# The diurnal boundary layer wind oscillation above sloping terrain[1]

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ABSTRACT

The role of thermal forcing in the diurnal oscillation of the planetary boundary layer wind above sloping terrain is investigated. Because the gravitational force vector has a component parallel to a sloping boundary, the diurnal temperature oscillation in the boundary layer provides a source of potential energy which drives a diurnal oscillation in the boundary layer wind. Thus, the viscous (Ekman) boundary layer and the thermal boundary layer are coupled.

A set of three second order differential equations which approximately governs the dynamics of the boundary layer is derived. Diurnally periodic solutions are obtained, and it is shown that the thermal forcing mechanism can account for the amplitude of the nocturnal low-level jet observed over the sloping Great Plains region of the United States. However, it appears that time and height variations in the eddy viscosity and eddy heat diffusion coefficients must be included to duplicate in detail the vertical structure and phase of the observed jet.

## Introduction

Meteorological soundings often reveal a nocturnal wind speed maximum in the planetary boundary layer a few hundred meters above the ground. This phenomenon, commonly termed the "low-level jet" has been observed in many locations but is apparently best developed in the Great Plains region of the United States. In this region the jet occurs most often in the spring and summer months. It is generally observed when the geostrophic wind is from the south and is usually called the "southerly low-level jet". The horizontal structure of this jet was delineated by Hoecker (1963) who reported the results of a special observational network which provided east-west cross sections of the boundary layer wind profile from the Mississippi basin to the Rocky Mountains. Hoecker's observations showed a core of maximum nocturnal wind speeds located near the center of this cross section above terrain which slopes upward towards the west.

The time-height variation of the boundary layer wind in this region was described in detail by Buajitti & Blackadar (1957). Their observations, based on routine meteorological soundings, indicated that the wind vector at each level rotates clockwise with diurnal periodicity. After removal of secular changes a hodograph of the wind velocity versus time of day describes an ellipse whose major axis is in the direction of, and whose magnitude is the order of, the mean wind at that level. The maximum wind speed occurs between midnight and dawn, and the oscillation has its greatest amplitude at a height of about 500 meters. Buajitti and Blackadar attributed the nocturnal wind maximum to the nocturnal decrease of eddy viscosity in the boundary layer. The low-level jet is in this view simply an inertial type oscillation driven by the diurnal variation of the frictional force. Buajitti & Blackadar performed numerical experiments for a wide range of assumed time and height variations in the eddy viscosity. The results showed that their mechanism, although it is no doubt an important contributing factor, cannot alone account for the observed amplitude and shape of the diurnal wind oscillation. Moreover, the variation of eddy viscosity alone can hardly explain the tendency for the low-level jet to be concentrated above the sloping plains east of the Rockies.

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Wexler (1961) attempted to account for the geographical location of the southerly low-level jet by assuming that the Rockies act as a lateral barrier which turns westward moving air northward to produce an atmospheric analogue of the Gulf Stream. This hypothesis alone cannot explain the diurnal oscillation of the low-level wind, but it does provide a mechanism for generating the weak daytime southerly low-level jet which was observed by Hoecker. Scale analysis of the equations of motion suggests, however, that the southerly low-level jet is not a close analogue to the western boundary currents in the oceans.

It is the purpose of this paper to analyze a third forcing mechanism originally proposed by Bleeker & Andre (1951). They suggested that the low-level nocturnal convergence of the wind over the Mississippi basin was the result of a large scale "drainage" wind caused by the downslope movement of radiationally cooled air along the slopes of the Rocky mountains to the west and the Appalachian mountains to the east. This gravitational drainage mechanism implies that the thermal and viscous boundary layers are coupled through the diurnally oscillating density field. In the next section a set of equations governing this coupled thermal-viscous boundary layer system is derived, and in the following section diurnally periodic solutions are discussed.

## The boundary layer equations

We define the cartesian $(x, y, z)$ coordinate system in the customary manner with $x$ increasing eastward, $y$ increasing northward and $z$ increasing vertically upward. For simplicity we require that the geostrophic wind and the ground contours be parallel to the $y$ axis. If we also assume that the diurnal temperature oscillation is independent of $y$ then all variables are functions of $x$ and $z$ alone. This approach is justifiable in the present problem because the north–south scale of the low-level jet is much greater than its east–west scale.

We are considering a system in which the lower boundary at $z \equiv h$ (where $h$ is height above sea level) is a function of $x$. It is convenient in this case to formulate the boundary layer equations in a coordinate system $(\chi, y, \zeta)$ in which $\chi$ is tangent to the ground in the $x$–$z$ plane, and $\zeta$ is normal to the ground. The relationship of this coordinate system to the $(x, y, z)$ system is shown in Fig. 1. The equations of motion in the $\chi^*$, $y^*$, $\zeta^*$ system (asterisks denote dimensional quantities) become

$$\frac{du^*}{dt^*} - fv^* \cos\phi - \frac{u^* w^*}{a} = -\frac{1}{\varrho^*}\frac{\partial p^*}{\partial \chi^*} - g \sin\phi + \varkappa_\nu \frac{\partial^2 u^*}{\partial \zeta^{*2}}, \quad (1)$$

$$\frac{dv^*}{dt^*} + fu^* \cos\phi - fw^* \sin\phi = \varkappa_\nu \frac{\partial^2 v^*}{\partial \zeta^{*2}}, \quad (2)$$

$$\frac{dw^*}{dt^*} + \frac{u^{*2}}{a} + fv^* \sin\phi = -\frac{1}{\varrho^*}\frac{\partial p^*}{\partial \zeta^*} - g\cos\phi + \varkappa_\nu \frac{\partial^2 w^*}{\partial \zeta^{*2}} \quad (3)$$

(valid for small values of $\phi$),
where $u^*$, $v^*$, and $w^*$ are the velocity components in the $\chi^*$, $y^*$, and $\zeta^*$ directions respectively, $f$ is the Coriolis parameter, $g$ the gravitational acceleration, $p^*$ is the pressure, $\varrho^*$ is the density, and $\varkappa_\nu$ is the vertical eddy viscosity coefficient. The parameters $a$ and $\phi$ denote the radius of curvature and the slope of the ground respectively:

$$a = \left[1 + \left(\frac{dh}{dx}\right)^2\right]^{3/2} \Big/ \frac{d^2h}{dx^2}, \quad \phi = \tan^{-1}\left(\frac{dh}{dx}\right).$$

Horizontal eddy stress terms have been neglected in (1)–(3) since they are small compared to the vertical eddy stresses in the boundary layer.

The continuity equation is

$$\frac{d\varrho^*}{dt^*} + \varrho^*\left(\frac{\partial u^*}{\partial \chi^*} + \frac{\partial w^*}{\partial \zeta^*} + \frac{w^*}{a}\right) = 0, \quad (4)$$

and the thermodynamic energy equation is

$$\frac{d \ln \theta^*}{dt^*} = \frac{Q}{\varrho^* c_p T^*}, \quad (5)$$

where $Q$ is the heating rate per unit volume, $T^*$ denotes temperature, $c_p$ is the specific heat at constant pressure, and $\theta^*$ is the potential temperature defined as

$$\ln \theta^* = \frac{c_v}{c_p} \ln p^* - \ln \varrho^* + \text{Const}, \tag{6}$$

where $c_v$ is the specific heat at constant volume.

The nonadiabatic heating $Q$ is approximated in this model by the sum of an eddy diffusion term and a radiation term. We write the eddy heat diffusion in the usual manner as $Q_E = \varrho^* c_p \varkappa_H (\partial^2 \theta^* / \partial \zeta^{*2})$. For simplicity we assume that the Prandtl number is unity so that the eddy heat exchange coefficient $\varkappa_H$ is equal to the eddy viscosity coefficient $\varkappa_v$.

The radiative heating is represented by a simple linear parameterization. We let

$$Q_R = \tilde{\alpha}(F - 2\sigma T^{*4}),$$

where $\tilde{\alpha}$ is a "grey body" extinction coefficient per unit length, $F$ is the radiative flux convergence per unit area, and $2\sigma T^{*4}$ is the black body emission per unit area. To simplify $Q_R$ we let $F = 2\sigma T_R^{*4}$, where $T_R^*$ is a radiative equilibrium temperature. We next define $T_R^* \equiv \bar{T}_R^* + T_R^{*\prime}$ and $T^* \equiv \bar{T}^* + T^{*\prime}$ where the bar denotes a horizontal and time average, and the prime denotes a departure from the average. Noting that in the boundary layer $\bar{T}_R^* \approx \bar{T}^* \approx \theta_s^*$, we can expand the relationship for $Q_R$ in a binomial series, and obtain correct to the first order

$$Q_R = 8 \tilde{\alpha} \sigma \theta_s^{*3} (T_R^{*\prime} - T^{*\prime}),$$

which states that the atmosphere is heated or cooled accordingly as the temperature is below or above the radiative equilibrium temperature. A similar linearization of the radiative heating was used by Charney (1959) in his general circulation model.

Equations (1)–(6) are now nondimensionalized so that all dependent variables are of order unity. We define a nondimensional time coordinate $t = t^* f$, and space coordinates $\chi = \chi^*/L$ and $\zeta = \zeta^*/\lambda D$ where $L$ denotes the horizontal scale of the motion, $D$ is the scale height of the atmosphere, and $\lambda \equiv (\varkappa_v/f)^{1/2}/D$ is the ratio of the depth of the Ekman layer to the scale height. We define the following nondimensional parameters: $Ro = U_s/fL$, $\delta = D/L$, $\mu = Dg\varrho_s^*/p_s^*$, $K = D d \ln \theta_s^*/dz^*$, and $\varepsilon = f^2L^2/gDK$. Here $U_s$ is a horizontal scale velocity and $p_s^*$, $\varrho_s^*$, and $\theta_s^*$ are standard values of pressure, density, and potential temperature respectively and are functions of $z$ alone. The dependent variables may now be scaled as follows:

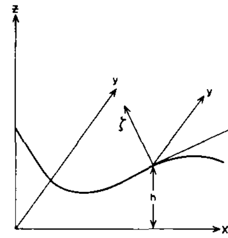


*Fig. 1.* Schematic diagram of the boundary layer coordinate system.

$$u^* = uU_s, \quad v^* = vU_s, \quad w^* = wU_s\lambda\delta,$$
$$p^* = p_s^*(1 + \mu K \varepsilon Ro p),$$
$$\varrho^* = \varrho_s^*(1 + K \varepsilon Ro \varrho),$$
$$\theta^* = \theta_s^*(1 + K \varepsilon Ro \theta).$$

We define $p_s^*$ and $\varrho_s^*$ so that they satisfy the hydrostatic approximation identically

$$\frac{-1}{\varrho_s^*} \frac{dp_s^*}{dz^*} = g, \tag{7}$$

and we define $\theta_s^*$ so that

$$\frac{d \ln \theta_s^*}{dz^*} = \frac{c_v}{c_p} \frac{d \ln p_s^*}{dz^*} - \frac{d \ln \varrho_s^*}{dz^*}. \tag{8}$$

For motions on the scale of the southerly low-level jet, the parameters $Ro$, $\lambda$, and $K$ are all small. In addition the slope of the Great Plains is gentle so that $a \gtrsim o(D)$ and $\cos \phi \approx 1$. Substituting the above-defined nondimensional variables into (1)–(6) and utilizing (7) and (8), we find that after neglecting terms multiplied by $Ro$, $\lambda$, or $K$ we obtain the following nondimensional equations for the boundary layer:

$$\frac{\partial u}{\partial t} - v \cos \phi = -\frac{\partial p}{\partial \chi} - \frac{\varrho \sin \phi}{\delta} + \frac{\partial^2 u}{\partial \zeta^2}, \tag{9}$$

$$\frac{\partial v}{\partial t} + u \cos \phi = \frac{\partial^2 v}{\partial \zeta^2}, \tag{10}$$

$$\frac{\partial p}{\partial \zeta} = 0, \tag{11}$$

$$\frac{\partial u}{\partial \chi} + \frac{\partial w}{\partial \zeta} - u \left( \frac{\sin \phi}{\delta} \mu \frac{c_v}{c_p} \right) = 0, \tag{12}$$

$$\frac{\partial \theta}{\partial t} + u\frac{\sin\phi}{\varepsilon\delta} = \frac{\partial^2\theta}{\partial\zeta^2} + H(\theta_R - \theta), \tag{13}$$

$$\frac{\partial \theta}{\partial \zeta} = \frac{c_v}{c_p}\mu\frac{\partial p}{\partial \zeta}\frac{-\partial \varrho}{\partial \zeta}, \tag{14}$$

where $H \equiv 8\tilde{\alpha}\sigma\theta_s^*/\varrho_s^* c_p f$ and $\theta_R$ is the nondimensional amplitude of the diurnal temperature variation.

The system (9)–(14) is subject to the boundary conditions

$$u,\ \theta,\ \varrho \to 0, \quad \text{and} \quad v \to v_g \text{ as } \zeta \to \infty; \tag{15}$$

and $u = v = w = 0$, and $\theta = \theta_r$ at $\zeta = 0$, where $v_g$ is the geostrophic velocity.

It remains only to specify $\theta_R$, the diurnal temperature variation. If we approximate the heating cycle by a harmonic oscillation of the temperature which decreases exponentially with height we may write

$$\theta_R = \theta_R(0)\,e^{-\zeta}\cos(\Omega t^*),$$

where $\theta_R(0)$ is the amplitude of $\theta_R$ at the ground and $\Omega$ is the angular velocity of the earth. Now $t^* = t/f$ where $f = 2\Omega \sin$ (latitude) so that if we evaluate $f$ at 30° N latitude we may write

$$\theta_R = \theta_R(0)\,e^{-\zeta}\cos t. \tag{16}$$

Thus, $t = 0$ corresponds to the time of maximum temperature at the ground. Before proceeding with the solution of the set (9)–(14) we comment briefly on the physical content of these equations. The $\chi$ momentum equation (9) contains the usual terms for the time dependent Ekman layer plus the term $-\varrho \sin\phi/\delta$. In the Great Plains region $\sin\phi \sim 0(\delta)$ and this term must be retained in the equations. Physically this term represents the gravitational force in the $\chi$ direction which exists when $\phi \neq 0$, and produces downslope (upslope) accelerations when $\varrho > 0$ ($\varrho < 0$). Thus, over sloping terrain the diurnal temperature cycle provides a source of gravitational potential energy which can drive a diurnal oscillation in the boundary layer current. Equation (11) implies that, in the first approximation perturbation pressure is independent of depth in the friction layer and that the geostrophic wind is, therefore, constant in that layer. The term $u \sin\phi/\varepsilon\delta$ in (13) represents the rate of potential temperature change due to advection of the mean potential temperature field along $\chi$. If $\varepsilon > 0$ so that the atmosphere is stably stratified, downslope (upslope) motion will create a positive (negative) potential temperature anomaly which will in turn create a buoyancy force in opposition to the motion. Thus, over sloping terrain the east–west component of the diurnal boundary layer wind oscillation will tend to be suppressed in a stable atmosphere.

## Solution and discussion

If we use (11) and (14) to eliminate $\varrho$ from (9), we obtain a closed set (9), (10) and (13) in the variables $u$, $v$ and $\theta$. We assume solutions of the form

$$\left.\begin{aligned} u &= \bar{u}(\zeta) + RE[U(\zeta)\,e^{it}], \\ v &= \bar{v}(\zeta) + RE[V(\zeta)\,e^{it}], \\ \theta &= \bar{\theta}(\zeta) + RE[\theta'(\zeta)\,e^{it}], \end{aligned}\right\} \tag{17}$$

where the barred quantities are time averaged values, and $U$, $V$ and $\theta'$ are complex amplitude functions for the diurnally periodic fields.

Substituting (17) into (9), (10), and (13), averaging in time and combining to eliminate $\bar{u}$ and $\theta$, we obtain a single sixth order equation in $\bar{v}$:

$$\frac{\partial^6 \bar{v}}{\partial\zeta^6} - H\frac{\partial^4 \bar{v}}{\partial\zeta^4} + (1+\propto)\frac{\partial^2 \bar{v}}{\partial\zeta^2} - H\bar{v} = -Hv_g, \tag{18}$$

where we have set $\cos\phi = 1$ and $\propto \equiv \sin^2\phi/\varepsilon\delta^2$. By a similar procedure we obtain for the periodic component of the motion:

$$\frac{\partial^6 V}{\partial\zeta^6} - (H+3i)\frac{\partial^4 V}{\partial\zeta^4} + (\propto + 2Hi - 2)\frac{\partial^2 U}{\partial\zeta^2} - i\propto U = H\frac{\sin\phi}{\delta}\theta_R(0)\,e^{-\zeta}. \tag{19}$$

From (9), (10), (15) and (17) we obtain the boundary conditions:

$$\left.\begin{aligned} &\bar{v} \to v_g \quad \text{and} \quad V \to 0 \quad \text{as} \quad \zeta \to \infty, \\ &v = \frac{\partial^2 \bar{v}}{\partial\zeta^2} = 0 \quad \text{and} \quad \frac{\partial^4 \bar{v}}{\partial\zeta^4} = v_g \quad \text{at} \quad \zeta = 0, \\ &\text{and} \quad V = \frac{\partial^2 V}{\partial\zeta^2} = 0 \\ &\text{and} \quad \frac{\partial^4 V}{\partial\zeta^4} = -\theta_R(0)\frac{\sin\phi}{\delta} \quad \text{at} \quad \zeta = 0. \end{aligned}\right\} \tag{20}$$

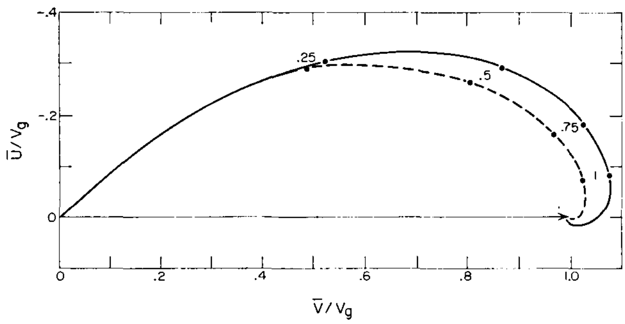


*Fig. 2.* Hodographs of the time averaged boundary layer wind profile for neutral stability (solid line) and an isothermal atmosphere (dashed line). Labelled points on the curves refer to heights in kilometers.

The solutions to (18) and (19) are obtained by standard methods and the results used to determine expressions for the other variables. The resulting rather complicated solutions may be written

$$\bar{u} = \sum_{j=1}^{3} r_j^2 C_j e^{-r_j\zeta}, \tag{21}$$

$$\bar{v} = v_g + \sum_{j=1}^{3} C_j e^{-r_j\zeta}, \tag{22}$$

$$\bar{\theta} = \frac{-\delta}{\sin\phi} \sum_{j=1}^{3} (r_j^4 + 1)\, C_j e^{-r_j\zeta}, \tag{23}$$

$$U = \sum_{j=1}^{3} (s_j^2 - i)\, D_j e^{-s_j\zeta} + (1-i)\, P, \tag{24}$$

$$V = \sum_{j=1}^{3} D_j e^{-s_j\zeta} + P, \tag{25}$$

$$\theta' = -\frac{\delta}{\sin\phi}\left\{\sum_{j=1}^{3} [(s_j^2 - i)^2 + 1]\, D_j e^{-s_j\zeta} - (1-2i)\, P\right\}, \tag{26}$$

where

$$P = H \frac{\sin\phi}{\delta}\, \theta_R(0)\, e^{-\zeta} / [(\propto - 1 - H) + i(2H - 3 - \propto)].$$

The subscripted constants $r_j$ and $s_j$ are defined as follows:

$$r_j = (x_j + H/3)^{1/2},$$

$$s_j = (x_j + H/3 + i)^{1/2},$$

where

$$x_1 = A + B,$$

$$x_2, x_3 = -\tfrac{1}{2}(A+B) \pm \frac{i\sqrt{3}}{2}(A - B),$$

with $A, B = \left(-\frac{b}{2} \pm \sqrt{\frac{b^2}{4} + \frac{a^3}{27}}\right)^{1/3}$,

where $a = (3\propto + 3 - H^2)/3$,

$$b = (-2H^3 + 9H\propto - 18H)/27.$$

Finally, $C_j$ and $D_j$ are defined as

$$C_j = \frac{(1 - r_k^2 r_l^2)\, v_g}{(r_j^2 - r_k^2)(r_j^2 - r_l^2)},$$

$$D_j = \frac{\sin\phi}{\delta}\, \theta_R(0) \times \frac{[(\propto - 1) + i(2H - 3 - \propto) + H(r_k^2 r_l^2 - r_k^2 - r_l^2)]}{[(\propto - H - 1) + i(2H - 3 - \propto)(r_j^2 - r_k^2)(r_j^2 - r_l^2)]}$$

for $1 \leq j, k, l \leq 3$ and $j \neq k \neq l$.

We note that the amplitudes $U$, $V$ and $\theta'$ of the diurnally periodic component depend only on $\propto$, $\delta$, $H$, and $\sin\phi\theta_R(0)/\delta$. They are independent of the geostrophic wind speed. The complexity of these solutions makes it desirable to consider specific examples. To evaluate the influence of stability we will consider two cases: *Case I*, neutral stability; and *Case II*, an isothermal atmosphere.

We recall that $H \equiv 8\tilde{\propto}\sigma\theta_s^*/\varrho_s^* c_p f$, and assigning the values $\tilde{\propto} = 10^{-4}$ cm$^{-1}$, $\theta_s^* = 300$°K, and $\varrho_s^* = 10^{-3}$ g cm$^{-3}$, we obtain $H \approx 1$ at 30° N latitude. From the definition of $\propto$ we have

$$\propto \equiv \frac{\sin^2\phi}{\varepsilon\delta^2} = \frac{\sin^2\phi}{f^2}\left(g \frac{d \ln \theta_s^*}{dz^*}\right).$$

In the Great Plains the terrain slope is typically of the order 1/400 so that $\sin\phi \approx 0.0025$. Therefore, for Case I $\propto = 0$, and for Case II $\propto \approx 0.5$. To evaluate the thermal forcing function we let the dimensional amplitude of, the diurnal temperature wave at the ground be

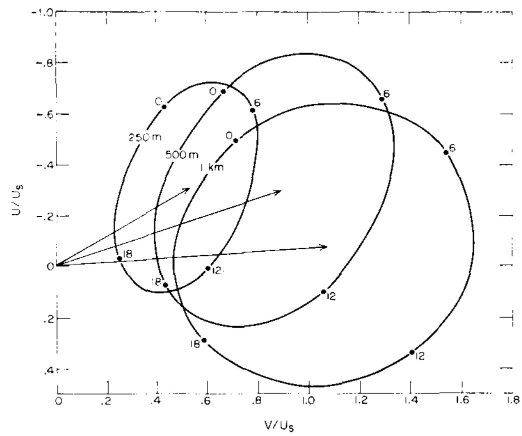


*Fig. 3.* Hodographs of the diurnal wind oscillation at 250 m, 500 m and 1 km for $\alpha = 0$ (neutral stability). Arrows indicate the mean velocity at each level. Times shown on each hodograph indicate hours after time of maximum surface temperature.

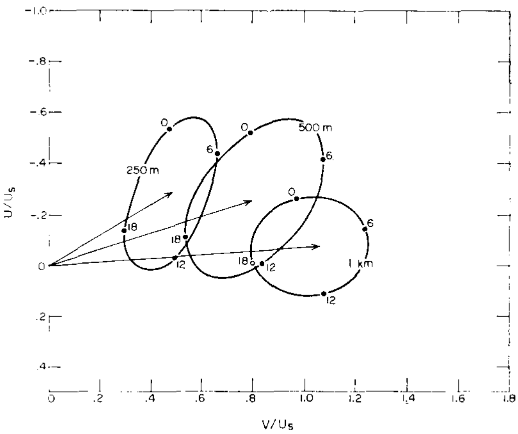


*Fig. 4.* Same as Fig. 3, except that $\alpha = 0.5$ (isothermal atmosphere).

$\theta_R^*(0) = 8.5°$ C and scale the velocity with $U_s = 10$ m sec$^{-1}$. Thus,

$$\frac{\sin\phi\,\theta_R(0)}{\delta} = \frac{\sin\phi\, g}{f U_s}\left(\frac{\theta_R^*(0)}{\theta_s^*}\right) = 1.$$

We first consider the influence of terrain slope on the time mean boundary layer profile. In Fig. 2 wind hodographs for $\alpha = 0$ (solid line) and $\alpha = 0.5$ (dashed line) are displayed. For $\alpha = 0$, which corresponds either to level ground or neutral stability, (21) and (22) reduce to the usual Ekman spiral formula. But in a stable atmosphere over sloping terrain ($\alpha = .5$) cross contour motion is suppressed by the buoyancy force. Hence, the boundary layer spiral is flatter, and the cross isobaric mass transport is smaller than in the neutral case. For the moderate slope considered here the effect of stability on the time mean Ekman layer is rather small. However, since $\alpha$ depends on the square of the slope, the stability effects discussed here may well substantially reduce the cross-isobaric mass transport in mountainous regions.

In Fig. 3 and Fig. 4 hodographs of the diurnal wind variation at 250 m, 500 m and 1 km are plotted for Case I and Case II respectively. Positive stability obviously reduces the amplitude of the oscillation, decreases the height of maximum amplitude, and increases the ellipticity of the hodographs. These solutions indicate that thermal effects can contribute substantially to the amplitude of the diurnal wind oscillation over sloping terrain. However, this constant eddy viscosity model cannot duplicate the observed structure of the oscillation in detail. In particular, the observed phase of the oscillation lags the computed phase by two or three hours and the observed jet has a sharper maximum near 500 m. These discrepancies might be eliminated in a model which included time and height variation of the eddy viscosity and heat diffusion coefficients. Such a refinement would require numerical integration of the coupled viscothermal boundary layer equations.

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## СУТОЧНЫЕ КОЛЕБАНИЯ ВЕТРА В ПОГРАНИЧНОМ СЛОЕ НАД НАКЛОННОЙ МЕСТНОСТЬЮ.

Исследуется роль нагрева в суточном колебании ветра над наклонной местностью. Так как вектор гравитационной силы имеет компоненту, параллельную наклонной границе, суточное колебание температуры в пограничном слое создает источник потенциальной энергии, который вызывает суточное колебание ветра в пограничном слое. Таким образом, вязкий (экмановский) пограничный слой и термический пограничный слой связаны.

Выводится система трех дифференциальных уравнений второго порядка, которые приближенно описывают динамику пограничного слоя. Получены решения с суточным периодом и показано, что механизм нагрева определяет амплитуду ночной приземной струи, наблюдаемой над наклонной областью Великих Равнин США.

Однако представляется, что следует учитывать переменность по времени и по высоте коэффициентов турбулентной вязкости и диффузии тепла для того, чтобы можно было воспроизвести в деталях вертикальную структуру и фазу наблюдаемой струи.