THE INTERFACE EFFECT AND THE FORMATION OF A LOW-LEVEL JET ALONG THE EAST SIDE OF THE ROCKY MOUNTAINS

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ABSTRACT

Based upon the analysis of several different causes for the low-level jet along the east side of the Rocky Mountains, the concept of "Interface Effect" is established. The basic mechanism for the formation of the low-level jet in North America has been found to be the compression and divergence, under the driving of ageostrophic winds, of the air columns between two surfaces—the ground and the bottom of inversion-which slope with different patterns in a cross-section normal to the jet stream. As a result, the air parcel is accelerated along the streamline and the anticyclonic shear of the current increased. Also, the diurnal variation of the jet is determined by the interface effect.

I. INTRODUCTION

It is known that frequently there is a low-level jet (LLJ) in the lower layer of the troposphere along the east side of the Rocky Mountains. Meteorologists have made several studies to discover the features and investigate the weather associated with the LLJ (Means, 1954; Blackadar, 1957; Izumi and Barad, 1963; Bonner, 1966, 1968; Djuric and Damiani, 1980). Blackadar (1957) advanced the inertial oscillation theory to explain the formation of LLJ. Wexler (1961) has drawn an analogy between the LLJ stream and the Gulf Stream. He uses the inertial boundary layer theory developed by Charney (1955) and Morgan (1956) to explain the formation of a large-scale LLJ in the lee of the Rocky Mountains. Recently, meteorologists have done some numerical simulations to approximate the process of the forming of the LLJ (Uccellini and Johnson, 1979; Djuric, 1981); thus, the research in this field is of interest today.

There are other LLJ streams in the world, such as the LLJ in Somalia and the jet in East Asia ahead of the "Meiyu" front. Chinese and Japanese meteorologists have made investigations of the LLJ in East Asia (Wang, 1974; Zhu, 1975, 1979; Huang, 1981; Matsumoto, 1972; Zhu and Zhu, 1982). From this work comes two points of view: one is that part of the momentum of the high-level current is transported to lower levels; the second view is that, because of the ageostrophic departure which is very evident, the air flow begins to move faster and the LLJ is formed. The feedback of the latent heat released by water vapor condensation to the temperature and the pressure fields is included in the second viewpoint.

This paper will provide another point of view based on a demonstration of the physical mechanics of the formation and will explain some aspects of the LLJ by the analysis of what will be called the "Interface Effect."

II. THE GENERAL CONDITIONS AT FORMATION OF THE LLJ

Djuric and Damiani (1980) state that in all cases the patterns of circulation and the temperature-pressure fields are similar during the formation of the LLJ along the east side of the Rocky Mountains. The conditions associated with the LLJ during 10 April 1977 are provided as an example. The 700 hPa chart is shown as Fig. 1. A high-pressure center is located over the lower reaches of the Mississippi River with a ridge extending northward. A warm center of 10°C is over Colorado. Cold core troughs are along both coasts. The same pattern occurs at the 850 hPa level. The main feature is that, in the lower troposphere, a cold high is to the east and a warm low is to the west of the area where the LLJ forms.

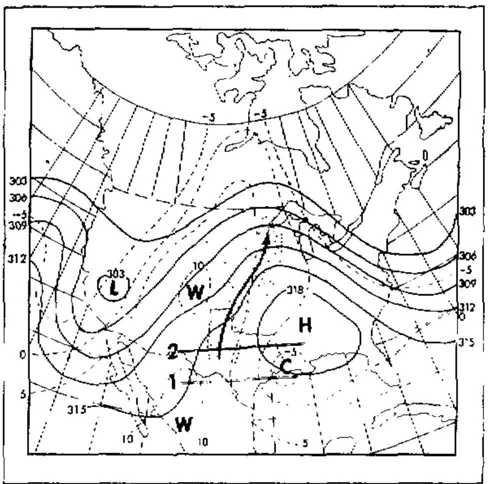

Fig. 1. Map of 700 hPa for 0000 GMT 10 April 1977. Arrow is location of axis of LLJ at 850 hPa level. Hatching represents area where the LLJ forms. Cross-sections 1 and 2 are identified.

It is convenient for discussion to define two east-west cross-sections at the entrance area of the LLJ. The locations of two cross-sections are shown in Fig. 1 and the lines are indicated by "1" and "2". Fig. 2 is the vertical distribution of the wind and temperature fields in these two cross-sections at 0000 and 1200 GMT 10 April 1977. The important feature of Fig. 2 is that an inversion is above the LLJ. The inversion slopes downward to the west and contacts the ground which slopes downward to the east. To the lee of the Rocky Mountains, the LLJ is usually associated with an inversion (Bonner, 1968) and the inversion sloping down toward the west is a general pattern in this area (Djuric, 1980).

The axis of maximum wind speed of the LLJ is about one km above the ground, a little higher in the southern segment than the northern segment. It is a little higher in daytime than at night and the wind speed is stronger at night than in the daytime. All of these features can be found in Fig. 2. These features are the same as the climatological summary by Bonner (1968).

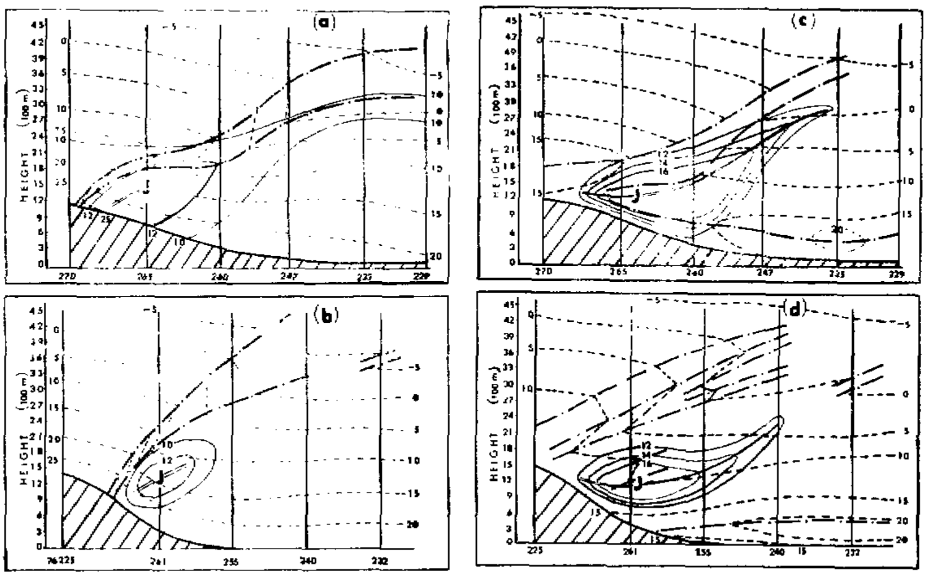

Fig. 2. Cross-sections showing inversions. (a) Section 2 at 0000 GMT 10 April 1977; (b) Section 1 at 0000 GMT 10 April 1977; (c) Section 2 at 1200 GMT 10 April 1977; and (d) Section 1 at 1200 GMT 10 April 1977. Dashed lines, isotherms (°C); thin solid lines, isotherms (m s^{-1}); dot dash lines, the boundaries of inversion.

Fig. 3 shows the vertical distribution of potential temperature and specific humidity along the northern cross-section at 0000 GMT 10 April 1977. The inversion separates the air into two layers, the hot and dry air (cT air mass) above and cool moist air (cP air mass; in other cases it may be mT) below. The fact that there are several isohumes concentrated below the inversion illustrates that the atmospheric elements such as temperature, humidity, etc. are not mixed through the inversion.

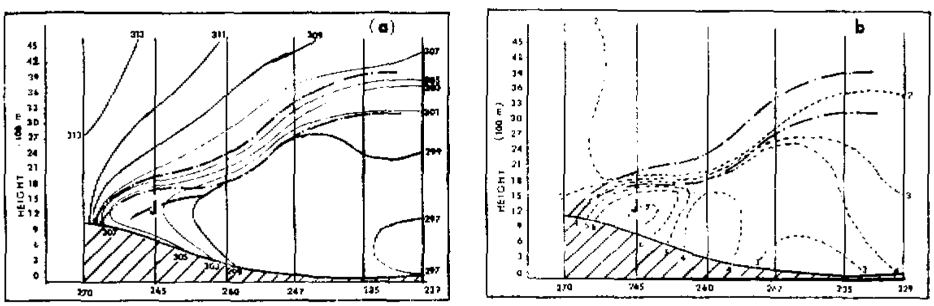

Fig. 3. (a) Distribution of potential temperature in relation to the inversion and to the LLJ and to the LLJ axis along cross-section (1) at 0000 GMT 10 April 1977. (b) Distribution of specific humidity at the same time and place as (a).

The satellite cloud pictures and the weather maps of rainfall amounts for this case indicate no precipitation and no convective activity (clouds) along the east side of the Rocky Mountains from Texas to North Dakota during the time of the LLJ formation. This feature shows that there was no vertical exchange momentum between the Polar Jet and the incipient LLJ. It also indicates that there was no release of latent heat, thus no addition of temperature effects to the pressure and isallobaric fields.

III. ANALYSIS OF THE CAUSES OF THE LLJ

1. *Analysis of Ageostrophic Wind on the West Side of the High-Pressure Area*

Zhu (1975, 1979) and Huang (1981) pointed out that the LLJ, which occurs ahead of the cold or stationary front in East Asia, forms in the southerly current in the western portion of the high-pressure area and is associated with ageostrophic winds.

According to the result of calculations of the ageostrophic wind in this case, there is a strong ageostrophic component crossing the contours toward lower pressure on the west side of the high-pressure area. The distributions of actual wind, geostrophic wind and ageostrophic wind at the 850 hPa level at 0000 GMT 10 April 1977 are shown in Fig. 4. As compared with contours, the east component of observed winds is attributed to ageostrophic movement. The ageostrophic component is very strong, especially in the entrance region of the LLJ where the speed of the ageostrophic wind is almost as large as the speed of the geostrophic and the observed winds. The strongest ageostrophic component wind exists near the axis of maximum wind speed.

Fig. 5 shows the variations of $\bar{u}$, the mean east-west component of observed winds in the entrance region of the current, and $\bar{v}$, the mean north-south component of the winds in the area near the axis of the LLJ at the 850 hPa level during the 72 h period. The variation of $\bar{v}$ demonstrates the formation of the LLJ. The variations of both curves are related. The speed of $\bar{u}$ increases ahead of the speed of $\bar{v}$, (see t_1 and t_2 in Fig. 5). The formation time of the LLJ is defined as the time that the wind speed reaches 12 m s^{-1}. This is later than the time at which the southerly wind begins to strengthen (see t_3 and t_2 in Fig. 5). Therefore the increase in $\bar{u}$ could not be the result of the formation of LLJ.

2. *The First Function of the Interface Effect of the Stable Layer (Inversion)*

It has been shown in Fig. 3 that the bottom of the inversion coincides with potential temperature surfaces. During adiabatic movement without condensation, the air parcels will remain at constant potential temperature. If a parcel is in the inversion layer when the movement begins it will remain in the inversion as long as the potential temperature isotherm does, which for all practical purposes is continuous. In such a case, the movement of the air layer below the inversion is confined within the space between the inversion and the ground. Fig. 6 shows the inversion positions at the time of 0000 GMT 9 April and 0000 GMT 10 April 1977 at both cross-sections. The position of the inversion has a very small change during 24 h, especially at the western part of the cross-sections. When the inversion is stationary, the observed wind is a movement of the air relative to the inversion (and of course to the ground). Therefore, with an easterly component of observed wind, the length of the air column between the ground and the inversion will become shorter and shorter with time. This process may be

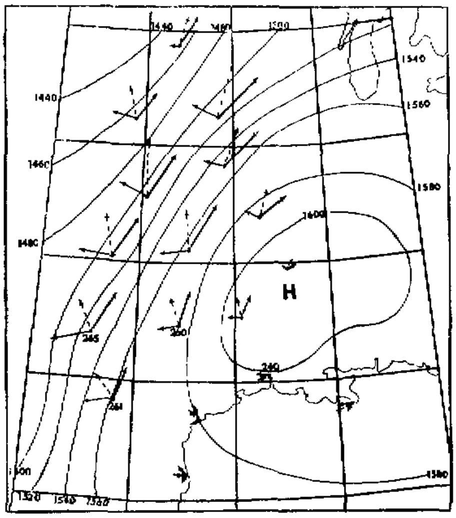

Fig. 4. the 850 hPa contours for 0000 GMT 10 April 1977. The dashed arrows are the actual winds, the double arrows are the geostrophic winds and the single line arrows are the ageostrophic winds.

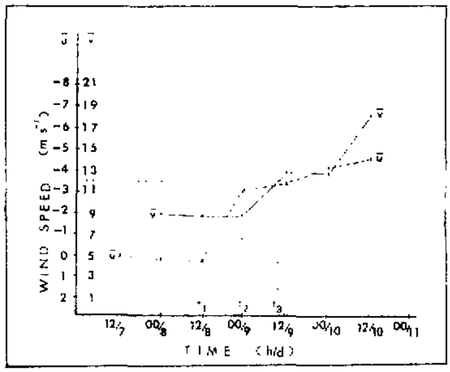

Fig. 5. Change of the $\bar{u}$ and $\bar{v}$ components with time in the entrance region of the LLJ. The east-west components ($\bar{u}$) are the average of stations 72255, 72260, 72261 and 72265; the north-south components ($\bar{v}$) are the average of stations 72451, 72363, 72265 and 72261. t_1, t_2 and t_3 are times as shown. h is hour in GMT and d is day of month, April 1977.

called the "Interface Effect", which is important for the formation of a LLJ. When the inversion is moving the same results will hold if the wind relative to the inversion has a component directed toward the intersection of the inversion with the ground.

Acceleration of an air parcel along the horizontal direction in the entrance region of the current is needed for the formation or for the maintenance of a LLJ. The first function of the interface effect is to force the air parcel to accelerate along the streamline. It may be demonstrated by a computation in this actual case.

For the first or second cross-section, the mean divergence of the wind in the vertical plane under the inversion can be stated as (see Fig. 7)

$$D_{xz}=\frac{\partial u}{\partial x}+\frac{\partial w}{\partial z}=\frac{\Delta S}{S}\approx\frac{H\times\bar{u}}{(H\times L)/2}=2\bar{u}L^{-1}, \tag{1}$$

where S is the area of the triangle, ΔS is the change of the area, $\bar{u}$ is the mean speed of the easterly component of observed wind, H and L are the height and the length of the bottom of the triangle, respectively. x and z are the coordinates. If the inversion is not stationary, $\bar{u}$ is the mean speed of the air relative to the speed of the inversion.

Values of D_{xz} evaluated in the two cross-sections at 0000 GMT 10 April 1977 are shown in Fig. 7. The patterns of D_{xz} are the same in both diagrams. There is divergence present in the eastern portion of the section, i.e., in the high-pressure center, which gradually changes to convergence to the west. Where the interface is closest to the ground, the maximum conver-

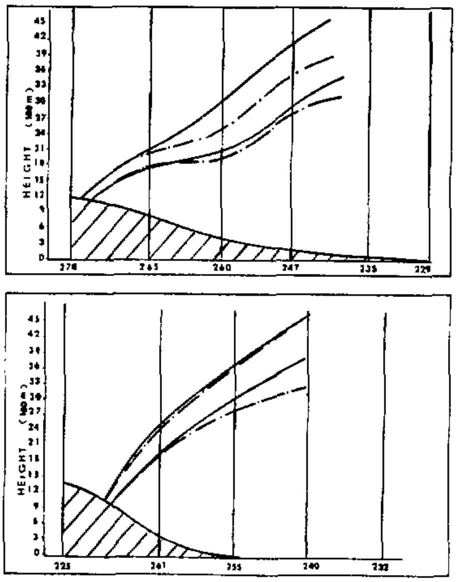

Fig. 6. The change of positions of the inversion during 24 hours along cross-section 2 (upper) and cross-section 1 (lower) between 0000 GMT (solid lines) 9 April 1977 and 0000 GMT (dot dash lines) 10 April 1977.

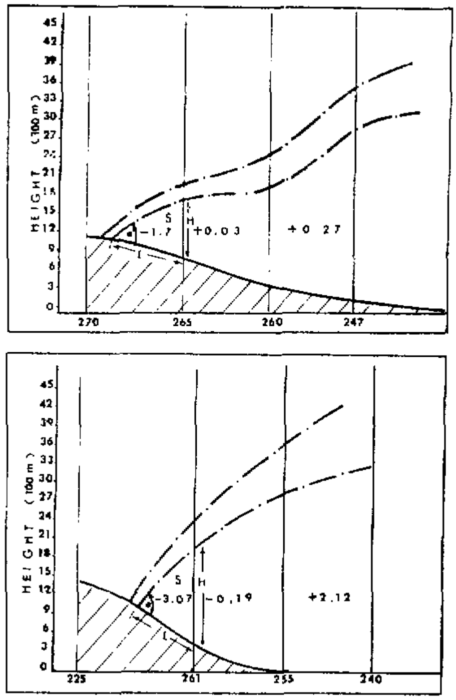

Fig. 7. The distribution of divergence D_{xz} (10^{-5} s^{-1}) in the cross-sections 2 (upper)and 1 (lower) at 0000 GMT 10 April 1977.

gence occurs and it is associated with the axis of the LLJ.

Some days the inversion does not extend to the ground. It disappears at a level some height above the earth's surface. But the pattern of divergence is the same as shown in Fig. 7—the maximum convergence is in the area where the altitude of the inversion is the lowest.

Fig. 8 is the flow pattern at the level of 850 hPa in this case. Assume that the trajectories of air columns under the inversion are well-represented by the streamlines at 850 hPa. Using the incompressible continuity equation, the horizontal acceleration in the north-south direction between the two cross-sections can be estimated by the integration of vertical divergence D_{xz} along the streamline, thus

$$v_2 - v_1 = -\int_l D_{xz} dy \simeq -\frac{1}{2}(D_{xz1} + D_{xz2}) \times l. \qquad (2)$$

Here the subscripts 1 and 2 indicate the values in cross-sections 1 and 2, respectively. l is the length of the streamline between two cross-sections. If an air parcel moves from cross-section 1 to cross-section 2 along the trajectory numbered 3 in Fig. 8, the meridional speed difference is

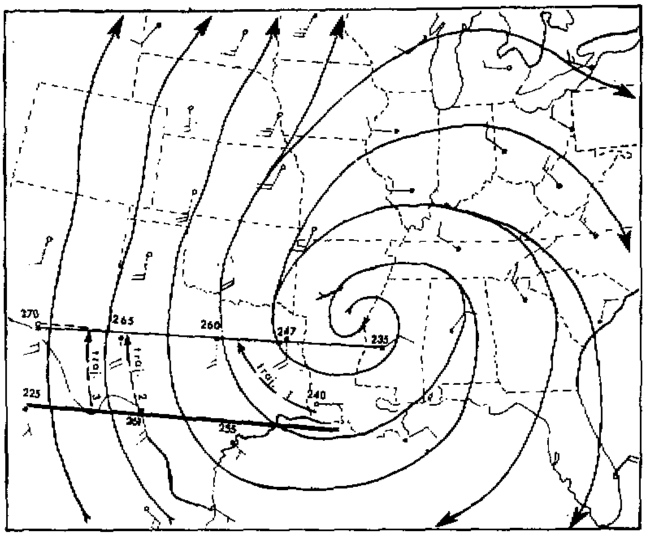

Fig. 8. Flow pattern of 850 hPa level at 0000 GMT 10 April 1977. The locations of three trajectories are shown.

$$v_2 - v_1 = -\frac{1}{2}(-1.7-3.07)\times10^{-5}\times2.3\times10^{5}=5.4\ \mathrm{m\ s^{-1}}. \tag{3}$$

This means that, confined by the stationary inversion and the ground, and because the east component of observed wind becomes stronger before the south component, the air current is accelerated along a S–N direction. The value calculated here is almost the same as the observed value. According to (1) the favorable condition for acceleration is a stronger current from east to west, i.e., a strong ageostrophic wind toward the west.

3. *Thr Second Function of the Interface Effect of the Stable Layer (Inversion)*

The LLJ has an axis of the maximum wind speed in the horizontal field. The second function of the interface effect is to form the axis of maximum wind in a direction parallel to the streamlines. For a dry adiabatic process the mean relative vorticity ζ of a column of air follows the relationship of potential vorticity conservation. In the natural coordinate system, the conservation equation can be written as

$$\frac{d}{ds}\left(-\frac{\partial v}{\partial n}\right)=-\frac{d}{ds}\left(\frac{v}{R_s}\right)-\frac{df}{dt}+\frac{f}{H}\frac{dH}{ds}, \tag{4}$$

where H is the thickness of the air column, R_s, the radius of curvature of streamline, $-\partial v/\partial n$ and v/R_s are the vorticity terms of shear and curvature respectively, and ζ has been neglected in the third term on the right of (4) in comparison with f. Eq. (4) shows that the anticyclonic shear will increase if during motion the length of column decreases and the latitude as well as the curvature vorticity increases.

If the current is in steady state, the streamlines numbered 1 and 2 in Fig. 8 may be treated as good approximations to trajectories. A rough computation has been made of the three terms on the right side in Eq. (4) along these trajectories. The results are shown in Table 1.

Table 1. The Value of Each Term in Eq.(4) (Units: 10^{-4} s^{-1})

Path and Section of Computation	$-\Delta\left(\frac{v}{R_s}\right)$	$-\Delta f$	$+\frac{f}{H}\Delta H$	$\Delta\left(-\frac{\partial v}{\partial n}\right)$
Trajectory 1, from 72240 to 72260	+0.097	−0.044	−0.357	−0.304
Trajectory 2 from 72261 to 72265	+0.006	−0.066	−0.510	−0.570

As shown in Table 1, the value of $-d(v/R_s)/ds$ is positive, but in absolute value, it is much smaller than the sum of the second and third terms, which are negative. Therefore the change of shear vorticity along streamlines in the entrance region is negative. This means that, when air columns move westward or northward along streamlines, the anticyclonic shear vorticity of the current is increased and the air parcel on the left-hand side will move faster than the air parcel on the right-hand side. If we consider the frictional effect caused by the western boundary, the ground, a maximum wind speed center must occur some place west of the high-pressure center but east of the mountains. This maximum wind speed center is the jet axis oriented north-south as seen on the weather map.

It is clear in Table 1 that the change of thickness has a greater impact on the change of shear vorticity than does the change of latitude. For example, if the air column goes from the point near the station 72240 to the point near the station 72260 along the trajectory numbered 1 in Fig. 8 (assume the flow is constant), the shear vorticity is decreased by $-0.304\times10^{-4}s^{-1}$. This effect is about the same as the change caused only by latitude effect if the air column moved north at a distance of about 1700 km and if its thickness remained constant. In fact, the distance between these two stations along the north-south direction is only 250 km. The ratio of the second to the third term in Eq. (4) is about 1/8 in this case. At the entrance region of a jet stream, especially close to the Gulf of Mexico (south of 30°N), there is generally westward movement, hence little latitude effect (see Fig. 8). However, it is in this area that an air parcel has a distinct acceleration and then turns to the right because of the interface and develops increasing anticyclonic shear. Therefore, the interface effect is playing an essential role in the formation of the low-level jet.

IV. INTERFACE EFFECT AND DIURNAL VARIATION OF LLJ

The wind speed of LLJ has a diurnal variation, especially when the sky is clear (Bonner, 1968). This variation may be explained by the diurnal variation of interface effect.

At night a radiation inversion occurs on the ground but during the day the lapse rate is close to dry adiabatic. Fig. 9 is a schematic diagram of diurnal variation of atmospheric stratification near the ground. The day and night temperature soundings are shown schematically in Fig. 9a. The distance L from station to the place where the inversion extends to the ground (or radiation inversion) is smaller at night, as shown in Fig. 9b, because of the radiation inversion. Therefore, the convergence of the eastern component of the observed wind in the vertical section will increase if $\bar{u}$ does not decrease too much. As shown in Fig. 9b. in the daytime

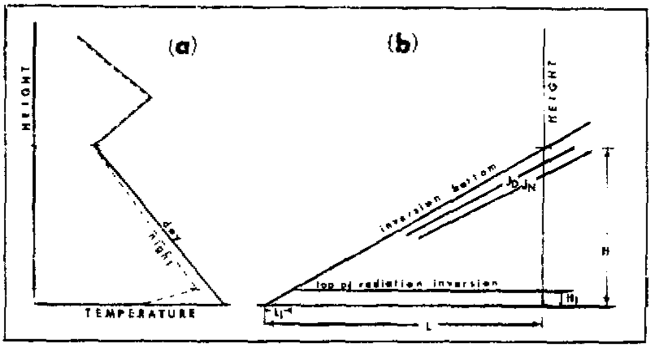

Fig. 9. The diurnal change of the interfaces. (a) represents the lapse rate and (b) shows the change of the parameters H and L and the change of the location of the LLJ and LLJ axis between day and night.

the convergence is

$$D_d=\frac{\Delta S_d}{S_d}=\frac{H_d\times\bar{u}_d}{\frac{1}{2}\times H_d\times L}=2\bar{u}_d L^{-1}. \tag{5}$$

During the night there is

$$D_n=\frac{\Delta S_n}{S_n}=2\bar{u}_n(L-L_1)^{-1}. \tag{6}$$

Therefore, the convergence ratio between daytime and nighttime is:

$$\frac{D_d}{D_n}=\frac{\bar{u}_d}{\bar{u}_n}\,\frac{(L-L_1)}{L}. \tag{7}$$

As has been stated before, the value of convergence D has a leading role in determination of the value of horizontal acceleration in south–north direction. Based on (7), the LLJ must be stronger at nighttime than during the daytime, except in the case when $\bar{u}_d \geqslant \bar{u}_n L(L-L_1)^{-1}$. Therefore, the diurnal variation of the interface effect gives an explanation of the diurnal wind speed oscillation of the LLJ.

V. CONCLUSIONS

(1) There are two basic causes of the formation of the LLJ along the east side of the Rocky Mountains. The first is a strong ageostrophic wind (divergence) that occurs on the west side of a high pressure in the lower atmosphere which gives the observed wind an easterly component. The second is the influence of the two surfaces, the inversion and the ground, on the east side of the Rocky Mountains. Because of the different slope of these two surfaces, the space between the two surfaces is shaped like a wedge which is thicker to the east. As soon as the east component of the observed wind occurs the air is moved westward under the inversion. This movement forces the air current to converge within the wedge space and forces the air column to become shorter. Thus the interface effect is taking place.

(2) Based on the two essential but different principles, the role of conservation of mass and the role of conservation of potential vorticity, the appearance of a maximum wind axis is a result of the interface effect. From the point of view of the mass conservation, the ageos-

trophic flow indicates the convergence in the wedge space and the air parcel is accelerated along a streamline. From the point of view of the conservation of potential vorticity the anticyclonic shear vorticity is increased as the air column moves westward and becomes shorter. Therefore the air parcels on the left–hand side will move faster than those parcels on the right hand side. So the maximum wind axis will occur in some place west of the high pressure, but east of the point at which the inversion intersects the ground.

(3) The increasing of anticyclonic shear of horizontal wind speed, $d(-\partial v/\partial n)/ds<0$, which is related with the LLJ formation, is determined by two factors: the change of thickness dH/ds; and the f effect. Of these factors, the change of thickness is more important for the formation of LLJ, especially in the entrance region of a jet stream.

(4) The diurnal variations of the LLJ are correlated with the diurnal variation of the interface effect.

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