

NOTES AND CORRESPONDENCE

On the Role of Sloping Terrain in the Forcing of the Great Plains Low-Level Jet

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ABSTRACT

The summertime Great Plains low-level jet (LLJ) has been the subject of numerous investigations during the past several decades. Characteristics of the LLJ include nighttime development of a pronounced wind maximum of typically $15\text{--}20\text{ m s}^{-1}$ at levels $300\text{--}800\text{ m}$ above the surface and a clockwise rotation of the wind maximum during the course of the night. Maximum frequency of occurrence of the LLJ is found in the southern Great Plains. Theories proposed to explain the diurnal wind maximum of the Great Plains LLJ include inertial oscillation of the ageostrophic wind, the diurnal oscillation of the horizontal pressure field associated with heating and cooling of the sloping terrain, and the western boundary current interpretations. A simple equation system and output from the 12-km horizontal resolution Weather Research and Forecasting Nonhydrostatic Mesoscale Model (NAM) for July 2008 are used to provide evidence as to the importance of the Great Plains topography in driving the LLJ. Summertime heating of the sloping terrain is critical in establishing the climatological position for the Great Plains LLJ. Heating enhances the background geostrophic flow associated with the Bermuda high, resulting in a maximum low-level mean summertime flow over the Great Plains region. Maximum geostrophic winds in the NAM are found during late afternoon, providing a large background wind on which frictional decoupling can act. The nighttime LLJ maximum is the result of an inertial oscillation of the unbalanced components that arise fundamentally from frictional decoupling. Diurnal heating of the sloping terrain forces a cycle in the geostrophic wind that is out of phase with the wind maximum.

1. Introduction

The summertime Great Plains low-level jet (LLJ) of the central United States is one of the most intensely studied mesoscale features of the past 50 years (e.g., Lettau and Davidson 1957; Hoecker 1963; Bonner 1968). Wind profiles at Great Plains sites during the daytime show weak southerly winds, often less than 5 m s^{-1} . Speeds increase significantly in the hours after sunset with an LLJ developing at levels $300\text{--}800\text{ m}$ above the ground. LLJ wind speeds reach a maximum sometimes in excess of 20 m s^{-1} during the early morning hours with a direction that varies in general from southerly near midnight to southwesterly by dawn. Observations from soundings and profilers (e.g., Bonner 1968; Zhong et al. 1996; Whiteman et al. 1997) and results from model

simulations show the summertime LLJ to be centered geographically over the southern Great Plains from Texas northward to Nebraska with a maximum over northern Oklahoma and southern Kansas.

A number of theories have been used to explain the occurrence of the LLJ. In this note the focus is on the summertime LLJ. Effects such as leeside troughing in forcing a LLJ over the Great Plains or the interaction of the LLJ to an upper-level jet core (Uccellini 1980) are not considered. Blackadar (1957) considered the LLJ to be supergeostrophic, resulting from an inertial oscillation of the ageostrophic wind owing to the sudden decay of turbulence in the boundary layer after sunset. Results from airborne measurements (Parish et al. 1988) and numerical experiments (Zhong et al. 1996) suggest that the Blackadar mechanism is the primary reason for the nighttime boundary layer wind speed maximum.

Holton (1967) noted that the oscillation in the wind may be due in part to the diurnal heating and cooling of the terrain slopes, a theme also discussed in Bonner and Paegle (1970). Wexler (1961) viewed the LLJ to be

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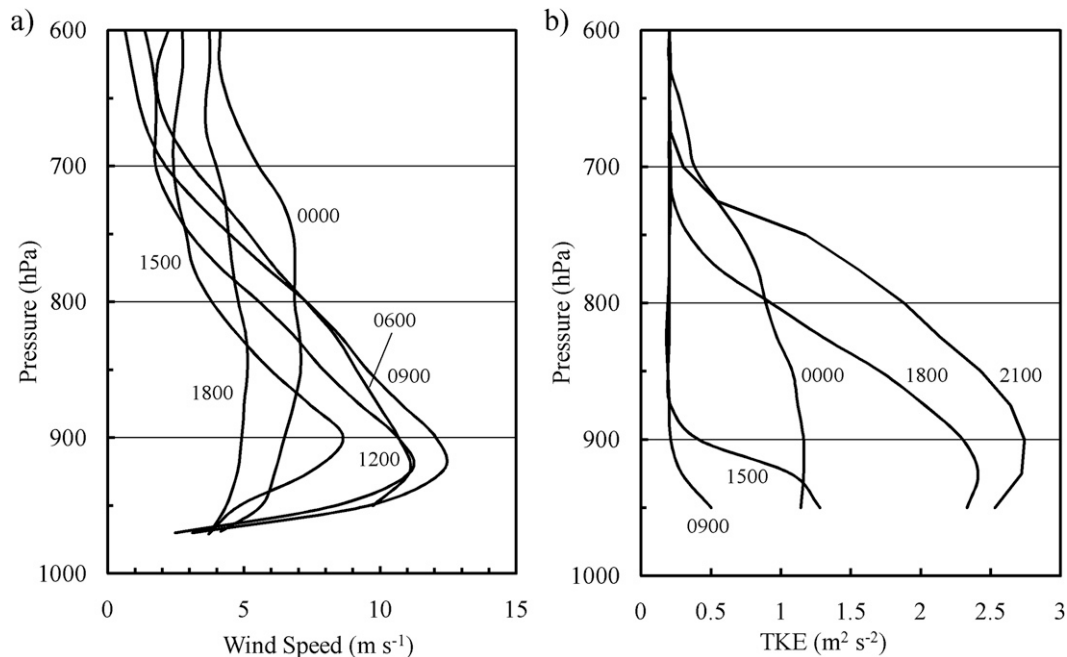


FIG. 1. Mean NAM July 2008 (a) wind speed profile (m s^{-1}) and (b) TKE profile ($\text{m}^2 \text{s}^{-2}$) at various UTC times.

similar to western boundary currents in the ocean, explaining the wind maxima from conservation of potential vorticity of northward-directed flow associated with the Bermuda high. This theme has been echoed in recent papers such as Pan et al. (2004), Ting and Wang (2006), and Jiang et al. (2007). In particular, Ting and Wang (2006) suggest that such mechanical forcing provided by the North American topography is central to the formation of the LLJ and that thermal influences are insignificant, a conclusion that has prompted this note.

It is critical to note that an understanding of the Great Plains LLJ requires answers to two fundamental questions: What factors are responsible for the diurnal variation of the LLJ? Why is the highest summertime LLJ frequency centered over the southern Great Plains? There seems little doubt that the theory of Blackadar (1957) can fundamentally address the first question. The purpose of this note is to address the geographical frequency of the summertime LLJ over the Great Plains. It will be argued here that thermal influences are critical to the observed wind and pressure fields and that there is little to support the claim that North American Cordillera blocking is central to the formation of the mean Great Plains flow.

2. Mean summertime LLJ simulations from the 12-km NAM

An extensive suite of numerical simulations have been conducted as a means to study the LLJ (e.g., Fast and McCorcle 1990; Zhong et al. 1996; Pan et al. 2004; Ting

and Wang 2006; Jiang et al. 2007). Results from the 12-km horizontal resolution Weather Research and Forecasting Nonhydrostatic Mesoscale Model (NAM) for July 2008 are used here to depict the basic structure of the LLJ. Three-hourly output grids are used for each day based on the 0000 UTC forecasts, commencing at 0300 UTC and continuing until 0000 UTC the following day. Inspection of the mean values for July 2008 shows that the NAM is able to simulate the LLJ. As an example, Fig. 1a shows the mean July 2008 LLJ vertical profile at selected times of the day for the grid point corresponding to Enid, Oklahoma, near the climatological center of the LLJ (e.g., Bonner 1968). Mean 925-hPa July 2008 wind speeds over central Oklahoma from the NAM double from 0000 to 0600 UTC and a maximum jet of approximately 13 m s^{-1} occurs around 0900 UTC. Mean July wind directions veer during the night from 159° at 0300 UTC to 209° at 1200 UTC, consistent with an inertial oscillation such as shown in Bonner and Paegle (1970). That frictional decoupling is taking place in the NAM simulations can be confirmed by the mean July 2008 turbulent kinetic energy (TKE) profile (Fig. 1b), which shows rapid changes near sunset. Note that TKE values at 925 hPa near the level of the jet maximum decrease more than a factor of 2 from 2100 to 0000 UTC, suggesting that frictional decoupling occurs in the early evening hours. There seems no doubt that the Blackadar (1957) mechanism is a key factor in the development of the summertime LLJ.

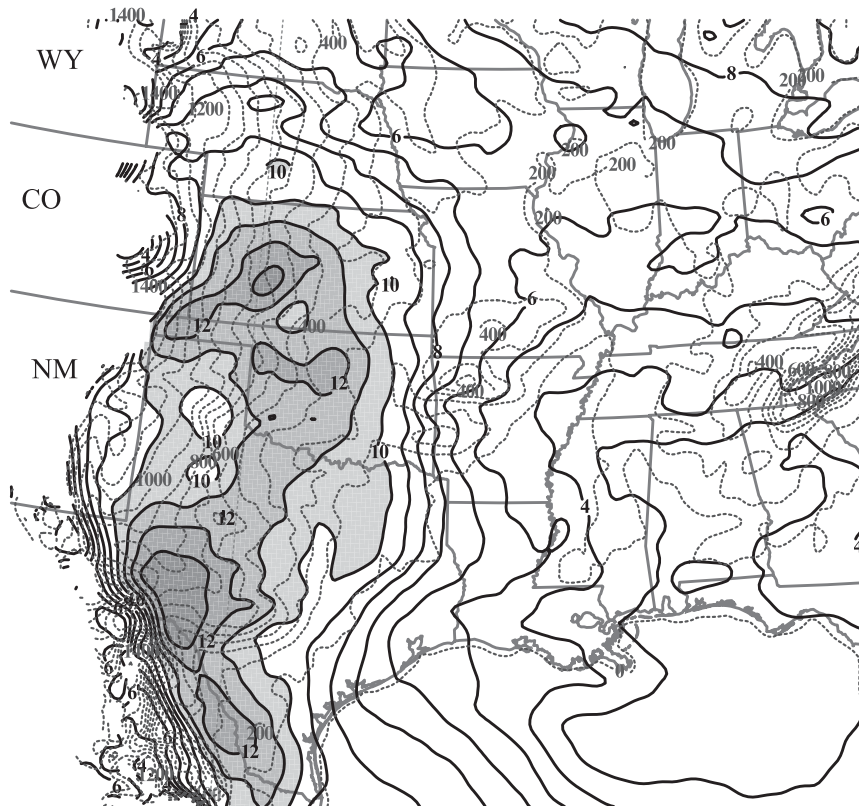


FIG. 2. Maximum mean July 2008 wind speeds (solid lines, m s^{-1}) between surface and 800 hPa at 0900 UTC over the south-central United States from 12-km NAM. Wind speeds in excess of 10 m s^{-1} are indicated by shading. Dashed lines indicate terrain contours (m).

Inspection of the NAM grids shows the LLJ to be a southern Great Plains phenomenon. July 2008 maximum boundary layer winds at 0900 UTC between the surface and 800 hPa from the NAM (Fig. 2) show that the strong LLJ zone extends from southwestern Texas northward to central Kansas. Note that the magnitude of maximum winds decreases rapidly toward the east such that the LLJ core from the July 2008 NAM output does not extend far past the eastern edges of Kansas, Oklahoma, and Texas. Such results are similar to those reported by Bonner (1968) based on rawinsonde data. The LLJ geographical core from the NAM can be seen to coincide with the terrain slope associated with the Great Plains, implying a link between topography and the LLJ that has been noted by many.

Given the acknowledged relationship between sloping terrain and the LLJ, it is appropriate to inquire as to physical mechanisms that produce such a link. Recent studies have examined the role of topography (e.g., Pan et al. 2004; Ting and Wang 2006; Jiang et al. 2007). Results from the NAM for July 2008 have been examined to infer the importance of the diurnal cycle of heating and cooling of the sloping terrain and the impacts on the

wind field. Figure 3 illustrates the mean July 2008 925-hPa x and y components of the wind u and v , respectively; geostrophic wind (subscript g); and ageostrophic wind (subscript ag) at Enid from the NAM. Diurnal increases in the x component of the geostrophic wind of approximately 2 m s^{-1} are present from 0000 to 1200 UTC (Fig. 3a), implying an isallobaric component in the y direction of only 0.5 m s^{-1} . Since the Great Plains slopes upward to the west over central Oklahoma as shown in Fig. 2, no significant terrain-induced diurnal changes in the geostrophic wind component in the x direction should be expected. Larger magnitude oscillations are seen in the x component of the wind and x component of the ageostrophic wind with a period of oscillation roughly that of the inertial period T .

Figure 3b illustrates the July 2008 925-hPa diurnal changes in the y -direction wind components at Enid. Geostrophic wind components in the y direction vary from maximum values of approximately 10 m s^{-1} near 0000 UTC to approximately 5.5 m s^{-1} by 0600 UTC with little subsequent change during the night. Isallobaric wind components during early evening are approximately 2.3 m s^{-1} directed toward the east. Oscillations in the

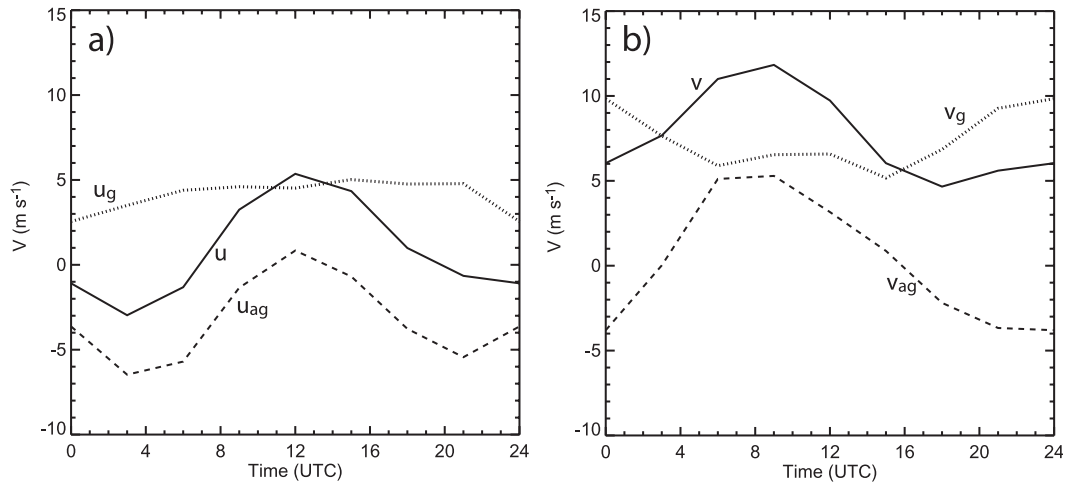


FIG. 3. Mean July 2008 NAM 925-hPa wind components at grid points corresponding to Enid, OK, for the (a) x and (b) y directions.

y -direction wind and ageostrophic components are of comparable magnitude to those in the x direction. Note that the phase angles between the x and y components of the ageostrophic wind are roughly what would be expected from a purely inertial oscillation. Such analyses indicate that the LLJ is supergeostrophic, as noted by Blackadar (1957). Although results shown in Figs. 1 and 3 are taken from one point, analyses from other grid points in the same region yield similar results. From such analyses, it is concluded that the mean summertime diurnal heating and cooling of the Great Plains sloping terrain is responsible for a periodic change in the geostrophic wind component such as reported in Bonner and Paegle (1970) and measured by airborne platform in Parish et al. (1988). This offers further evidence of the validity of the NAM LLJ simulations.

3. Idealized forcing of the Great Plains LLJ

Although informative studies are possible using output from numerical models, a useful description of the LLJ can be obtained using a simple analysis of the equation of motion (e.g., Hess 1959). In absence of friction, and assuming that geostrophic flow exists only in the y direction and varies linearly with time t , the scalar x and y equations of motion can be expressed as

$$\frac{du}{dt} - fv = -f \left[v_g(0) + \frac{\partial v_g}{\partial t} t \right] \quad \text{and} \\ \frac{dv}{dt} + fu = 0.$$

The focus here is on the hours subsequent to frictional decoupling, and the geostrophic wind change with time

is induced by the cooling of the sloping Great Plains terrain. Observations and previous work such as from Bonner and Paegle (1970) as well as results from the NAM for July 2008 shown in Fig. 3 suggest that the amplitude of the geostrophic wind change from late afternoon to dawn is approximately 5 m s^{-1} . Multiplying the y equation by i and adding to the x equation yields

$$\frac{d(u + iv)}{dt} - f(u + iv) = -f \left[v_g(0) + \frac{\partial v_g}{\partial t} t \right].$$

This is a linear ordinary differential equation if it is assumed that the Coriolis parameter f is constant and can be solved by conventional methods. The solution can be shown to be

$$u = u_{\text{isal}} + [u(0) - u_{\text{isal}}] \cos(ft) + [v(0) - v_g(0)] \sin(ft)$$

and

$$v = v_g(t) - [u(0) - u_{\text{isal}}] \sin(ft) + [v(0) - v_g(0)] \cos(ft),$$

where $u_{\text{isal}} = -(1/f)(\partial v_g / \partial t)$ represents the isalobaric wind component.

These two scalar equations contain three components that define how the wind changes with time: the time-varying geostrophic wind that acts in the y direction for conditions specified in the equations of motion above, the isalobaric wind component that for the assumptions listed acts in the x direction, and an inertial oscillation of the initially unbalanced flow. This unbalanced flow consists of the initial imbalance between the u component and the isalobaric component as well as the imbalance between the initial v component of motion and the initial geostrophic wind. Hence the total wind can be

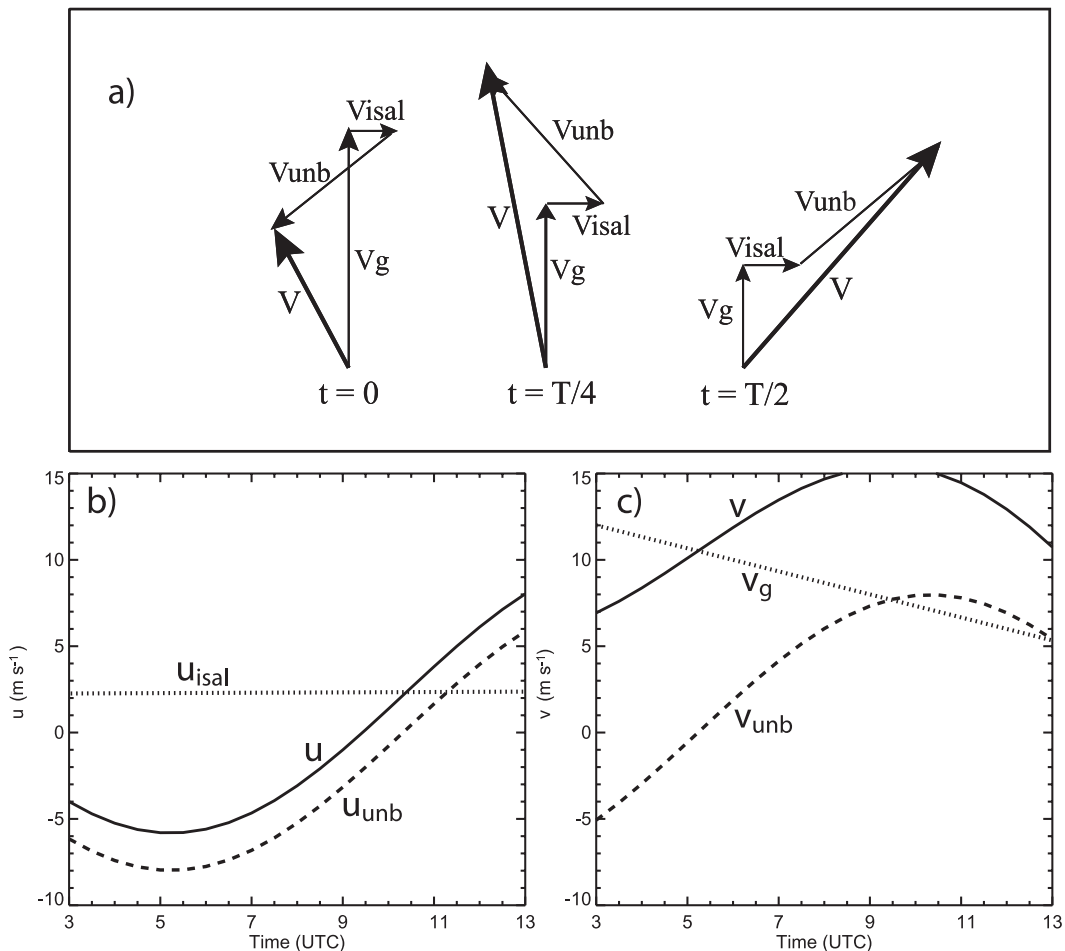


FIG. 4. (a) Schematic depiction of LLJ evolution based on equation system discussed in text. (b), (c) Evolution of the (b) x and (c) y components of motion for the time period in (a).

thought of as the sum of the geostrophic, isalobaric, and inertial components.

In the case of the Great Plains boundary layer, it can be imagined that about the time of sunset the geostrophic wind is near maximum. This is consistent with the NAM analyses presented earlier as well as results from Bonner and Paegle (1970). Subsequent cooling of the sloping terrain forces a relaxation of the pressure gradient in the lowest levels of the atmosphere, equivalent to an isalobaric component that is directed downslope or to the east and the southerly geostrophic component must decrease with time during the night.

A simple example to illustrate consists of an initial geostrophic wind of 12 m s^{-1} at 0300 UTC that decreases to 6 m s^{-1} by 1500 UTC. If it is assumed that the geostrophic wind varies linearly, the isalobaric component is 2.2 m s^{-1} . If it is assumed that the actual wind is 8 m s^{-1} directed 30° to the left of the geostrophic wind at 0300 UTC at the level of the nocturnal jet and that the initial inertial oscillation commences at that time, the

unbalanced component of motion has a magnitude of 8 m s^{-1} . The inertial period at 36°N is roughly 20 h and the initially unbalanced components oscillate 90° in a 5-h period. Assuming the onset of the inertial oscillation occurs at 0300 UTC (2000 LT), the initially unbalanced flow components rotate 180° by 1300 UTC (0600 LT) the next day.

Figure 4a illustrates the evolution of the wind in the lower boundary layer based on the idealized equation system described above for times corresponding to one-quarter and one-half the inertial period. As seen in the NAM analyses, diabatic heating of the sloping Great Plains terrain is responsible for an enhancement to the local pressure gradient force near the time of sunset. This is important since the background flow on which the frictional decoupling operates is at a near maximum. At night, however, cooling of the sloping terrain reduces the magnitude of the geostrophic wind with time. Components of motion accompanying the flow evolution are illustrated in Figs. 4b and 4c. Unbalanced flow components (subscript

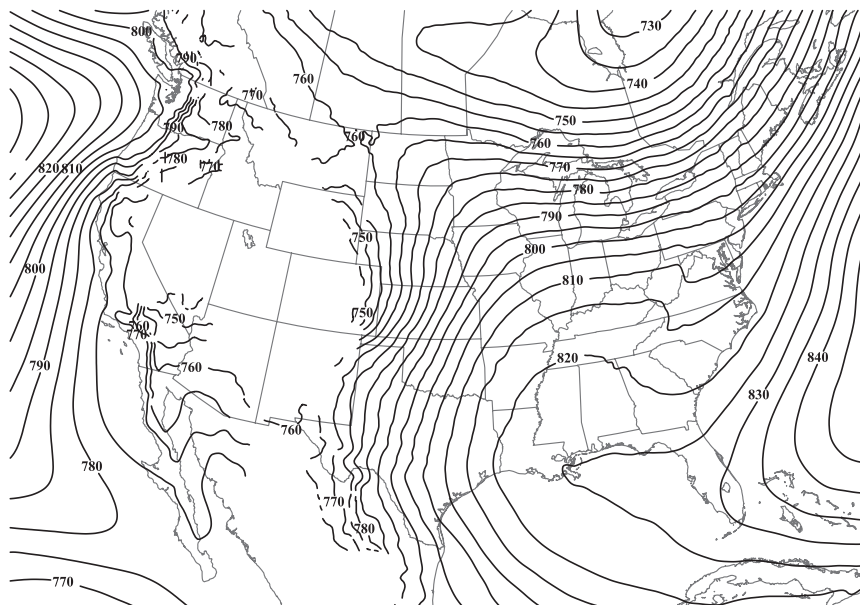


FIG. 5. Mean July 2008 925-hPa height field (m) from 12-km NAM.

unb) from this simple model display a pattern that resembles the ageostrophic wind components shown in Fig. 3.

Sensitivity studies show the importance of the intensity of the background flow on the strength of the LLJ nighttime maximum. The choice of initial conditions shown in Fig. 4 is representative and underscores the importance of decreasing geostrophic flow during the night, which acts as a brake on the intensity of the LLJ maximum. Calculations in which the isobaric wind component is reduced and hence the geostrophic wind decrease is smaller consistently show a greater LLJ intensity. The strength of the mean flow is closely tied to the maximum strength of the LLJ and is thus a significant factor. Based on the simple example shown in Fig. 4 and sensitivity studies from the equation system above, it is concluded here that the effect of nocturnal diabatic cooling of the sloping terrain is to reduce the magnitude of the LLJ maximum. Thus it is argued from this simple analysis that the diabatic heating and cooling of the sloping Great Plains terrain has a mixed effect on the intensity of the nocturnal LLJ. Afternoon heating enhances the intensity of the base geostrophic flow on which frictional decoupling acts in creating the nocturnal LLJ, but the cooling process during the night serves to reduce the magnitude of the geostrophic wind and hence the intensity of the LLJ maximum.

4. Geographic location for the summertime LLJ

It is relevant to inquire as to why the LLJ displays a frequency of occurrence maxima over the Great Plains.

Numerical results of Pan et al. (2004), Ting and Wang (2006), and Jiang et al. (2007) stress the importance of the North American Cordillera in the development of the summertime LLJ. These simulations show that removal of the North American mountain chain leads to a significant reduction in the LLJ frequency. The existence of a strong background mean flow—the geostrophic wind component in the above equation set—is important for development of a strong nighttime LLJ. Not surprisingly, observations and analyses using output from the NAM for the summer of 2008 (not shown) indicate a close relationship between the magnitude of the nocturnal summertime LLJ maximum and the strength of the southerly component of the geostrophic wind. The importance of the mean flow on the LLJ has been pointed out by many including the recent work of Pan et al. (2004), Ting and Wang (2006), and Jiang et al. (2007). Wexler (1961) and Zhong et al. (1996) note that the existence of a strong background flow is critical in the development of the LLJ since it provides the base state on which the diurnal frictional effects can operate.

Figure 5 illustrates the mean 925-hPa level geopotential height field for the Great Plains region taken from the July 2008 NAM. Prominent on the map are the broad anticyclonic circulations associated with the Bermuda high over the eastern United States and with the Pacific high over the western United States. A maximum in the mean July 2008 geostrophic wind magnitude, as implied by the gradient of the height contours, is seen over the Great Plains, stretching from southern Texas to northern Kansas. Intensification of the low-level flow associated

with the Bermuda high is present along the eastern slopes of the North American Cordillera from Mexico to the Northern Plains. Enhancement of the horizontal pressure gradient force occurs over the region of sloping terrain, again suggesting a link between the Great Plains topography and the forcing of the mean flow. Similar results have been shown in Jiang et al. (2007) based on results of the North American Regional Reanalysis grids and output from the GFDL global atmosphere and land model. Given the acknowledged importance of the mean flow, it can be concluded with little surprise that the frequency of occurrence of the LLJ over the Great Plains is in part due to the mean summertime low-level circulation associated with the Bermuda high. The decrease in LLJ intensity and frequency reported by Pan et al. (2004), Ting and Wang (2006), and Jiang et al. (2007) for their simulations in which the North American Cordillera is removed is tied to the mean flow, which is reduced considerably in response to the flat terrain in their simulations.

What mechanisms are responsible for the enhancement of the horizontal pressure gradient force over the Great Plains as shown in Fig. 5? A central theme echoed by numerous authors is tied to the physical blocking of the flow at the western end of the Bermuda high by the Great Plains topography (e.g., Wexler 1961). The North American Cordillera serves as a barrier to the flow, which becomes deflected northward (Wexler 1961). Strong anticyclonic shear is thought to develop as the flow moves northward parallel to the North American topography, resulting in an enhanced mean flow in which the LLJ evolves. This idea stresses conservation of potential vorticity with attendant mechanical generation of anticyclonic vorticity to explain the strong low-level winds over the Great Plains. Ting and Wang (2006) have concluded that such mechanical forcing is dominant for the LLJ and that the role of thermal forcing is negligible. It is suggested here that the 12-km NAM horizontal pressure and temperature fields support neither the theory discussed by Wexler (1961) nor the importance of blocking such as discussed in Ting and Wang (2006) and others. Further, it is argued that thermal effects play a central role in the mean low-level southerly flow across the Great Plains during summer, in addition to the diurnal LLJ characteristics that have been noted by many, including Jiang et al. (2007).

It has repeatedly been stated (e.g., Wexler 1961) that the North American Cordillera from the high plateau in central Mexico to the Rocky Mountains of the United States serves as a “blocking” mechanism or “barrier” for the flow associated with the Bermuda high. Such a description does not fit with observations of the horizontal pressure field in the lower atmosphere over the

Great Plains such as shown in Fig. 5. If stably stratified flow is blocked by terrain, the speed of the low-level airflow impinging on the barrier must decrease as the air is forced to ascend the terrain. Low-level convergence on the eastern slopes of the Great Plains must result and a piling up of flow with an attendant pressure increase must occur. As shown in Fig. 5, an enhancement of the mean horizontal pressure gradient develops for July 2008 over the sloping terrain with the *lowest* pressure being situated to the west. The Great Plains mean flow shown in Fig. 2 thus is not an example of blocked flow.

Schwerdtfeger (1975) wrote the seminal paper on the upstream effect of a mountain barrier on impinging flow. Since that time descriptions of flow directed toward a topographic barrier have gone by the names of “cold air damming,” “barrier flow,” or, as used here, “blocking.” In each case the relevant parameters are the static stability of the airstream, height of the terrain obstacle, and initial kinetic energy of the wind; hence, the Froude number is a critical nondimensional number used to describe such flow situations (e.g., Pierrehumbert and Wyman 1985). Characteristics of blocking include modulation of the horizontal pressure field along the windward slopes such that a positive pressure perturbation becomes established. If the situation persists for periods of several hours, mountain-parallel or barrier wind components can develop with the highest terrain to the right of the flow in the Northern Hemisphere (e.g., Olson et al. 2007).

In addition to the horizontal pressure field being incompatible with blocking, the robustness of potential vorticity conservation in the summertime boundary layer is questionable. Wexler (1961) notes that the summertime lower atmosphere is well mixed for at least 9 h day^{-1} , a claim validated by the 12-km NAM analysis, and thus potential vorticity conservation in the lower atmosphere becomes suspect. Finally, numerical experiments by Zhong et al. (1996) have shown that the variation in the Coriolis parameter, a key argument in the Wexler (1961) theory, plays only a minor role in the strength of the LLJ. It is concluded here that the mean summertime flow in the lowest 2 km or so over the Great Plains cannot be the result of mechanical forcing processes.

To understand the effect of the sloping Great Plains on the mean horizontal pressure field, heating of the sloping terrain must be considered. As shown in Jiang et al. (2007), perturbations develop in the geopotential height field with lower perturbation heights adjacent to the terrain. The mean height field in the lower atmosphere (shown in Fig. 5), as well as the diurnal oscillation of the horizontal pressure gradient force, is tied to the heating of the sloping Great Plains.

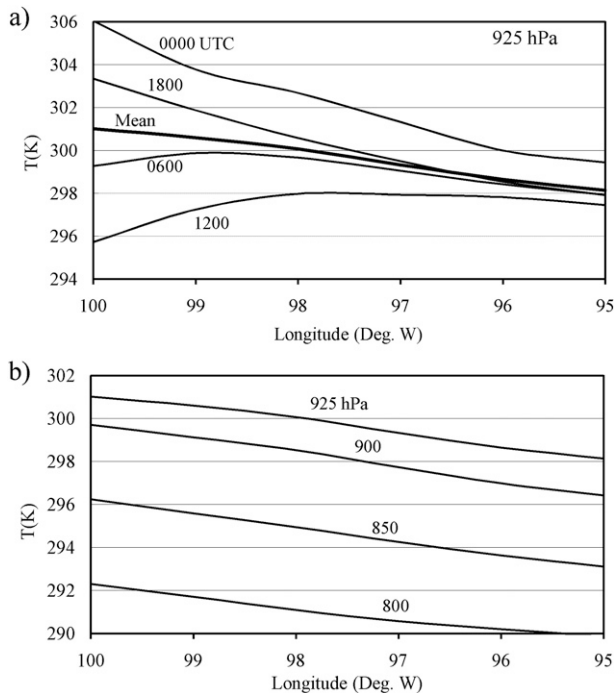


FIG. 6. Mean July 2008 12-km NAM (a) mean 925-hPa temperatures (K) along 100°–95°W, averaged between 35° and 39°N at 6-h intervals, and (b) isobaric temperatures (K) at 925, 900, 850, and 800 hPa along 100°–95°W, averaged between 35° and 39°N.

Diurnal changes in temperatures are most pronounced in the lower boundary layer below approximately 900 hPa where the sign of the temperature gradient between 95° and 100°W over Oklahoma and southern Kansas can change because of nighttime cooling of the terrain slopes. Bonner and Paegle (1970) also note that the strongest geostrophic wind oscillation associated with diabatic heating of the sloping terrain is found at the surface. Figure 6a shows the evolution of the mean July 2008 temperature gradient at 925 hPa to illustrate that from local midnight to dawn cooler temperatures are seen in the mean between approximately 97° and 100°W, which is consistent, of course, with the diurnal changes in the geostrophic wind.

For the mean monthly 925-hPa geostrophic wind conditions, the pattern of heating over the sloping terrain is critical. Figure 6b illustrates the mean July 2008 NAM temperatures at several isobaric levels in the lower atmosphere over the LLJ region encompassing a rectangle centered over Oklahoma and southern Kansas between 35° and 39°N and between 95° and 100°W. Temperature gradients exist over the Great Plains in the July 2008 mean with warmest temperatures to the west over the elevated terrain. It is impressive that mean monthly isobaric temperatures show warmest air over the elevated

terrain for all levels below 500 hPa. Isobaric temperature gradients below 850 hPa average approximately $0.7^{\circ}\text{C} (100 \text{ km})^{-1}$ from the July 2008 NAM. A thermal wind from north to south is implied by the temperature field, which in turn implies an increase of the southerly geostrophic wind below 850 hPa and a strengthening of the horizontal pressure gradient force. An enhancement of the mean horizontal pressure field over the Great Plains sloping terrain at low levels such as shown in Fig. 5 is consistent with the thermal wind.

Inspection of the mean July 2008 NAM conditions over the central Great Plains shows that the geostrophic wind and hence horizontal pressure gradient is nearly zero at 600 hPa (not shown). Given the horizontal temperature field as shown in Fig. 6 and the attendant thermal wind, a southerly geostrophic wind must become established below 600 hPa. Since the horizontal temperature gradient is strongest at levels between the surface and 850 hPa, the most rapid change will take place in the lower atmosphere. The intensity of the southerly geostrophic wind will increase with distance below the 600-hPa level, with the maximum being reached at the ground. Thermal wind calculations from 600 to 1000 hPa over the southern Great Plains averaged for the entire month of the July 2008 suggest an approximately 8 m s^{-1} enhancement of the southerly flow associated with the Bermuda high, which is consistent with the tightening of the height field shown in Fig. 5 over the sloping Great Plains terrain.

Thermal wind vectors computed from the vector difference of geostrophic winds between 750 and 950 hPa superimposed on the mean horizontal temperature gradient at 850 hPa from the NAM (Fig. 7) are directed parallel to the mean isotherms, indicative of thermal wind balance. Magnitudes of the thermal wind vectors shown in Fig. 7 can be compared with thermal wind calculations from the mean horizontal temperature gradients. It can be seen that typical horizontal temperature gradients between 750 and 950 hPa are approximately $0.8 \text{ K} (100 \text{ km})^{-1}$, which corresponds to a thermal wind of approximately 6.5 m s^{-1} , representative of the magnitudes of vectors shown in Fig. 7. The horizontal pressure gradient force is thus close to thermal wind balance. It is concluded that the sloping terrain gives rise to heating differences during summer that are fundamental to the enhancement of the mean low-level horizontal pressure gradient and hence the background flow over the Great Plains. Weak isobaric temperature gradients exist to the east of the Great Plains and consequently minimal thermal wind enhancement occurs. Hence the mean horizontal pressure field at 1000 hPa associated with the Bermuda high east of approximately 94°W is weak.

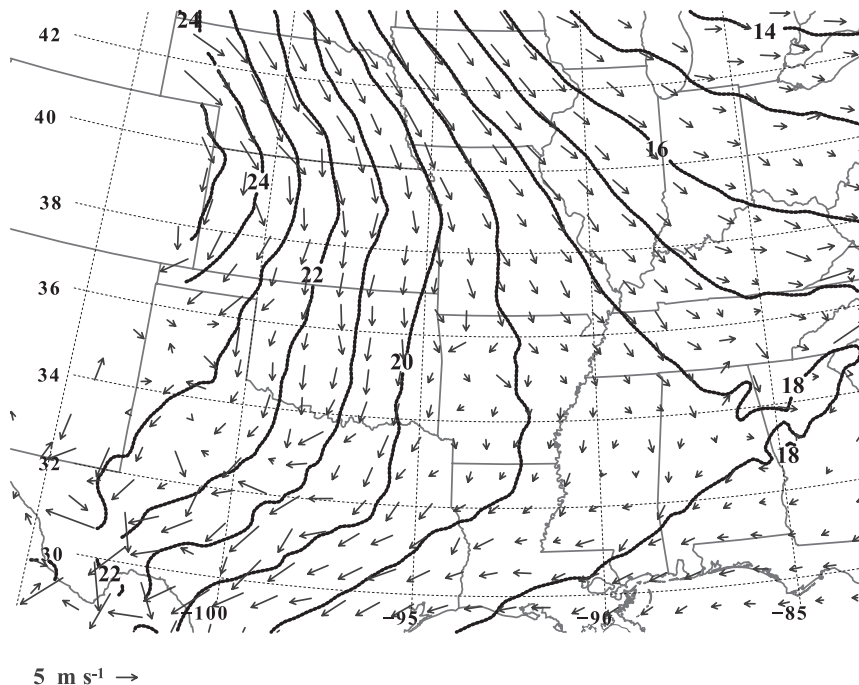


FIG. 7. Mean 850-hPa isotherms (solid lines, °C) and thermal wind vectors between 750 and 950 hPa (arrows, m s^{-1}) from July 2008 12-km NAM.

An analogy to the Great Plains pressure field is that associated with the strong summertime low-level jet stream found off the California coast. Intensification of the horizontal pressure gradient on the eastern edge of the Pacific high along the West Coast occurs because of the heating of the mountainous coastal range and interior valley of California. Strong horizontal temperature gradients develop that, through thermal wind arguments, result in an enhancement of the low-level horizontal pressure gradient force near the coast (e.g., Zemba and Friehe 1987). The heating acts to intensify existing horizontal pressure gradients in both cases and draw strongest geostrophic winds toward the heat source.

5. Summary

Analyses using simplified forms of the equation of motion indicate that the Great Plains LLJ occurs because of the combined effects of a mean geostrophic component, an isobaric component, and an unbalanced or inertia component that arises when turbulent fluxes in the boundary layer diminish. It has been argued that a strong mean flow is necessary for the development of the nocturnal Great Plains LLJ maximum. It is noted here that the mean summertime low-level pressure field is enhanced as the result of the mean summertime heating of the Great Plains terrain. The notion that the North American Cordillera blocks the flow is not compatible

with observations of the pressure field. Summertime heating of the elevated sloping terrain is responsible for the establishment of a mean isobaric temperature gradient, which (through thermal wind arguments) implies an enhancement of the horizontal pressure gradient force associated with the Bermuda high over the Great Plains. The geographical frequency of the LLJ over the southern Great Plains is tied to the strengthened geostrophic wind; thus, thermal forcing is critical to enhancing the mean flow in the lowest two kilometers above the surface.

The diurnal course of daytime heating and nighttime cooling of the Great Plains sloping terrain is responsible for an oscillation of the background geostrophic wind, as noted by Holton (1967), Bonner and Paegle (1970), and others. That the maximum geostrophic wind is found near the time of the onset of frictional decoupling enhances the development of the LLJ nighttime maximum, although the nighttime cooling of the sloping terrain decreases the geostrophic flow. Since at any time the wind can be thought of as the sum of the geostrophic and ageostrophic components, such nighttime diabatic forcing acts to reduce the actual wind. In this manner, the sloping terrain plays a key role in the establishment of the mean flow but plays an adverse role in the nighttime maximum intensity of the LLJ. It is again noted here that the mechanism described by Blackadar (1957) is responsible for the nocturnal wind maxima.

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