INFLUENCE OF GEOSTROPHIC WIND VARIATION

We now examine the influence of the time and height variation of the geostrophic wind on production of ageostrophic flow and nocturnal accelerations. Even though we have selected fair weather conditions, time variation of the geostrophic wind appears to exert an important influence on accelerations during the course of the night. The magnitude of the vector change of the synoptic scale geostrophic wind, between 1500h and 0300h, averages 95% of the geostrophic wind speed at 1500h and exceeds 50% of the 1500h geostrophic
Figure 4. (a) The time dependence of the ageostrophic wind speed, averaged over fair weather days, with respect to the geostrophic winds computed from the 1500h, 300m balance of forces (solid line) and computed from the synoptic scale pressure field (broken line). (b) Vertical profiles of these averaged ageostrophic wind speeds for different observational times (hours).

wind speed on eight of the nine nights.

Much of the time-variation of the geostrophic wind is related to systematic diurnal variation of the geostrophic wind as is evident from calculation of the diurnal variation of the resultant, synoptic scale, surface geostrophic wind (Fig. 5). The southward geostrophic flow increases by about 3 m s\(^{-1}\) at night or a factor of 2. Averaged over all of the Wangara
days, the nocturnal acceleration of the resultant geostrophic wind is about half as large and oriented more southeastward than southward.

The diurnal variation of the surface geostrophic wind is probably due to diurnal variation of low-level baroclinity. The nocturnal increase in the southward component of the surface geostrophic wind corresponds to colder air to the east and warmer air to the west. This nocturnal temperature gradient is probably generated by surface cooling over the terrain which rises towards the east with a slope of about $7 \times 10^{-3}$ depending on the
scale of the slope computation. The terrain slope to the west of the 'Wangara' site is quite weak. Caughey et al. (1979) found that an organized slope of only $1.4 \times 10^{-3}$, on a similar horizontal scale, could produce significant nocturnal accelerations. Although the computation of the synoptic geostrophic winds for the Wangara experiment included stations on the slope from the mesoscale network, it is not clear how well this geostrophic wind represents the effect of cooling over the sloped terrain.

One can also estimate the geostrophic wind contribution due to cooling along the slope by using the hydrostatic equation. The magnitude of this contribution can be estimated as

$$(g \theta' / \theta_0) (\gamma / f)$$

where $\theta'$ is the boundary layer vertical average of the deviation of the potential temperature from $\theta_0$, the potential temperature at the top of the cooled layer. $\gamma$ is the slope equal to approximately $7 \times 10^{-3}$. For the Wangara experiment, a typical nocturnal value of $\theta' / \theta_0$ is approximately $3 \times 10^{-2}$ in which case this geostrophic wind contribution is a little more than $2 \text{ m s}^{-1}$. This is in approximate agreement with the above.

The actual resultant winds follow a similar diurnal variation. The dashed vectors in Fig. 5 connect the resultant geostrophic wind with the resultant 300 m actual wind and thus serve as an estimate of the resultant ageostrophic wind. The rotation of this ageostrophic flow with time is somewhat erratic in the morning but indicates the possible influence of an inertial oscillation during the day and at night. Note the large increase of the ageostrophic flow during the evening transitional period. Thus a large part of the diurnal variation of the flow can be described by the superposition of the diurnal variation of the geostrophic wind and an inertial oscillation whose amplitude increases sharply during the evening transition. However, continuity between the 0600h and 0900h winds cannot be assumed since the nine fair weather days were not in succession and the sample size is apparently too small to eliminate synoptic trends.

The thermal wind also influences the height variation of the ageostrophic flow prior to the evening transition. The resultant geostrophic wind at any level is simply the resultant surface geostrophic wind plus the resultant thermal wind. The resultant 1500h thermal wind in the lowest kilometre on Wangara fair weather days is directed primarily toward the east with a magnitude of approximately $1.75 \text{ m s}^{-1}$. As a consequence, the 1500h resultant geostrophic wind rotates with height opposite to the direction of frictional rotation of the wind (Fig. 6). Since the mixed-layer depth averages a little less than one kilometre, the layer-averaged, resultant, geostrophic wind (estimated with an 'X' in Fig. 6) appears to be oriented a few degrees away from low pressure compared to the geostrophic wind at the pre-jet level (usually 200–300 m). If the actual wind is well mixed, and thus responds to the layer-averaged geostrophic wind, then baroclinity produces ageostrophic flow at the pre-jet levels which acts to oppose the frictional generation of ageostrophic flow. However, since the mid-level of the mixed layer is only 100–200 m above the pre-jet level, this effect is predicted to average less than $0.5 \text{ m s}^{-1}$. Comparison of these resultant winds must be made with caution due to neglect of certain non-linear effects associated with the averaging as well as the influence of the thermal wind above 1 km on days with deep mixed layers.

Additional evidence can be constructed by estimating the importance of each effect for each day and then averaging over all of the fair weather days. Such computations for 1500h (Table 1) indicate that the thermal wind effect produces an ageostrophic flow at 300 m which averages only $0.38 \text{ m s}^{-1}$. In fact when added to the frictional generation of ageostrophic flow, thermal wind effects have no appreciable direct influence on the net production of ageostrophic flow in agreement with the above resultant wind analysis. However, shear induced by baroclinity could enhance mixing and indirectly influence the change of
Figure 6. Resultant 1500h wind vectors for Wangara fair weather days. Plotted are synoptic geostrophic winds at the surface ($V_{gs}$), 300 m ($V_{g,300}$) and 1 km ($V_{g,1km}$) and vertically averaged over the boundary layer ($x$); actual 300 m wind ($V_{300}$) and the ‘balance’ estimate of the geostrophic wind ($V_{gb}$).

ageostrophic flow during the transitional period as suggested by Zeman (1979).

The difference between the actual 300 m wind and the synoptic scale geostrophic wind at 300 m averages about 1.85 m s$^{-1}$ at 1500h. Since the idealized frictional and thermal wind effects lead to ageostrophic flow of just less than 1 m s$^{-1}$, about half of the total ageostrophic flow must be associated with accelerations and various errors and assumptions in the analysis. As noted in section 4, the ageostrophic flow continues to increase by another 1-2 m s$^{-1}$ during the transitional period.

Since the daytime mixed-layer wind is only slightly weaker than the geostrophic wind, the maximum increase of wind speed due to a hypothetical inertial oscillation with constant geostrophic wind would be approximately equal to the ageostrophic wind speed at the end
of the transitional period. For the Wangara experiment, the ageostrophic flow at the end of the transition averages only about half of the average observed acceleration of about 6.7 m s\(^{-1}\). The additional acceleration is probably due mostly to time dependence of the geostrophic wind and associated increased ageostrophic flow. Between 1500h and 0300h, the geostrophic wind speed increases on the average by almost 2 m s\(^{-1}\) (Table 1).

The associated actual flow acceleration induced by the geostrophic acceleration could be considerably larger than the geostrophic acceleration itself depending on initial imbalances and the particular orientation of the geostrophic tendency to the initial observed and geostrophic winds (Young 1973, Mahrt 1974). Such a possibility is consistent with the fact that the ageostrophic flow, computed from the synoptic scale geostrophic wind, continues to increase by another 1.3 m s\(^{-1}\) during the night after the early evening transition.

Thus the geostrophic tendency appears to induce an acceleration of a little more than 3 m s\(^{-1}\). Such a value is approximately the right magnitude to explain the excess of the observed acceleration over that predicted by an inertial oscillation.

Thus both resultant winds and averages of statistics for individual nights indicate that the geostrophic accelerations account for an appreciable fraction of the observed nocturnal accelerations and that daytime baroclinity has little direct influence on the generation of ageostrophic flow for Wangara fair weather days.

### TABLE 1. Generation of ageostrophic flow (m s\(^{-1}\)). \(V_o\) is the synoptic scale geostrophic wind. Overbar refers to boundary layer average and subscript ‘300’ refers to 300 m level, ‘s’ refers to surface level.

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6. Conclusions and Further Discussion

Nocturnal accelerations and strength of the low-level jet appear to be significantly enhanced by an increase of the low-level ageostrophic flow generated during the transition from the daytime well-mixed boundary layer to the nocturnal boundary layer. During this transition on fair weather days in the Wangara experiment, this ageostrophic flow increases from an initial average value of between 1 and 2 m s\(^{-1}\) to a value between 3 and 3.5 m s\(^{-1}\).

Ageostrophic flow is generated by increased stress divergence in the lower layers apparently due to the fact that the surface stress decreases more slowly than the decrease of downward transport of momentum associated with decreasing depth of the turbulence. In particular, a stable layer (high Richardson number) forms, usually between 100 m and
200 m, due to mixing of cooled air upward above the surface layer into air with weaker shear. The downward transport of momentum through this stable layer probably becomes weak. As a result, the surface flow decelerates and rotates toward low pressure during the transitional period.

Due to these processes, the ageostrophic flow increases especially in the lowest few hundred metres. After the transitional period, an elevated low-level wind maximum develops due to the decrease of the ageostrophic flow with height and due to frictional retardation of the flow near the surface.

The geostrophic wind normally changes significantly during the night, partly due to cooling over weakly sloping terrain. On the average, the geostrophic wind speed increases by about 2 m s⁻¹. The vector magnitude of the geostrophic wind change is normally considerably larger and sometimes generates additional accelerations corresponding to increasing ageostrophic flow. The nocturnal geostrophic tendency seems to account for about half of the observed nocturnal acceleration. The total nocturnal acceleration averages between 6 and 7 m s⁻¹. When averaged over all days, regardless of weather situation, the various effects discussed above are reduced in importance by roughly a factor of 2.

For the Wangara data, even though surface winds become quite weak during the transitional period, the turbulence does not completely collapse. On the contrary, the low-level stress divergence reaches a maximum during this period. These results seem to support the usefulness of considering the boundary layer depth to be a continuous, although rapidly decreasing function of time, during the transitional period as modelled by Zeman (1979) and Smeda (1979). On the other hand, models of diurnal variation of boundary layer flow which replace the transitional period with an instantaneous collapse have appealing simplicity in that they avoid the unresolved problem of rigorously defining the boundary-layer depth during the transitional period (Thorpe and Guymer 1977). In the Wangara simulation of Thorpe and Guymer, the increase in the observed ageostrophic flow during the transitional period was indirectly compensated for by using a large daytime stress coefficient which produced daytime ageostrophic flow considerably larger than observed. An attractive possibility would be to modify the model of Thorpe and Guymer by attempting to parametrize the observed increase of ageostrophic flow during the transitional period. The ageostrophic flow increases most rapidly when the depth of the turbulence decreases rapidly compared to the decrease of surface stress.

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