MESOSCALE METEOROLOGY

3.5 Nocturnal Low-Level Jet

The broadest definition of a low-level jet (LLJ) is simply any lower-tropospheric maximum in the vertical profile of the horizontal winds.

A LLJ can occur under favorable synoptic conditions anywhere in the world.

Of practical interest is their impact on the transport of moisture. Of theoretical interest is the large amount of vertical wind shear associated with them and the observation that they are typically supergeostrophic by a large ($> 50\%$) amount.

A large number of specific geographic locations all over the world have been identified as especially favorable for LLJ development. Among these locations is the Great Plains region of the United States. Here, the LLJ is one of the most significant in terms of its impact on precipitation and severe weather.

Specifically, LLJs often lead to the initiation of long-lived mesoscale convective complexes that produce heavy overnight rains and sometimes flash floods. These LLJs also often increase the low-level vertical wind shear, which may enhance the potential for severe convective storms, including tornadoes. The nocturnal low-level wind maximum is also believed to be responsible for the observed nighttime maxima in warm-season rainfall accumulations in the central United States

The Great Plains LLJ has consequently been studied more than any other.

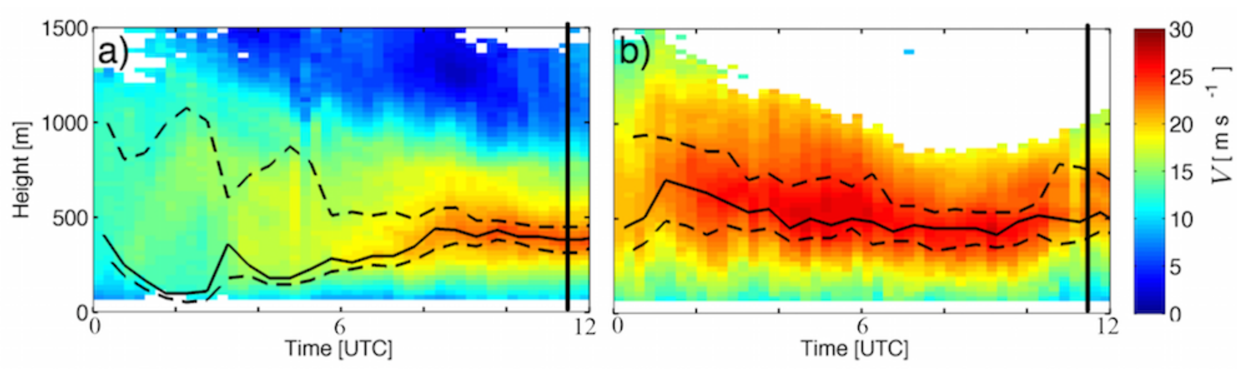

Figure 1: Evolution of the LLJ at the Lamont, OK ARM site on the nights of 09 October (left) and 24 October 2012 (right) case studies. [From Klein *et al.* 2015]

3.5.1 Introduction

Definition

- Fast moving current of air near the surface.
- Large wind shear ($\partial\vec{V}/\partial z$) above and below the jet level.
- Maximum wind speed at least 12-16 ms^{-1} (peak speeds of $\sim 30\ \mathrm{ms}^{-1}$ observed)
- Wind speed above jet is typically 50-75$\%$ or less of the maximum.
- Strong lateral shear on both sides, with a typical width of $\sim$ 200-300 km.

3.5.2 Climatology of LLJs

In the U.S., the LLJ most common flows from the south over the Great Plains, especially in the spring and summer. LLJs are also found around the world, such as the Koorin Jet and Southerly Buster in Australia, or the Somali Jet in East Africa.

For the U.S. Great Plains Jet

- Strong diurnal oscillation with strongest wind speeds at night.
- Average height 500-1000 m AGL, near the level of nocturnal inversion.
- Often the maximum winds are supergeostrophic.

In the climatology study of Bonner (1968), it was found that

- LLJ tended to have a wind maximum near 800 m AGL.
- Strong jets were primarily a nighttime feature.
- LLJ occurred with greatest frequency during the spring and summer.
- The states of greatest activity include: Texas, Oklahoma, Kansas, Nebraska, Iowa, Missouri, and Arkansas.
- Favorable synoptic conditions for LLJ formation are those which have a strong west to east pressure gradient across the Great Plains and an uninterrupted flow of air from the Gulf of Mexico.

A more recent study by Whiteman *et al.* (1997) showed that

- 50$\%$ of LLJ maxima actually occur below 500 m
- The temporal wind maximum typically occurs around 2 am LST (local standard time).

3.5.3 Meteorological Importance of LLJs

- Increased northward transport of moisture at jet level.
- Increased low-level convergence at nose of jet.
- Involved in sustaining convection at night.
- Partly responsible for nighttime thunderstorm maximum observed in the Great Plains.

The low altitude and southerly flow of the LLJ make it a key element in the return-flow cycle of air from the Gulf of Mexico (a typically springtime event). In this cycle, northerly flow advects dry, typically cool continental air out over the Gulf of Mexico where it is modified by surface processes and gains moisture. To complete the return-flow cycle, this modified air then advects northward back onto the continent by way of low-level winds. The LLJ is a principal mechanism by which this moist and unstable air from the Gulf is advected northward into the United States where it ultimately becomes precipitation.

Higgins *et al.* (1997) found that low-level flow of moisture from the Gulf of Mexico at night is increased by 48% from mean values when a LLJ is present.

Arritt *et al.* (1997) showed that the widespread Great Plains flooding event of 1993 was associated with a prolonged period of strong LLJs.

The LLJ can also promote convection by inducing uplifting from convergence along the nose of the jet (Zhong et al. 1996) which can combine with divergence aloft from an upper-level jet (Beebe and Bates 1955).

The strong nocturnal phase of the jet is widely believed to be particularly important in promoting nighttime convection.

The LLJ has been linked to the occurrence and intensity of mesoscale convective systems and appears to be an essential ingredient in the environment that produces mesoscale convective complexes. This is due presumably to the enhancement of both warm advection and the advection of moist, unstable air (Maddox 1983).

Further, the presence of a LLJ, especially when combined with an upper-level jet, provides a veering of winds with height that is favorable for the development of severe weather and tornadoes (Uccellini and Johnson 1979).

3.5.4 Causes of LLJs

- Inertial oscillation.
- Baroclinicity over sloping terrain.
- Coupling with return circulation in the jet streak, and others related to synoptic-scale dynamics.

We will consider first two causes only here, as they relate to boundary layer dynamics.

The Inertial Oscillation

One of the perhaps most important theories for understanding the LLJ was due to Blackadar (1957).

This theory accounts for both the daily oscillation in jet intensity and for the significantly supergeostrophic velocities observed during the nocturnal phase.

Blackadar explains that:

- The cycle of the LLJ as an inertial oscillation that relies on the retardation to subgeostrophic speeds of lower tropospheric air due to vertical, turbulent mixing with the heated surface during the day.
- Once surface heating ceases near nightfall, the layer of air in contact with the ground undergoes radiative cooling, becomes statically stable, and decouples from the layer of air above.
- This layer becomes nearly frictionless and turbulence free, causing the flow to accelerate due to the synoptic pressure gradient.
- The effect of the Coriolis force on this accelerating, frictionless air stream is to cause an inertial oscillation with supergeostrophic speeds being reached after several hours.

Consider the classic example of an oscillating system, a pendulum. The stable (or balanced) position of a pendulum is pointing straight down. A push of the pendulum to one side will cause it to return to the balance point, but its momentum will carry it past that point, and it will swing up on the other side.

Now consider the "balance point" for atmospheric motion: *geostrophy*. Frictional drag keeps the wind below geostrophy in the mixed PBL during the day. At night, the stable layer reduces drag to only the lowest tens of meters. Air above this shallow layer accelerates as friction is "released".

Mathematically, we begin with the horizontal momentum equations in which Coriolis force is considered. Friction, on the other hand, is not included since we are considering the case when the flow is decoupled form the surface.

$$\frac{du}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv \tag{1}$$

$$\frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu \tag{2}$$

Next, we partition the flow into geostrophic (g) and ageostrophic (a) components

$$u = u_g + u_a \tag{3}$$

$$v = v_g + v_a \tag{4}$$

We substitute the above into Eqs.(1) and (2), and assume that $d\vec{v_g}/dt = 0$ (in other words we assume that the horizontal pressure gradient is constant in time), which yields

$$\frac{du_a}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + f(v_g + v_a) \tag{5}$$

$$\frac{dv_a}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - f(u_g + u_a) \tag{6}$$

By definition

$$u_g = -\frac{1}{\rho f}\frac{\partial p}{\partial y} \quad \text{and} \quad v_g = \frac{1}{\rho f}\frac{\partial p}{\partial x} \,. \tag{7}$$

Substitution into Eqs.(5) and (6) yields

$$\frac{du_a}{dt} = f v_a \tag{8}$$

$$\frac{dv_a}{dt} = -f u_a \tag{9}$$

Next, we take d/dt of Eq.(8) and use Eq.(9) to obtain

$$\frac{d^2 u_a}{dt^2} = f\frac{dv_a}{dt} = -f^2 u_a \tag{10}$$

If we take the f-plane approximation (f is constant), then Eq.(10) is a linear ODE with constant coefficients.

$$u_a = C_1 \cos\left(ft\right) + C_2 \sin\left(ft\right) \,, \tag{11}$$

where C_1 and C_2 are constants. Also note that using Eq.(11) together with Eq.(8) will yield

$$\begin{aligned} v_a &= \frac{1}{f}\frac{du_a}{dt} \\ &= \frac{1}{f}[-C_1 f \sin\left(ft\right)] + \frac{1}{f}[C_2 f \cos\left(ft\right)] \\ &= -C_1 \sin\left(ft\right) + C_2 \cos\left(ft\right) \end{aligned} \tag{12}$$

At $t = 0$, we define $u_a = u_{a0}$ and $v_a = v_{a0}$. Therefore,

$$u_{a0} = C_1 \cos 0 + C_2 \sin 0 = C_1 \tag{13}$$

$$v_{a0} = -C_1 \sin 0 + C_2 \cos 0 = C_2 \tag{14}$$

Substitution of Eqs.(13) and (14) into Eqs.(11) and (12) yields

$$u_a(t) = u_{a0} \cos\left(ft\right) + v_{a0} \sin\left(ft\right) \tag{15}$$

$$v_a(t) = v_{a0} \cos\left(ft\right) - u_{a0} \sin\left(ft\right) \tag{16}$$

We can apply trigonometric identities (this is left as an exercise for you, see page 107 in Markowski and Richardson), we rewrite Eqs.(15) and (16) as

$$u_a(t) = |\vec{v}_{a0}| \cos\left(\psi_0 - ft\right) \tag{17}$$

$$v_a(t) = |\vec{v}_{a0}| \sin\left(\psi_0 - ft\right) \,, \tag{18}$$

where $\vec{v}_{a0} = (u_{a0}, v_{a0})$, $|\vec{v}_{a0}| = \sqrt{u_{a0}^2 + v_{a0}^2}$, and ψ_0 is the the angular constant designating the initial orientation of the ageostrophic wind.

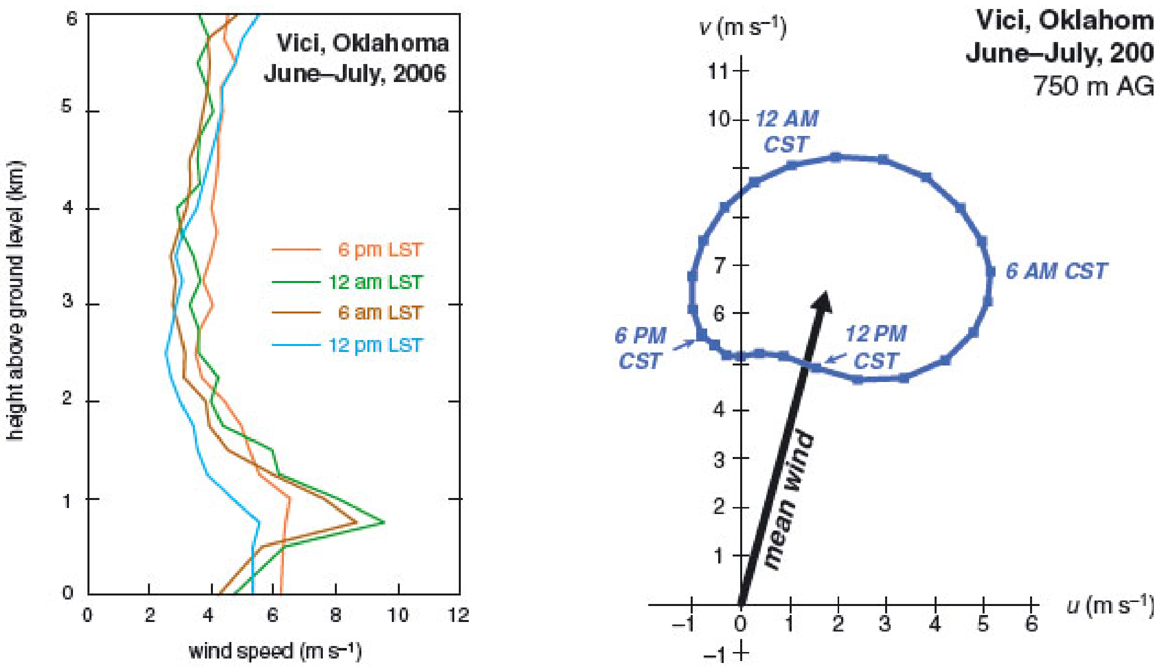

Figure 2: (left) Vertical profiles of the mean horizontal wind speed at the Vici, OK, wind profiler at 6 pm, 12 am, 6 am, and 12 pm local standard time in June and July, 2006; (right) Mean diurnal cycle of the horizontal wind velocity components at 750m above ground level observed by the Vici, OK, wind profiler in June and July, 2006. [From Markowski and Richardson]

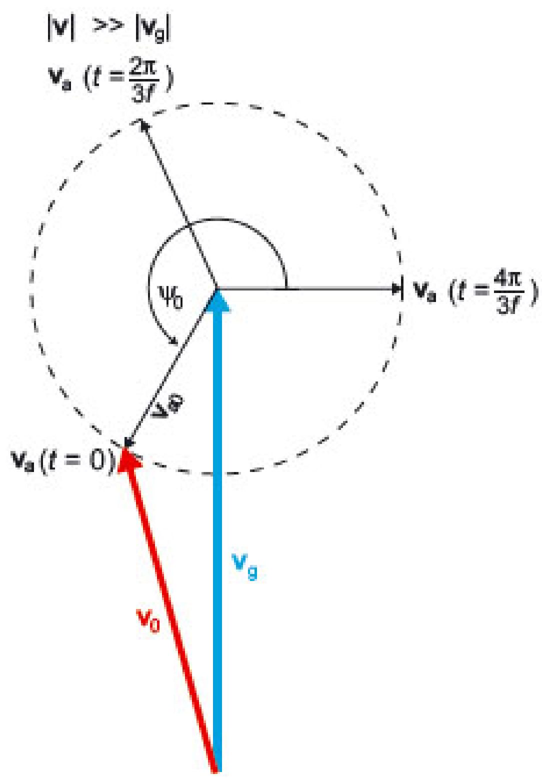

Figure 3: Schematic illustrating $\vec{v}_g$, $\vec{v}_a(t)$,$\vec{v}_{a0}$, and ψ_0. The horizontal wind is shown at $t = 0(\vec{v}_0)$. [From Markowski and Richardson]

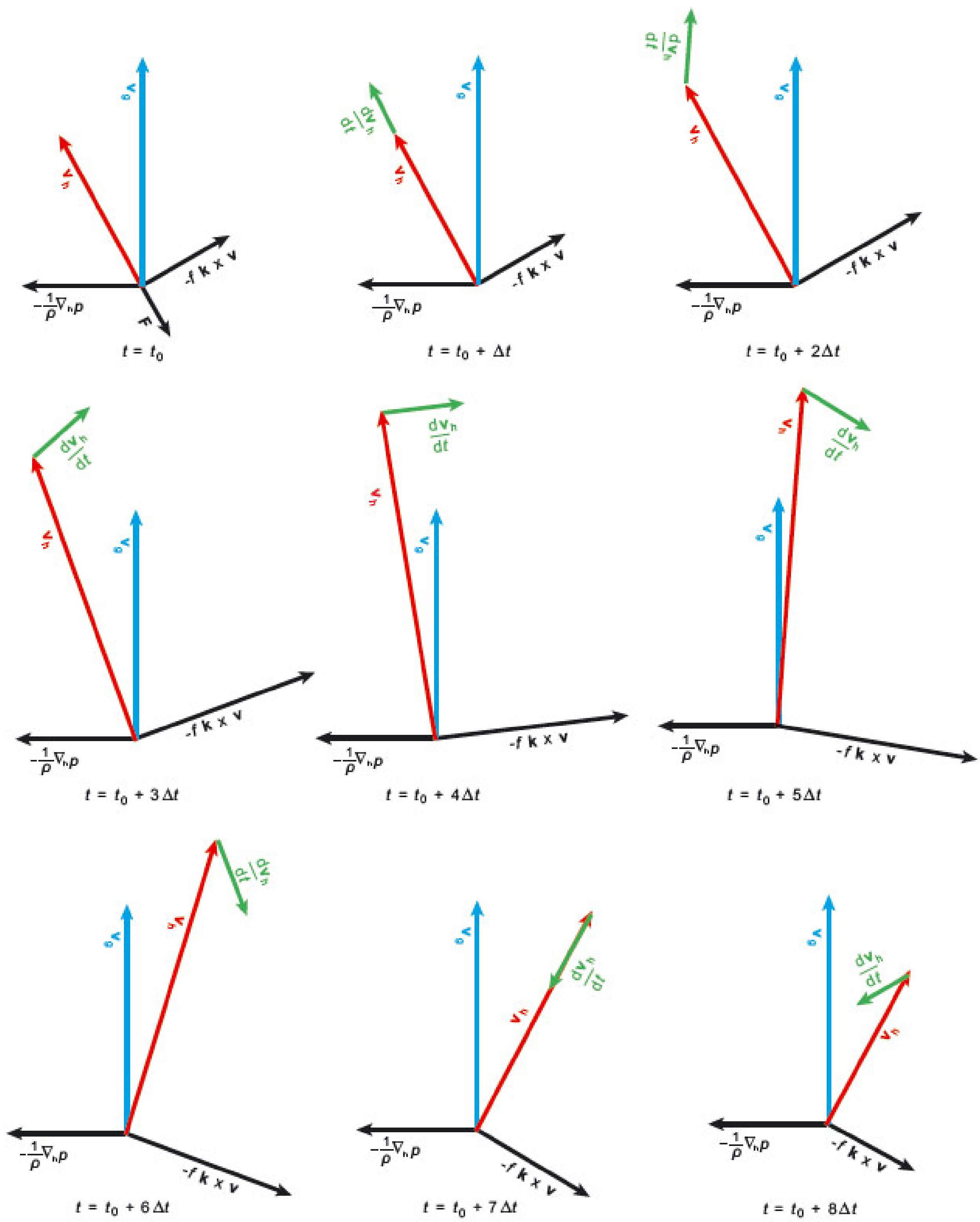

Figure 4: Schematic illustrating how the nocturnal low-level wind maximum develops, for the case of a southerly geostrophic wind. [From Markowski and Richardson]

Equations (17) and (18) show that the ageostrophic wind rotates clockwise (in the northern hemisphere) in a circular pattern (called an inertial circle). This behavior starts after turbulence shuts down and the boundary layer decouples from the surface, usually ~6-7 pm local time.

Figure 3 shows that $\vec{v}_a$ will point north after a period of approximately $5\pi/6$. This means that the horizontal wind reaches a maximum at that time since it is supergeostrophic ($\vec{v}_h = \vec{v}_g + \vec{v}_a$).

A full circle is completed in $2\pi/f$. Recall that $f = 2\omega \sin(\phi)$, where ϕ is latitude and $\omega = 2\pi/24\ \mathrm{h}$ is the Earth's angular frequency. Thus, the LLJ will reach its maximum in Oklahoma ($\phi \approx 35°$) after

$$\frac{5}{6}\frac{\pi}{f} = \frac{5}{6}\frac{\pi}{4\pi}\frac{24\ \mathrm{h}}{\sin(\phi)} = \frac{5\ \mathrm{h}}{\sin(35°)} \approx 9\ \mathrm{h}.$$

That means if the adjustment period starts around 6-7 pm local time, the LLJ will reach its peak strength at ~2-3 am local time.

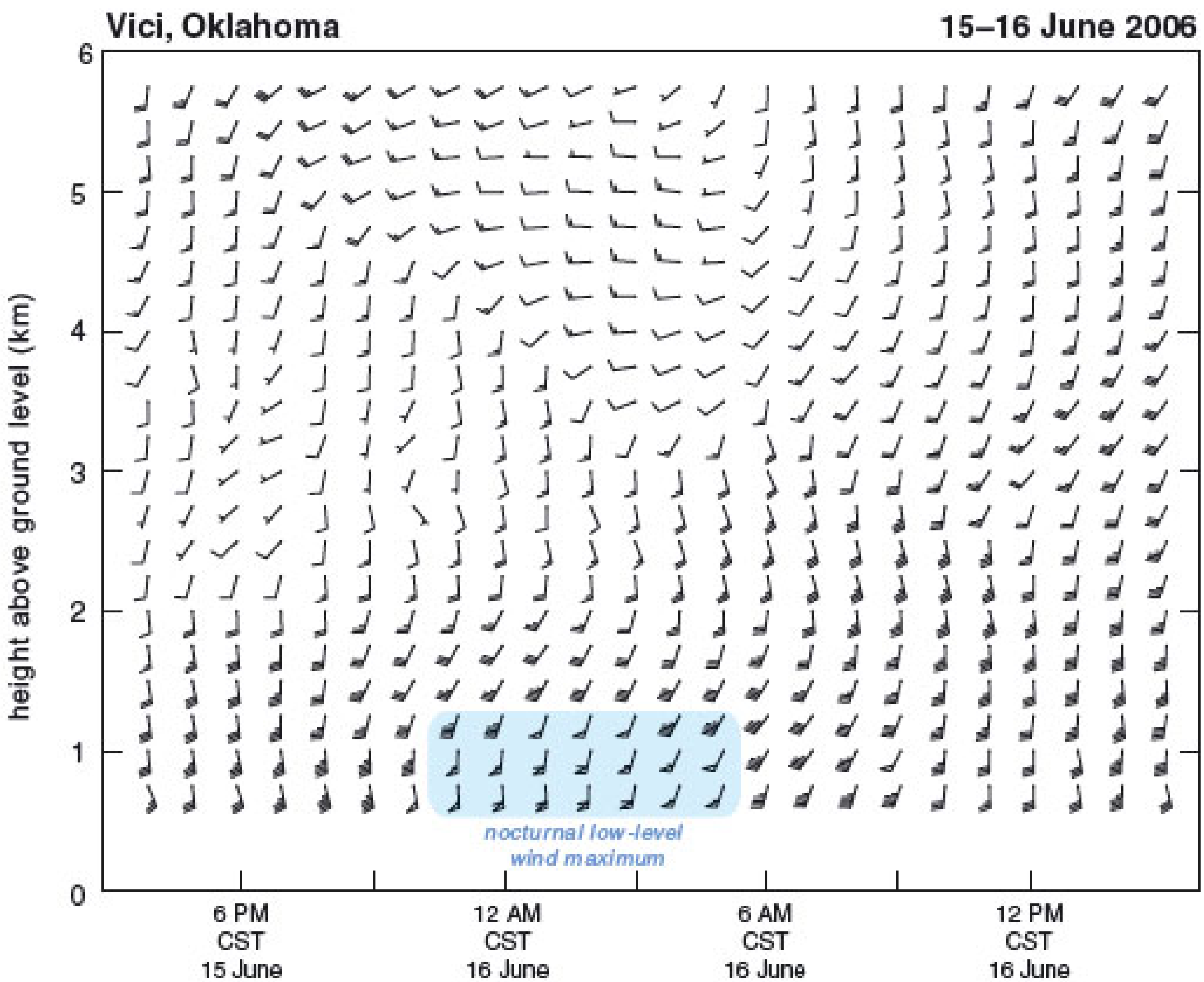

Figure 5: Vertical wind profiles observed by the Vici, OK, wind profiler from 4 pm CST 15 June to 3 pm CST 16 June 2006. [From Markowski and Richardson]

Baroclinicity Over Sloping Terrain

Theories other than the Blackadar theory have been proposed to account in whole or part for the LLJ. One mechanism analyzed by Holton (1967) describes the nature of the LLJ as a response to the diurnal heating and cooling of sloping terrain, which results in a periodic variation in thermal wind and a consequent surface geostrophic wind oscillation.

This mechanism makes no appeal to variations in turbulent mixing and has the advantage of explaining why the LLJ tends to be located over the (gently sloped) Great Plains, which the Blackadar theory does not address.

Let's recall the thermal wind relation

$$\frac{\partial u_g}{\partial z} = -\frac{g}{f\overline{T}}\frac{\partial \overline{T}}{\partial y} \tag{19}$$

$$\frac{\partial v_g}{\partial z} = \frac{g}{f\overline{T}}\frac{\partial \overline{T}}{\partial x}\,, \tag{20}$$

where $\overline{T}$ is the mean layer temperature.

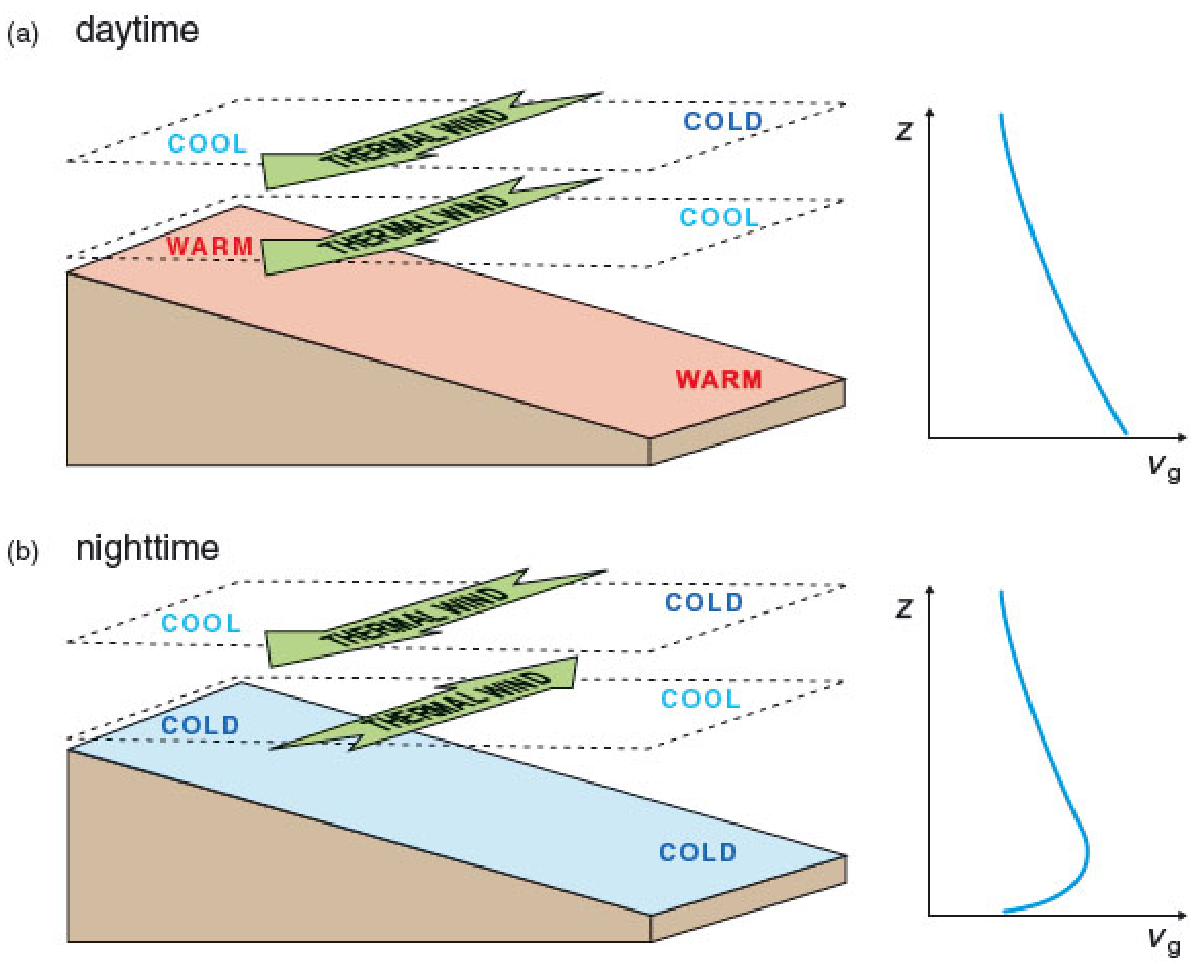

Figure 6: Diurnal oscillation of the low-level thermal wind owing to the diurnal heating and cooling of sloped terrain. [From Markowski and Richardson]

During the **daytime** (Fig.6a), solar insolation warms ground and forms a mixed layer with a near-adiabatic lapse rate such that the west-to-east temperature gradient (measured along a horizontal surface) is negative ($\partial\overline{T}/\partial x < 0$) near the ground and aloft. That means $\partial v_g/\partial z < 0$.

During the **nighttime** (Fig.6b), the ground cools more quickly than the overlying air. This results in a reversal of the horizontal temperature gradient ($\partial\overline{T}/\partial x > 0$). Thus, $\partial v_g/\partial z > 0$. Above the inversion level, the gradient may again reverse ($\partial\overline{T}/\partial x < 0$), as does the thermal wind $\partial v_g/\partial z < 0$.

This means that $|v_g|$ reaches a maximum near the nocturnal inversion level, which means that $|v|$ would reach a larger maximum than could be accomplished strictly through the release of friction. The stability of the air (lack of mixing) below the inversion helps the jet to persist.

Stated another way, in the derivation of the ageostrophic wind components, we assumed that $d\vec{v_g}/dt = 0$. The thermal wind theory shows that this assumption is false due to sloping terrain in the Great Plains. The thermal wind mechanism does not explain the super-geostrophic wind, however.

But, if we impose the diurnal changes in $\vec{v}_g$ on the inertial oscillation of $\vec{v}_a$, the magnitude of LLJ is altered. In this way, the thermal wind mechanism is best used together with the inertial oscillation theory.