The Evolution of Afternoon Boundary Layer winds

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1. INTRODUCTION

The original concept of a boundary layer (BL) in fluid flow can possibly be attributed to Froude, who carried out a series of experiments in the early 1870s to study frictional resistance of a thin flat plate when towed in still water (Garratt 1992). His work recognized the different characteristics present near a stationary surface with flowing fluid overhead. Prandtl (1905) also conducted several experiments that highlighted interactions between a solid surface and the air just above it. In meteorology, the boundary layer is defined as the layer of air directly above the Earth's surface in which the effects of the surface (i.e. friction, heating and cooling) are felt directly on time scales less than a day, and which carries significant fluxes of momentum, heat or matter by turbulent motions on a scale of the order of the depth of the boundary layer or less (Garratt 1992). Changes within the boundary layer and their relationship to the formation of the low level jet (LLJ) are of particular interest over the Midwest United States due to the large horizontal moisture advection, the development of strong convergence, and wind profiles (which are critical to the maintenance of large scale thunderstorms).

The LLJ is related to deep convective activity and is also responsible for most of the moisture transport in the central United States. For example, Means (1954) concluded that the LLJ transported enough moisture to produce a region of rainfall covering the entire state of Kansas with 4-7 cm of water. Uccellini and Johnson (1979) computed moisture and sensible heat transports and found that the moisture transport increased by a factor of 3 and the sensible heat transport increased by a factor of 2 owing to the development of a LLJ (Stensrud 1996). LLJs are also related to convective activity, according to studies of mesoscale convective complexes (MCCs) (Maddox 1980). Maddox found that a strong LLJ is a common sign of the environment necessary to produce these large thunderstorm clusters that are responsible for most of the warm-season rainfall over the Great Plains (Fritsch et al 1986).

The formation of the LLJ is thought to be associated with the decoupling of the daytime boundary layer from the ground. By late morning the BL winds are well mixed due to strong positive vertical velocities caused by incoming solar radiation heating the earth’s surface. The surface heating causes the formation of rising, buoyant air currents or parcels. The interaction of horizontal winds with the surface also enhances mixing. The rising parcels and frictional effects cause the winds to be subgeostrophic and cross the isobars at a small angle towards low pressure. When the effects of turbulent mixing come to an end late in the day, the frictional effects are reduced and the winds above the shallow nocturnal inversion are decoupled from the surface layer and are no longer in balance (Stensrud 1996). Blackadar’s (1957) study concluded that the daytime mixing dies down, on average, at about the time of sunset.

The imbalance between the coriolis and pressure gradient force causes the wind to speed up and veer to the right (i.e. the first stage of LLJ formation). According to Blackadar, the LLJ is initiated by a nocturnal inversion that begins when the air near the surface starts to cool while the air in the middle and top portions of the BL is still warmer. Buoyant parcels no longer rise from the ground, and mixing created by wind is suppressed when the air is stable.
Recent radar and profiler data suggests that such a late decoupling may be inaccurate; and that the veering of the wind might occur earlier than previously thought (Julie Lundquist, personal communication 2000). Prior evidence for the ceasing of daytime mixing was primarily from temperature profiles derived from radiosonde data. This method may not be an ideal measure of vertical mixing taking place within the BL. We used the Cooperative Atmosphere Surface Exchange Study’s (CASES-97) radiosonde, aircraft, and profiler data from 5, 10, and 20 May 1997 to study the LLJ’s initiation period during the afternoon hours. These data provided an opportunity to document the wind and to study the decoupling mechanisms that are thought to play a role in the LLJ’s evolution.

The following sections will include a description of data collection and analysis strategies, a results section describing the evolution of the wind profile, a description of the BL and synoptic evolution, results, summary, and conclusion.

2. DATA

a. Data collection

FIG. 2. The instrumentation and preset flight tracks for CASES-97

We used temperature, wind, and turbulence data collected during the Cooperative Atmosphere Surface Exchange Study (CASES-97). CASES-97 was conducted from 21 April to 17 June 1997 in the Walnut River Watershed east of Wichita, Kansas. The area was chosen as the focus of investigations of the lower atmosphere with the land surface (LeMone et al. 2000). There were three profiling sites in the CASES-97 array (Figure 2), at Beaumont (BEA), Whitewater
(WHI), and Oxford (OXF). The three locations were equipped with 915-MHz radar wind profilers
to collect wind data from 100 m to a few kilometers above ground level (AGL). Automatic
Weather Stations (AWSs) sampled wind at 10 m, temperature and humidity at 2 m, and
precipitation at Beaumont and Oxford (LeMone et al 2000). During the six Intensive Observing
Periods (IOPs), which occurred during 24-hour periods of fair weather when winds were expected
to be steady, additional measurements were taken.

During IOPs the radiosondes were released every 90 minutes from the three profiler sites.
The profilers collected data 50 minutes out of every hour. A technique was used for removing the
effects of migrating bird echoes in the wind profiler data (Pekour and Coulter 2000). The
radiosondes and profilers were used to obtain temperature and wind data, respectively.

Two aircraft were used to gather wind and turbulence data during IOPs. The NOAA Twin
Otter (TO) and the University of Wyoming King Air (KA) sampled vertical flux profiles,
horizontal fields, and BL depth using a combination of roughly cross-wind straight-and-level
flights at multiple levels or stack patterns of approximately 40 kilometers in length (Figure 2), and
alternating flights just above the surface and just below the inversion intermingled with flights
around the perimeter of the array. The aircraft have a gust probe system and fast temperature and
water vapor sensors for estimating heat and moisture fluxes. A more detailed description of the
instruments used on the aircraft as well as a description of the surface collecting data can be found

We used synoptic analyses (surface and 850 millibar) data originating from the National Center
for Environmental Prediction (NCEP) for 5, 10, and 20 May 1997.

3. Data analysis

a. Evolution of wind profile

Wind data from three sources (KA, TO, and wind profiler) were plotted as a function of
height as well as at a given level as a function of time to determine when the BL began to decouple
from the surface. Aircraft wind data were spatially averaged along the flight track.

b. Evolution of boundary layer

Radiosonde potential temperature, aircraft vertical velocity standard deviation, and
turbulent kinetic energy (TKE) were used to track the evolution of the BL. The potential
temperature profile data were used to show the initiation of the inversion near the surface. The
vertical velocity standard deviation ($\sigma_w$), which is a reflection of thermals rising from the ground (a
source of mixing daytime boundary layer), is used to aid in the determination of the decoupling
process. TKE, a measure of mixing from rising plumes plus the mixing induced by the wind, was
also examined to determine the timing of the decoupling process. Both $\sigma_w$ and TKE were
evaluated over each aircraft track after linearly detrending the data.

c. Evolution of synoptic scale

Synoptic analyses were used to verify if any wind change was due, in part, to synoptic scale
conditions. The isobar angle of the surface data was measured in 3-hour increments to calculate
geostrophic wind direction changes just above the surface.

4. RESULTS
The BL winds during the daytime hours on all three days were typical for fair weather convective boundary layer conditions, with little, if any change with height. Winds on the 4 and 20 May remained primarily the same until at least 1800 while winds on the 10 May began to veer at approximately 3:00 PM Local Standard Time (Figure 3, more vertical wind profiles are available at http://gonzalo.er.anl.gov/ABLE/dataarchive.html). The wind veering on the 10 May was thus considerably earlier than expected.

Figure 6 shows the U and V components of the wind for 10 May, by level, as a function of time. U is positive east, and V positive north. Both components of the wind tend to increase with time. The increase in U is consistent with the shift of the wind from south to southwest in Figure 3.

Synoptic scale surface maps were investigated to determine whether synoptic conditions explained the wind veering on the 10th. The maps indicated a slight backing of the geostrophic wind on the 10th (Figure 7). The backing indicated that the wind shift did not appear to be "synoptically" driven. Radiosonde potential temperature profiles indicated that the surface inversions formed at approximately sunset (Figure 8). This result was expected from the work of Blackadar (1957). Blackadar expected decoupling at sunset associated with the inversion development. Since decoupling apparently occurred before inversion development, the development of the low-level inversion did not explain wind behavior on the 10th.

We used two measures of BL mixing for changes that might explain why decoupling occurred on the 10 May but not 4 May. The first was vertical velocity standard deviation ($\sigma_v$), a measure of mixing by plumes rising from the surface. The second was the turbulence kinetic energy (TKE), a measure of vertical mixing within the BL,

$$TKE = \frac{1}{2}(\sigma_u^2 + \sigma_v^2 + \sigma_w^2)$$

$\sigma$ denotes standard deviation

Vertical velocity standard deviations behaved similarly for the observed three days (Figure 10), and thus do not provide an explanation for the difference between the 4 May and the other two days. TKE values fell sharply in the late afternoon on both 10 and 20 May (Figure 8), but they remained high on 4 May. TKE remained high on 4 May because of higher winds that kept the BL well mixed (Figure 3). We interpret this as meaning that the moderate wind resulted in more turbulence, which, in turn, led to a continued mixing of the BL on the 4 May. Thus TKE separates out the non-veering 4 May from the veering 10 May, but does not explain why wind veered on the afternoon of 10 May but not the 20 May.

5. SUMMARY AND CONCLUSIONS

The purpose of this study was to compare the afternoon winds on the three days and to look for possible causes responsible for wind shifts. The wind evolution on 4 May and 20 May fit our expectations gained from prior studies. Winds on the 10 May, however, did not. The winds on the 10 May veered earlier than expected and synoptic conditions did not appear to have any effect on wind direction or speed. Investigation of the decoupling mechanism, as represented by changes in $\sigma_v$ and TKE, failed to completely explain the results.

TKE measurements suggested that the BL mixing continued on the 4 of May. We have yet to find a cause that is able to explain the differences observed between the 10 and 20 of May. All of the observations would indicate little, if any, differences between the two days.
Future work will include examination of the vertical divergence of vertical flux of horizontal momentum difference between the three days. Also, synoptic conditions could be studied more rigorously to possibly locate “up-stream” perturbations that may have affected overnight winds.
Figure captions

Figure 1. Pressure gradient diagram.

Figure 2. CASES-97 array.

Figure 3-5. Vertical wind profiles from Whitewater profiler for (a) 4 May, (b) 10 May, and (c) 20 May.

Figure 6. Average flight-level winds as a function of time, for four ranges of altitude, 10 May. U is positive east, V is positive north.

Figure 7. Isobar angle with respect to the east from surface analyses, 10 May, as a function of time. High pressure was to the east, so a 45-degree angle signifies geostrophic wind from the southwest.

Figure 8. Radiosonde potential temperature profiles at Beaumont during late afternoon, 10 May. Note the nocturnal inversion is just beginning to form at 0030 UTC 11 May. This is typical behavior for all the IOPs.

Figure 9. Standard deviation of vertical velocity, by flight leg, as a function of flight-leg center time for 4, 10, and 20 May, based on King Air data.

Figure 10. Turbulence kinetic energy, by flight leg, as a function of flight-leg center time for 4, 10, and 20 May based on King Air data.
ABLE-WHITE Wind Profile (high) 970504  18 LST

Height (km)

Hours Before 18 LST

Wind Speed (knots)
Figure 6
References

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