On the Initiation of the 20 June 2015 Great Plains Low-Level Jet

THOMAS R. PARISH
Department of Atmospheric Science, University of Wyoming, Laramie, Wyoming

RICHARD D. CLARK
Department of Earth Sciences, Millersville University of Pennsylvania, Millersville, Pennsylvania

(Manuscript received 18 May 2016, in final form 14 April 2017)

ABSTRACT

Extensive measurements were made of the summertime Great Plains low-level jet (LLJ) in central Kansas during June and July 2015 as a component of the Plains Elevated Convection at Night (PECAN) field study. Here, the authors describe the early phase of the LLJ development on 20 June 2015. Half-hourly soundings were launched to monitor the progress of the jet. An airborne mission was also conducted using the University of Wyoming King Air research aircraft. Vertical sawtooth patterns were flown along a fixed track at 38.7°N between longitudes 98.9° and 100.3°W to document changes in the potential temperature and wind profiles. Ageostrophic winds during the LLJ formation were also assessed. In addition, a high-resolution numerical simulation of the 20 June 2015 LLJ case was conducted using the Weather Research and Forecasting Model. Observations and model results show that the early stage of development consisted of a rapid increase in wind speed in the hours just after sunset with less pronounced directional change. The LLJ evolution is similar to that expected from an inertial oscillation of the ageostrophic wind following the stabilization of the near-surface layer.

1. Introduction

The Great Plains low-level jet (LLJ) is among the most scrutinized mesoscale features of the lower atmosphere (e.g., Bonner 1968; Bonner and Paegle 1970; Mitchell et al. 1995; Whiteman et al. 1997). Studies have shown that relatively weak southerly winds in the daytime summer boundary layer often become significantly enhanced during the nighttime and early morning hours. A well-defined jet profile becomes established several hundred meters above the surface a few hours after sunset. Maximum wind speeds often exceed 20 m s⁻¹. Wind directions typically shift from southerly early in the evening to southwesterly later on at night. Observations from soundings and profilers indicate that the LLJ is geographically centered over the Great Plains states (e.g., Bonner 1968; Mitchell et al. 1995; Whiteman et al. 1997). Bonner (1968) depicts the center of LLJ frequency over central Oklahoma.

The forcing of the LLJ has received considerable attention during the past 60 years. The most prominent theory is that of Blackadar (1957). He views the development of a nocturnal wind maximum in the lower atmosphere to be the result of the sudden decay of turbulence arising from surface cooling near sunset. Such stabilization of the lowest levels decouples the lower atmosphere from surface processes; acceleration of the wind in the lower atmosphere occurs as an inertial oscillation of the ageostrophic wind. While this mechanism has received widespread support (Parish et al. 1988; Zhong et al. 1996; Shapiro and Fedorovich 2010), it has been pointed out repeatedly (e.g., Holton 1967; Shapiro and Fedorovich 2009) that such a process cannot explain the geographical preference of the LLJ to the sloping terrain of the Great Plains.

A second popular theory is that of Holton (1967), who has indicated the diurnal cycle of heating and cooling of the sloping terrain associated with the Great Plains is essential in understanding the nocturnal wind maximum. His work follows that from Bleeker and André (1951), which proposed that low-level nocturnal convergence is present west of the Rockies as a result of drainage flow over the slopes of the Great Plains.

There is no debate that the diurnal heating cycle is responsible for an oscillation in the horizontal pressure gradient force. Bonner and Paegle (1970) also considered the importance of the diurnal oscillation in temperature over the Great Plains in the development of the LLJ. They noted that the diurnal heating results in an
oscillation of $3 \text{ m s}^{-1}$ magnitude in the geostrophic wind at 600 m above the ground at Fort Worth, Texas; a maximum geostrophic wind magnitude is seen at about 2300 UTC and a minimum at 1100 UTC. Bonner and Paegle (1970) conclude that such an oscillation needs to be considered along with the oscillation in the eddy viscosity such as proposed by Blackadar (1957) to understand the LLJ.

Another often-discussed theory is that proposed by Wexler (1961). He notes that the North American Cordillera acts as a barrier to the broad trade wind current associated with the Bermuda high. In this view, air is blocked by the topography and then becomes deflected northward. Westerly intensification of the wind east of the Rocky Mountains is compared to processes responsible for strong western oceanic boundary currents such as the Gulf Stream. While this theory has received some recent support (e.g., Ting and Wang 2006), Holton (1967) originally dismissed this idea since scale analysis of the equations of motion show that boundary currents in the ocean are not comparable to the LLJ. Parish and Oolman (2010) note that observations and modeling do not support the Wexler (1961) contention that the North American Cordillera acts as a barrier to the air moving about the Bermuda high.

Theories proposed by Blackadar (1957) and Holton (1967) continue to be discussed as forcing mechanisms for the LLJ. Work has attempted to evaluate the LLJ in terms of these two theories or some combination of each (Jiang et al. 2007; Du and Rotunno 2014). Recently, Shapiro et al. (2016) have developed a coupled equation set to evaluate the respective forcing from Blackadar (1957) and Holton (1967) mechanisms. They conclude that the diurnal heating and cooling of the sloping terrain alone only produces weak jetlike profiles. Representative LLJ events are simulated by including effects of frictional decoupling as discussed by Blackadar (1957).

Here, we regard the summertime nocturnal wind maximum as forced fundamentally through the Blackadar (1957) mechanism. As originally pointed out by Parish and Oolman (2010), the high frequency of LLJ occurrence over the Great Plains is due to the time-averaged summertime heating of the sloping terrain. Emphasis is placed on the seasonal time scale of the summertime heating process. Net summertime solar radiative heating of the Great Plains sloping topography is responsible for an enhanced horizontal pressure gradient in the lower atmosphere and hence southerly geostrophic wind that is prerequisite to the LLJ such as shown in Parish and Oolman (2010) and Shapiro et al. (2016).

In this paper, we examine the initial development stages of an LLJ on 20 June 2015 from surface-based and airborne observations taken as part of the Plains Elevated Convection at Night (PECAN) field study. Particular attention is given to the diagnosis of the geostrophic and ageostrophic components of motion as the jet forms. This permits evaluation of the physical processes acting to force the LLJ. Observational analyses are supplemented with a high-resolution simulation using the Weather Research and Forecasting (WRF) Model, which allows comparison of the dynamics responsible during the LLJ formation.

2. Observations of the 20 June 2015 Great Plains LLJ

PECAN was a large multiagency field project conducted during June and July 2015 in the Great Plains region of the United States. The overarching goal of PECAN was to better understand nighttime convective storms. Given the acknowledged importance of the LLJ in convection over the Great Plains (e.g., Maddox 1980), studies were also conducted of the LLJ. One objective of the LLJ missions was to directly measure the mesoscale dynamics of the jet.

Figure 1 presents an overview of conditions at 0000 UTC 20 June 2015 from the operational North American Mesoscale Forecast System (NAM) for the PECAN area. The sea level pressure field (Fig. 1a) shows a strong horizontal pressure gradient across the Great Plains states from Texas to South Dakota. As noted previously, a prerequisite condition for an LLJ is a strong southerly geostrophic wind in the daytime boundary layer. Conditions on this day were ideal; southerly geostrophic winds with a magnitude exceeding $20 \text{ m s}^{-1}$ at the surface were present over western Kansas. Such a surface pressure field as shown in Fig. 1a is commonly seen during the late afternoon over the Great Plains during summer (Parish 2016).

The pronounced horizontal pressure gradient at the surface over the Great Plains as shown in Fig. 1a is a consequence of the summertime heating of the sloping terrain. During summer, the intense solar radiation results in strongest heating at the surface. Differential heating thus occurs at a particular isobaric level over the sloping Great Plains where the warmest air is found over the elevated terrain to the west. As a result, an isobaric temperature gradient becomes established with warm air to the west and cooler air to the east. The magnitude of the isobaric temperature gradient is largest at the lowest levels of the atmosphere. Observations from 20 June 2015 show that isobaric temperature gradients arising from the diabatic heating of the sloping terrain extend to about 600 hPa.

The effect of such isobaric temperature gradients is profound. The magnitude of the isobaric temperature gradient is proportional to the thermal wind. For the
case of the Great Plains summertime boundary layer, a thermal wind from the north must be present. This requires that the southerly component of the geostrophic wind must increase moving down from the free atmosphere, approximately 600 hPa, toward the surface. Large isobaric temperature gradients imply a large increase in the southerly component of the geostrophic wind at lower levels of the atmosphere.

Parish and Oolman (2010) note that the westerly intensification of the horizontal pressure gradient force at the surface over the Great Plains for time-averaged conditions is a consequence of the thermal wind. The same process can be seen for this case as well. To illustrate, mean isotherms between 850 and 950 hPa are also shown in Fig. 1a. Isotherms across the Great Plains states are oriented primarily in a north–south direction. This alignment follows the general terrain contours, with warmest air over the highest terrain to the west. A mean isobaric temperature gradient of approximately 1.6 K (100 km)$^{-1}$ is present between 850 and 950 hPa across southern Kansas. Such an isobaric temperature gradient corresponds to a 5.6 m s$^{-1}$ thermal wind from the north and hence requires that the southerly component of the geostrophic wind must increase by 5.6 m s$^{-1}$ from 850 to 950 hPa.

The 600-hPa height field at 0000 UTC 20 June 2015 (Fig. 1b) differs significantly from that shown at the surface in Fig. 1a. Isotherm orientation at 600 hPa displays only a weak relationship to the underlying terrain. We used the height field at 600 hPa as an estimate of the free atmosphere forcing during PECAN. For this case, weak northwesterly geostrophic winds were present at 600 hPa over central Kansas. The horizontal pressure gradient associated with the synoptic forcing at middle levels of the troposphere thus contains a northerly component, opposite to the direction of the LLJ.

Thermal wind forcing from 600 hPa down to the surface must first overcome the northerly geostrophic wind component at 600 hPa. We interpret the strong surface pressure gradients shown in Fig. 1a as the cumulative effects of the thermal wind from the free atmosphere near 600 hPa down to the surface. Integrating the thermal wind downward from 600 hPa to the surface over the general southern Kansas region (not shown) yields a geostrophic wind in excess of 15 m s$^{-1}$. This is sufficient to overcome the weak adverse northerly geostrophic wind at 600 hPa and support an increasingly strong southerly geostrophic wind at isobaric levels approaching the ground. Heating of the sloping terrain is vital to the existence of the LLJ in this case.

Figure 2a shows the location of the PECAN study region. The center of operations for PECAN was in Hays, Kansas (HYS; Fig. 2a); the University of Wyoming King Air research aircraft (UWKA) was based out of Great Bend, Kansas (GBD; Fig. 2a), about 75 km southeast of Hays. For the 20 June 2015 case, the UWKA was launched at 0045 UTC and returned to base at 0533 UTC. As with all LLJ flights during PECAN, the UWKA flight strategy was to conduct a climbout sounding from GBD followed by an isobaric leg to the north at about 885 hPa [about 500 m above ground level (AGL)], near the level of the LLJ maximum. Repeating vertical sawtooth and isobaric legs were then conducted along a fixed track at 38.7°N between longitudes 98.9° and 100.3°W (see dark solid line in Fig. 2a for track).

Details of the sawtooth profiling along the fixed track are shown in Fig. 2b. Starting at the east end of the fixed track, vertical sawtooth maneuvers were conducted
between approximately 150 and 1000 m AGL. Ascent–descent rates were limited to about 300 m \text{min}^{-1} to ensure accurate wind measurements, requiring a horizontal traverse of about 20 km per individual ascent–descent. A total of six ascent–descent soundings were conducted over the approximate 125-km leg. All vertical sawtooth legs began and ended at low levels.

Eight sawtooth legs were completed during the mission. After the first four were completed, repeating isobaric legs were flown at 885 hPa along the same east–west track. After the remaining sawtooth profiles were completed, an isobaric leg heading directly south was conducted again at 885 hPa prior to landing at Great Bend.

Two key measurements were made during the vertical sawtooth profiling. The series of soundings along the track enable determination of the LLJ kinematics and the thermodynamic structure of the lower atmosphere. In addition, the differential GPS measurements (i.e., Parish and Leon 2013) enable precise vertical positioning of the UWKA and thereby allow instantaneous \(D\) values to be determined. A \(D\) value is simply the difference between the height of a pressure surface and the height of that same pressure surface in a U.S. Standard Atmosphere. Isobaric variation of \(D\) values is proportional to the horizontal pressure gradient force, similar to geopotential height. As shown originally by Bellamy (1945), use of \(D\) values in a cross section eliminates the hydrostatic component of height in a U.S. Standard Atmosphere and thus can illustrate the vertical changes of the horizontal pressure gradient force and hence the thermal wind.

Sawtooth profiling permits construction of \(D\)-value cross sections along the track (Parish et al. 2016). Such \(D\)-value cross sections allow estimation of the vertical profile of the horizontal pressure gradient. To check measurements of the pressure gradient force (PGF) as determined from the sawtooth maneuvers, isobaric legs were conducted along the track at a height roughly at the middle of the sawtooth legs. As noted in Parish et al. (2007), height measurements provided by these isobaric legs enable precise determination of the isobaric slopes of the UWKA and hence the PGF.

Figure 3 summarizes soundings obtained from the UWKA sawtooth profiles during the flight. Sunset at Hays (near the center of the flight track) on 20 June is at 0205 UTC, about 80 min after takeoff. Soundings shown in Fig. 3 are for the westernmost ascent–descent legs. Figure 3a illustrates the UWKA potential temperature profiles. The earliest potential temperature profile in Fig. 3a is from 0124 UTC. Most of the sampled boundary layer is characterized by uniform potential temperature profile, indicating a well-mixed lower atmosphere. Cooling is evident in the potential temperature profile at 0215 UTC, just 10 min after sunset. Close inspection shows potential temperatures at the lowest levels of the sawtooth profile to have decreased by about 0.5 K from the upper boundary layer. Such a cooling is sufficient to decouple the surface layer from the so-called residual layer (e.g., Stull 1988). Observations on this and other days suggested a rapid stabilization process that alters the force balance in the lower atmosphere and prompts commencement of the inertial oscillation of the ageostrophic wind early in the evening.

The cooling process continues, and by 0355 UTC, temperatures at the lowest levels of the sawtooth have decreased by 2°C from that conducted 90 min earlier. Potential temperature profiles indicate strongest cooling at the lowest levels; above about 600 m above mean sea level (MSL), warming occurs. Dewpoint temperatures in Fig. 3a show a marked decrease from 0215 to 0355 UTC, suggesting subsidence. Such a feature was observed during other LLJ missions during PECAN and implies an organized east–west circulation that accompanies LLJ evolution.

Winds (Fig. 3b) display a rapid acceleration during the first few hours after sunset. Wind speeds at 1100 m MSL
increase by about 6 m s\(^{-1}\) from 0215 to 0355 UTC. A jet profile develops by 0445 UTC with maximum LLJ winds of 27 m s\(^{-1}\). Wind directions (Fig. 3b) display little directional change early in the evening. Clockwise rotation becomes evident between 0355 and 0445 UTC, roughly matching that expected from an inertial oscillation of the ageostrophic wind.

Personnel from Millersville University of Pennsylvania launched soundings at half-hour intervals throughout the night near Hays (https://doi.org/10.5065/D6GM85DZ). Figure 4 shows sounding profiles at 1-h intervals during the early evening coincident with the UWKA flight. An advantage of the ground-based sounding is that measurements within the first 100 m AGL are possible. Rapid destabilization is apparent from the potential temperature profiles (Fig. 4a) between 0000 and 0200 UTC; the potential temperature at the lowest level indicates cooling of about 2 K from 0000 to 0200 UTC. This suggests that the near-surface layer became decoupled from the residual layer between 0100 and 0200 UTC. As with the aircraft data, cooling is strongest below about 1000 m MSL (about 400 m AGL), and warming from subsidence occurs above that level.

Wind speeds (Fig. 4b) show a rapid intensification in the early evening. From 0200 to 0500 UTC, the wind speed at 890 hPa increases from about 15 to 25 m s\(^{-1}\), and a jet profile becomes developed. Little increase in wind speed occurs during the remainder of the night. The LLJ maximum occurs just above the inversion shown in Fig. 4a. Wind directions (Fig. 4c) show only minor variation early in the evening but more significant veering with time as the LLJ reaches its maximum speed.

To provide an idealized construct from which to evaluate physical processes at work at the level of the LLJ, a simple model following the differential equation system originally suggested by Blackadar (1957) was constructed. Simply stated, if we neglect friction, the x and y components of motion can be written as

\[
\frac{du}{dt} - f v = -f u_g \quad \text{and} \\
\frac{dv}{dt} + f u = f v_g.
\]

Here, \(u\) and \(v\) are the x and y wind components, \(u_g\) and \(v_g\) are the x and y geostrophic wind components, and \(f\) is the
If we assume that the geostrophic winds in the \( x \) and \( y \) directions are 10 and 15 m s\(^{-1}\), respectively. These values are taken from NAM analyses. As will be shown later, such values are also in reasonable agreement with airborne measurements. We also assume, based on the Millersville sounding data, that the initial wind components are \( u = 1 \) m s\(^{-1}\) and \( v = 11 \) m s\(^{-1}\). Decoupling is assumed in this case at \( t = 0 \).

Solutions to this simple equation system are shown in Fig. 5. In essence, the ageostrophic wind at the time of initial decoupling rotates about the geostrophic vector with a period equal to that of an inertial oscillation (about 19 h in central Kansas). Figure 5a illustrates the wind vectors at 2-h increments for the first 6 h. Wind speeds increase rapidly within the first few hours after decoupling with only minor variation in wind direction. Comparison of wind speed and wind direction with results from the Millersville soundings at the level of the LLJ maximum (about 1070 m MSL) is shown in Fig. 5b. Without question, the simple model provides a reasonable picture as to the observed changes in wind speed and wind direction at the level of the LLJ. The good agreement between the idealized analytical model and observations at the level of the maximum wind shows that the evolution of the wind is consistent with the Blackadar (1957) mechanism in the development of the nocturnal LLJ.

Given the above equation system, the ageostrophic components are out of phase by 90° (Fig. 5c). During the initial acceleration phase, the \( y \) component of the ageostrophic wind \( v_{ag} \) displays a rapid increase from values less than 0 m s\(^{-1}\) at the onset of the residual layer decoupling to about 7 after 4 h. Such a conceptual depiction of the ageostrophic wind components allows a framework with which to compare airborne observations. Given that the track of the UWKA was along a line of constant latitude, only the \( y \) component of the ageostrophic wind is continuously monitored during the eight vertical sawtooth maneuvers. Measurements of the \( x \) component of the ageostrophic wind are made only during the isobaric legs conducted to start the mission at 0050 UTC and on return to base at GBD at 0530 UTC.

To understand the dynamics of the LLJ, it is necessary to measure both the actual and geostrophic winds. Differential GPS measurements made during PECAN provide a precise and independent measure of the aircraft
height (see also Parish and Leon 2013) that, combined with a measurement of static pressure, enables detection of the horizontal pressure gradient force. As a means to determine the vertical profile of the pressure gradient force, $D$ values have been calculated for the vertical sawtooth profiles. In the original paper describing $D$ values, Bellamy (1945) used the U.S. Standard Atmosphere as his reference. The choice of the U.S. Standard Atmosphere as a reference atmosphere is arbitrary. For this case, we calculated $D$ values based on a reference atmosphere with an adiabatic lapse rate, a sea level temperature of 314 K, and a sea level pressure of 1003 hPa. The $D$-value analyses that follow use this reference atmosphere.

Figure 6 illustrates results from airborne measurements obtained from sawtooth profiling for the first leg at 0057–0119 UTC and fourth leg at 0326–0349 UTC. For the initial sawtooth (Fig. 6a), the wind measurements suggest only a weak jet with a maximum speed of 16 m s$^{-1}$. Potential temperature measurements indicated a weak gradient of about 1 K (100 km)$^{-1}$; warmest air is situated over the western part of the domain. Such a temperature gradient is equivalent to a thermal wind of about 3 m s$^{-1}$ between 925 and 850 hPa, suggesting that the geostrophic wind must decrease by about that amount between the lowest and highest profile levels along the flight track. The vertical orientation of the isentropes is indicative of a well-mixed boundary layer and is characteristic of the daytime. The $D$ values are oriented nearly vertical as well. A geostrophic wind of about 15 m s$^{-1}$ has been calculated for the lower levels of the sawtooth.

Significant changes have taken place in the lower atmosphere about 2.5 h later (Fig. 6b). The LLJ has become well established with a clearly defined jet profile and a maximum wind in excess of 23 m s$^{-1}$ along the westernmost part of the flight track. Coincident with LLJ development, isentrope orientation becomes more horizontal in response to surface cooling. Warmest air remains to the west. Evidence of subsidence is seen in the upper part of the vertical profile over the western end of the profile. Isobaric temperature gradients are apparent and result in a thermal wind. Note that the $D$-value gradients relax with height; this is associated with the isobaric temperature gradients and hence the thermal wind. Close inspection reveals a slight enhancement of the $D$-value gradients as compared with Fig. 6a, signifying an increase in the horizontal pressure gradient force over the 2-h period since the initial sawtooth.
The LLJ is near maximum strength by about 0500 UTC (Fig. 7). Wind speeds reach 25 m s\(^{-1}\) over the western part of the flight track at a height of about 1100 m MSL, corresponding to a height of about 375 m AGL. Isentropes are now directed in quasi-horizontal orientation. The temperature gradient is about zero near the lowest levels of the sawtooth profile, although highest temperatures remain to the west at levels above 300 m AGL. Previously, it has been noted that drainage flows develop at night (e.g., Holton 1967). Aircraft measurements from a case study in July 1983 (Parish et al. 1988) show that isobaric temperature gradients near the level of the LLJ, approximately 300 m or so, are reduced to near zero. The sea level pressure field at 0000 UTC 20 June 2015 in Fig. 1a shows lower pressure to the west. The \(D\) values as shown in Fig. 7 also indicate a horizontal pressure gradient force directed toward the west at lowest levels. We conclude that any drainage flows that might exist must be localized and shallow, certainly beneath the lowest level of the sawtooth profiles.

The \(D\)-value contours at this time take on a characteristic jet profile that has been observed in other LLJ cases during PECAN (Parish 2016). Nocturnal cooling is manifested by decreasing \(D\) values with height at the lowest few hundred meters above ground level. The LLJ core is usually situated near the inflection point of the \(D\)-value contours, near the top of the cooled layer at about 1200 m MSL (880 hPa or about 500 m AGL) over the west end of the flight track (see also Figs. 3 and 4). Above the jet core, isobaric slopes of \(D\) values become reduced owing to thermal wind effects. From Fig. 7, the relationship between the isobaric gradient of potential temperature and changing orientation of the \(D\)-value slope is apparent.

Decreasing LLJ winds above the jet core are linked to the lessening isobaric \(D\)-value slopes, in accordance with the thermal wind.

From analyses shown in Figs. 6 and 7, mean values were computed at isobaric levels that permit profiles of the wind and the \(y\) component of the geostrophic wind to be constructed. The \(y\) component of the geostrophic wind is simply the difference between the \(y\) component of the wind and \(y\) component of the geostrophic wind. Figure 8 illustrates the mean profiles for each time shown in Figs. 6 and 7. Wind speeds (Fig. 8a) display a rapid increase during the first three hours with the jet core beginning to develop.

From the field of \(D\) values, isobaric slopes can be estimated, and hence, the profile of the \(y\) component of the
geostrophic wind can be computed (Fig. 8b). Profiles of the geostrophic wind from the $D$-value fields show maximum values at the lowest levels and a decrease of the geostrophic wind with height. Such trends are consistent with the thermal wind as discussed previously. Heating of the sloping terrain is responsible for diurnal oscillation of the $x$ component of the horizontal pressure gradient force (e.g., Holton 1967); a maximum in the magnitude of the horizontal pressure gradient force occurs near the surface. Likewise, the thermal wind arising from the heating of the Great Plains ensures that the horizontal pressure gradient must decrease with height.

To determine the validity of the computed geostrophic wind profiles from the vertical sawtooth profiles as shown in Fig. 8b, comparison was made with isobaric legs conducted near the level of the LLJ maximum at about 1100 m MSL (about 350 m AGL or 880 hPa). Isobaric legs provide an accurate estimate of the horizontal pressure gradient force (Parish and Leon 2013). During PECAN, reciprocal (back and forth) legs were conducted as a means to check for the accuracy of the geostrophic wind calculations from the sawtooth profiling as well as to check for isallobaric effects owing to a changing pressure field. Such legs were completed just prior to the sawtooth leg shown in Fig. 7b from 0238 to 0323 UTC. From the profile shown in Fig. 8b, the $y$ component of the geostrophic wind was determined to be approximately 18 m s$^{-1}$ at 1100 m MSL. Results from the reciprocal isobaric legs (not shown) indicate the $y$ component of the geostrophic wind to be 18.2 and 18.7 m s$^{-1}$. The magnitude of the geostrophic wind matches with that interpolated from the $D$ values approximately 20 min later at the 1100 m MSL level. In addition, the geostrophic wind measurements between isobaric legs confirm isallobaric effects to be negligible at that time.

Hourly METAR data used to derive the vector wind components over the PECAN domain and the NWS Weather Prediction Center (WPC) surface analysis for this LLJ episode (surface charts not shown) provide additional evidence that the diurnal surface pressure tendency and, in turn, the isallobaric effects were weak relative to the geostrophic wind and friction-induced ageostrophic wind on 20 June.

As shown in the idealized model results in Fig. 5, the $y$ component of the ageostrophic wind $v_{ag}$ should show a rapid increase with time during the first few hours following decoupling of the surface layer. Profiles of $v_{ag}$ are illustrated in Fig. 8c. At the level of the jet core, $v_{ag}$ at 0100 UTC is negative and increases with time such that by 0500 UTC, it is approximately 8 m s$^{-1}$. Values are consistent with those expected assuming the simple model presented earlier (see Fig. 5c). We conclude from these measurements that actual and ageostrophic wind components in the $y$ direction as measured from the UWKA platform for the case on 20 June 2015 follow closely with those expected from the idealized model based on Blackadar (1957).
3. WRF Model study of the 20 June 2015 Great Plains LLJ

Numerical simulation of the 20 June 2015 LLJ was conducted using WRF, version 3.7 (e.g., Skamarock et al. 2008). Four domains were used in this simulation, which was centered over the midpoint of the flight track shown in Fig. 2. Horizontal grid resolutions of 27, 9, 3, and 1 km were employed; the innermost domain consisted of \(211 \times 211\) grid points centered on the midpoint of the UWKA flight track along which the vertical sawtooth profiling was conducted. A vertical grid of 84 sigma levels was used with increasing resolution toward the surface.

Key parameterizations used for the run are the following: Lin (Purdue) microphysics scheme, the new Goddard scheme for longwave and shortwave radiation physics, MM5 surface layer similarity with the unified Noah land surface model, and the Yonsei University boundary layer physics scheme. For this simulation, the model was initialized at 1200 UTC 19 June 2015 using the NAM 12-km horizontal resolution grids. Lateral boundary conditions were specified at 6-h intervals, and the simulation was run for 24 h. For brevity, only details from the innermost domain are presented.

The purpose of the modeling work is to document the evolution of the jet during the early evening. This enables a comparison with airborne observations and soundings. In addition, forcing mechanisms from the WRF Model can be evaluated and compared with analyses from the aircraft measurements. The \(D\) values were added to the WRF output to enable comparison of the forcing mechanisms. Figure 9 illustrates cross sections of \(D\) values, wind speeds, and potential temperatures along the same line as the UWKA sawtooth maneuvers at times corresponding those shown in Figs. 6a and 7. At 0100 UTC (Fig. 9a), isentropes in the lower atmosphere display a vertical orientation similar to that seen in Fig. 6a. This again reflects a well-mixed boundary layer with highest temperatures over the elevated terrain to the west. Such temperature gradients persist throughout the lowest entire column shown in Fig. 6a. Wind speeds from the WRF Model are uniform across a transect corresponding to the UWKA flight track with a maximum of \(14 \text{ m s}^{-1}\) over the western end of the cross section. Maximum winds are slightly weaker and a bit elevated as compared with observations.

The \(D\) values in the lowest 1 km are oriented in a near-vertical manner with the tightest isobaric gradient near the surface. Such a pattern indicates a large horizontal pressure gradient force at the surface. As expected with the simulated isobaric temperature gradients and attendant thermal wind, the horizontal pressure gradient decreases with height. This is especially apparent in the upper part of the cross section shown in Fig. 9a where the orientation and slope of the \(D\)-value contours change in response to the temperature field.

Significant changes in all fields occur as the LLJ evolves as shown in the 0500 UTC cross section (Fig. 9b). The LLJ is now well developed, with maximum winds of about \(22 \text{ m s}^{-1}\) over the highest terrain to the west. The LLJ pattern of wind speed shown in Fig. 9b is similar to the airborne observations, although the WRF Model–simulated jet winds are a bit weaker. Potential temperatures have transitioned to a quasi-horizontal orientation with warm air remaining over the high terrain throughout the lowest 2500 m. Pronounced isobaric temperature gradients persist above the height of the LLJ.

The \(D\)-value contours at 0500 UTC take on a pattern similar to that shown in Fig. 7 with a pronounced nose of...
minimum values just above the jet maximum. Isobaric slopes of $D$ values decrease with height in response to the thermal wind constraints. The position of the LLJ core is similarly located as observations show. The excellent agreement between the WRF Model and observed $D$-value cross sections demonstrate the high fidelity of the WRF simulation.

A summary of the WRF results averaged over the flight track cross section along and valid at times coincident with the airborne observations is shown in Fig. 10. The profile of potential temperature shows that surface cooling has commenced by 0100 UTC, perhaps a bit earlier than observed. Cooling is restricted to the layer below about 1100 m MSL (about 890 hPa), in good agreement with observations (e.g., Figs. 3a and 4a). As with observations, temperatures above about 1100 MSL increase with time as the LLJ develops. Profiles of dewpoint temperatures (not shown) support subsidence, similar to that observed.

Wind speeds (Fig. 10b) indicate a uniform increase from 0100 to 0500 UTC with a maximum reached at about 1100 m MSL, closely matching observations. Averaged over the entire cross section, wind speeds are slightly weaker than those observed. Wind direction at the level of the LLJ core shown in Fig. 10b agrees with observations (Fig. 3b) as well as that expected from the idealized model (Fig. 5b).

The profile of the $y$ component of the ageostrophic wind $v_{ag}$ shows that the maximum amplitude of the oscillation occurs near the level of the LLJ. At 0100 UTC, $v_{ag}$ is negative below the level of the LLJ core. At the level of jet core, $v_{ag}$ increases with time in a manner similar to that expected from the idealized model as shown in Fig. 5c. Note that the profile of the $y$ component of the ageostrophic wind matches that inferred from the $D$-value cross sections in Fig. 8c. These results provide additional testimony as to the validity of the inferred ageostrophic wind components from the vertical sawtooth profiles.

The $y$ component of the geostrophic wind shows an oscillation such as described by Holton (1967) that is maximized at the surface and decreases with height owing to thermal wind constraints described earlier and consistent with the thermal wind apparent in the cross sections shown in Fig. 9. The diurnal oscillation of the $y$ component of the geostrophic wind is about 3 m s$^{-1}$. It is important to note the time-scale differences from the diurnal oscillation (Holton effect) versus the seasonal
forcing. Note that all geostrophic profiles in Fig. 10 show a decrease with height, representing the effects of the thermal wind induced by isobaric temperature gradients. Even during the nighttime hours, warmest air resides over the highest terrain. The seasonal time scale of heating contributes to the mean or background geostrophic wind, whereas the diurnal cycle serves to perturb the mean horizontal pressure gradient force.

4. Summary

Development of the nocturnal wind maximum in the lower atmosphere over the Great Plains states has incited considerable interest during the past half century owing to its importance for severe weather, wind energy, bird migration, and transports of sundry biological and entomological agents. Forcing of the LLJ has been cast in terms of a variety of means, most notably through inertial oscillations first proposed by Blackadar (1957) and the diurnal course of heating and cooling of the sloping Great Plains terrain (Holton 1967; Bonner and Paegle 1970). LLJ studies during PECAN have been aimed at addressing the variety of proposed forcing mechanisms. We acknowledge the central importance of stabilization of the near-surface layer during the early evening. Such a process effectively cuts off the turbulent exchange between the surface and lower atmosphere, thereby decoupling the residual boundary layer from the surface layer. We argue that the “Blackadar mechanism” is the key process in the development of the summertime Great Plains LLJ.

Maximum enhancement of the southerly geostrophic wind through diurnal heating and cooling of the sloping terrain occurs at the surface during late afternoon, approximately 2200 UTC. At the level of the LLJ, typically 400 m or so above the surface, the oscillation in the geostrophic wind is considerably weaker. The net summertime heating of the sloping Great Plains terrain is critical to development of the nocturnal jet observed on this day. Observations and modeling work (e.g., Parish and Oolman 2010) show that the mean horizontal pressure gradient is significantly enhanced over the Great Plains during warm-season months. Net heating of the sloping terrain produces a marked horizontal temperature gradient that extends from the surface to nearly 600 hPa. A northerly thermal wind results from such heating. Integrating the thermal wind downward from 600 hPa to the surface from the NAM as well as WRF simulations yields a geostrophic wind in excess of 15 m s$^{-1}$. The climatological frequency of LLJ over the Great Plains is thus the result of the strong summertime heating of the sloping terrain. The importance of a strong background geostrophic wind in the development of the LLJ cannot be overstated (e.g., Shapiro et al. 2016).

For the LLJ in this case and others during PECAN, two important agents are acting. First, the $x$ component of the horizontal pressure gradient force in the free atmosphere needs to support southerly or at a minimum relatively tranquil north–south geostrophic wind components for an LLJ to form. Large adverse pressure gradients to the LLJ at 600 hPa, as evidenced by strong northerly winds, tend to oppose LLJ development in the lower atmosphere. Second, robust heating of the sloping terrain is prerequisite to the development of the LLJ. For periods of cloud cover or after significant rainfall, solar insolation of the sloping terrain is less effective in establishment of strong horizontal temperature gradients necessary to enhance the southerly geostrophic wind at low levels. Ideal conditions for the development of a strong LLJ thus consist of a supportive southerly geostrophic wind at 600 hPa with ample clear skies that promote heating of the sloping Great Plains terrain and thereby intensify existing horizontal pressure gradients at lower levels through the thermal wind.

The LLJ mission on 20 June 2015 was conducted to explore the thermodynamics and dynamics associated with the early nighttime initiation and development of the LLJ. The 600-hPa height field showed weak northwest winds across central Kansas, mildly opposing the development of an LLJ. Strong heating was present throughout the day, which enabled significant enhancement of the near-surface horizontal pressure gradient force that prompted the decision to conduct an LLJ mission. Soundings conducted at half-hourly intervals by Millersville University of Pennsylvania showed that surface cooling commenced about an hour prior to sunset. At that time, the near-surface layer became decoupled from the residual boundary layer and commencement of the inertial oscillation of the unbalanced wind components began. Within four hours of sunset, wind speeds doubled to about 25 m s$^{-1}$ at 400 m AGL and a well-defined LLJ core became established. Wind speed and wind direction changes during the early evening followed that expected from a simple inertial oscillation.

Airborne measurements were conducted to determine the ageostrophic wind component in the $y$ direction. The increasing LLJ winds were accompanied by a corresponding increase in the $y$ component of the ageostrophic wind. Profiles of the $y$ component of the geostrophic wind showed a maximum pressure gradient at the surface and a decrease with height following thermal wind constraints.

Results from a WRF simulation provided confirmation of the role of daytime heating of the terrain in enhancing the low-level geostrophic winds. As the LLJ developed during the early evening, the ageostrophic
wind components evolved in a manner consistent with the Blackadar (1957) theory and similar to the airborne observations.

Acknowledgments. This research was supported by the National Science Foundation through Grant AGS-1359645 to the University of Wyoming and AGS-1359720 to Millersville University of Pennsylvania. The authors gratefully acknowledge the entire PECAN team that made the LLJ component such a success. The authors thank pilots Brett Wadsworth and Tom Drew, scientists Jeff French and Larry Oolman, and the entire Wyoming crew for help with the PECAN field study and UWA measurements. The authors also thank Todd Sikora and the team of 17 students from Millersville University of Pennsylvania who worked tirelessly throughout the PECAN project. WRF simulations were conducted using resources provided by the Advanced Research Computing Center at the University of Wyoming. We thank Jared Baker for his continuing help with the WRF simulations.

REFERENCES