Boundary Layer Wind Maxima and Their Significance for the Growth of Nocturnal Inversions

ALFRED K. BLACKADAR

Department of Meteorology and Oceanography, New York University [1, 2]

ABSTRACT

A sharp maximum is frequently observed at night in the wind speed profile below 3000 ft. The wind speed maximum is usually at the top of the nocturnal inversion, is supergeostrophic, and is often associated with extremely large values of wind shear at low levels.

It is shown that the characteristic velocity profile tends to promote an orderly growth of the nocturnal inversion. The supergeostrophic wind speeds suggest that an inertia oscillation is induced when the constraint imposed by the daytime mixing is released by the initiation of an inversion at about the time of sunset.

INTRODUCTION

ATTENTION has recently been drawn by Lettau [11] and the author [3] to the frequent occurrence of distinct maxima of the wind speed below 5000 ft in the midwestern United States. An example of such a maximum is shown in FIGURE 1. Such profiles often occur in the daytime but they are almost always better developed at night because the diurnal wind variations usually lead to a nocturnal increase of the wind at these levels.

Studies of the horizontal distribution of the wind in these same layers have led to the use of the term *low-level jet* by Means [13] and the recognition of the importance of this jet stream in the formation of squall lines and severe weather phenomena. Lettau and the author have used the same term to describe the maximum in the vertical profile of the wind speed. It should be remarked that although the two kinds of maxima may and frequently do occur simultaneously, they are essentially different phenomena. The occurrence of a jet-like wind profile is by no means confined to the vicinity of the jet-stream as it appears upon the 850 mb surface.

Examples of low-level wind maxima may be found almost anywhere in the United States during any season of the year. FIGURES 2 and 3 show examples of the widespread nature of the occurrence on typical days in winter and in summer. An interesting and apparently rather typical case was observed in detail by Gifford [8] near Washington, D. C., and is shown in FIGURE 4. Despite the extensive area of occurrence, in the United States at least, the phenomenon is best developed over the great plains during the nighttime hours.

An interesting example of a boundary layer wind maximum is to be found in observations of the northeast trade winds near Lake Chad in North Africa at latitude 15°N. Mean wind speed profiles at Fort Lamy for the winter months, as reported by Goualt [9] are shown in FIGURE 5. Similar wind profiles and diurnal variations have been observed in the northeast trades over the Sudan at about the same latitude by Farquharson [7]. Although the winds at Fort Lamy may be affected by Lake Chad (about 60 miles distant), as suggested by Goualt, there is little doubt that boundary layer wind maxima are characteristic at this latitude over dry level surfaces when the weather is clear.

The phenomenon is of great interest because of

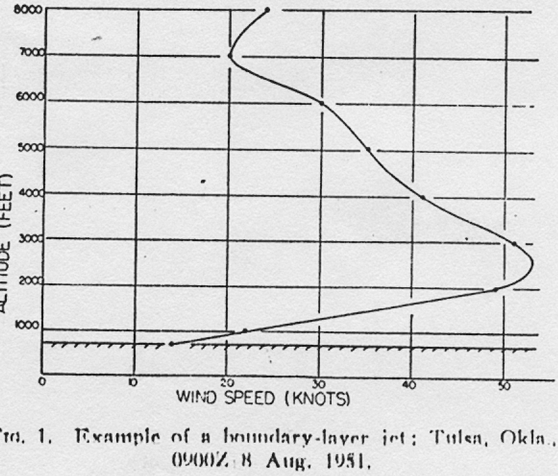

FIG. 1. Example of a boundary-layer jet: Tulsa, Okla., 0900Z 8 Aug. 1951.

[1] Present affiliation: The Pennsylvania State University.

[2] The research reported in this article has been sponsored by the Geophysics Research Directorate of the U. S. Air Force Cambridge Research Center under Contract No. AF 19(604)-1368.

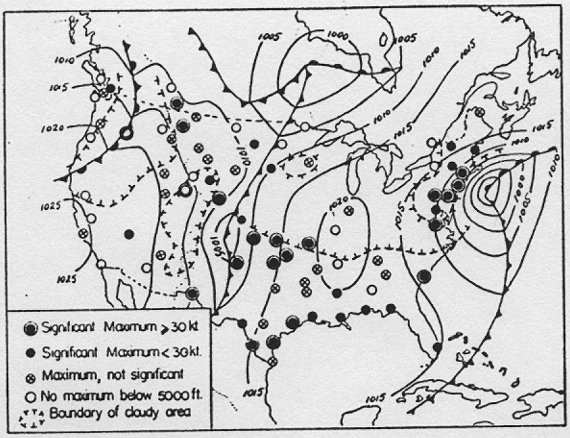

FIG. 2. Distribution of occurrences of boundary layer wind maxima at 0900Z 1 Jan. 1953 in relation to the pressure distribution and clouds at 0630Z.

indirect evidence which may be inferred about the turbulent mass exchange in the planetary boundary layer. It is also of practical interest. Byram [4] has pointed out that the blow-up phenomenon of forest fires, in which a convective column produces a sudden intensification of the burning and a rapid spread of the fire, occurs only when a jet profile of wind speed exists near the surface. The diurnal variation of the wind at low levels, which is closely associated with the occurrence of the wind speed maximum, produces a nocturnal increase of low-level temperature advection which may be a significant factor in causing the high frequency of nocturnal thunderstorms in this area [12].

Low level wind shear, which is always present beneath jet-like wind profiles, is at best a nuisance

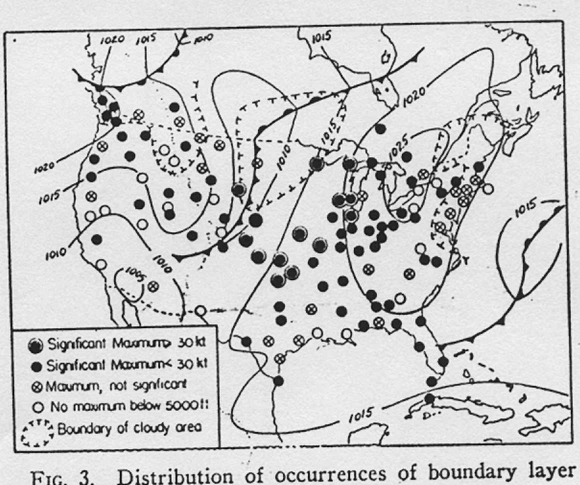

FIG. 3. Distribution of occurrences of boundary layer wind maxima at 0900Z 25 Aug. 1951 in relation to the pressure distribution and clouds at 1230Z.

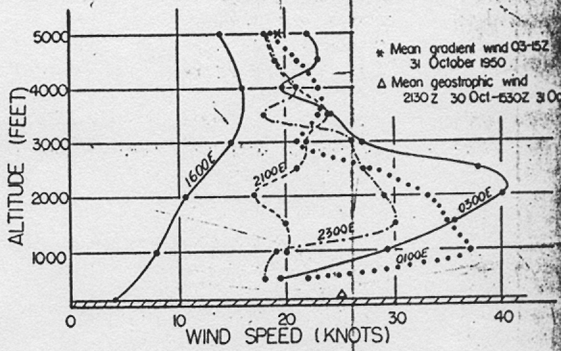

FIG. 4. Evolution of a boundary layer jet profile at Silver Hill, Md., on the night of 30–31 Oct. 1950. Observations were made by Gifford [8].

in tending to cause aircraft to land short of the runway, and occurrences of abnormally large shear are hazardous because of the rapid loss of lift suffered while letting down into layers of decreasing headwind component [14]. The effect is particularly serious in the case of jet aircraft which normally let down at speeds close to the stalling speed, and which require a considerable time to develop sufficient forward thrust after the need for corrective action is recognized.[3] Most of the cases of extreme wind shear in the lowest 100 meters which were recorded during kite observations at Drexel, Nebraska,[4] occurred just beneath low-level jet-profile wind speed maxima during the nighttime hours. In such situations the wind shear frequently averages 1 knot per 10 ft throughout the lowest 300 ft, and sometimes this rate is maintained through even deeper layers. At 11 p.m. on 18 March 1918 a wind speed maximum of 70 knots occurred at an elevation of 780 feet while at the surface the speed was only 5 knots. Particularly puzzling in this case is the fact that the isobars on the surface map indicated a speed of only 20 knots. A study of the conditions which generally accompany jet-like wind profiles may thus be important for understanding and forecasting occurrences of dangerous wind shear.

The regularity of the occurrence of the jet-like profile at certain places and certain times, and the sharpness of the maximum suggest that a real phenomenon is involved which is capable of a physical explanation. It is important, however, to eliminate insofar as possible many less conspicuous but more frequent maxima which are associated with the smaller scale fluctuations of

---

[3] Capt. Thomas B. Gray, Jr., Department of Meteorology, The Florida State University, Tallahassee; personal communication.

[4] *Monthly Weather Review Supplements* 5, 7, 8, 10, 11, 12, 13, 14 and 15 (1916–1918).

the wind and which are therefore probably due to different causes. In the following sections the term *significant maximum* is used to denote those occurrences in which the wind speed at the level of the wind maximum exceeds the wind speed at the level of the next higher minimum by 5 knots or more.

Relation to the Geostrophic Wind

During the hours after sunset, the wind speed at the level of the jet maximum increases rapidly and reaches a maximum some time between midnight and sunrise. Over a fixed station the evolution of the wind profile is probably quite irregular, being marked by fluctuations in the strength of the wind and the height of the wind maximum which have a period which is smaller than two hours. Such fluctuations are apparent in Figure 4 which has been derived from Gifford's observations at Silver Hill, Md. It should be pointed out in connection with these observations that each wind profile is the average of four pilot balloon observations made at fifteen minute intervals.

There is rather conclusive evidence that during the night time and early morning hours the wind speed at the level of the jet maximum is considerably supergeostrophic. During the period covered by Gifford's observations the surface pressure was nearly constant at stations in the area, and analyses of the surface weather maps and 850 mb charts show no change of pressure gradient during the period. The average geostrophic wind during the period was 25 mph at the surface and 17 mph at 850 mb; however, it may be seen from Figure 4 that in the jet maximum the wind speed reached at least 40 mph. This excess of wind speed cannot be accounted for by the anticyclonic curvature of the contour field, for at 850 mb the gradient wind correction amounts to only 2 miles per hour. Nor does it appear likely that much larger values of the geostrophic wind speed occurred in the layer between the surface and 5000 ft, for there was no horizontal variation of temperature whatsoever at 850 mb. It appears therefore that the winds at around 2000 feet were indeed supergeostrophic. Above the jet level the wind at each level exceeds the gradient wind at the same level by roughly the amount of the gradient wind deficiency at 1600 E on the previous afternoon.

Similar conclusions may be drawn from an analysis of the average of 16 occurrences of low-level wind maxima at 0300 Local Time at San Antonio, Texas, during the month of January 1953. During this month, low-level wind maxima occurred on all but two out of the 23 pilot-balloon observations available at that time of the night, but five of these cases were not significant according to the criterion adopted above. The averages of the wind speeds at each level are shown in Figure 6 together with the average wind speed profile at 1500 CST of the previous day. The average geostrophic wind speeds were 23.0, 27.4, and 23.3

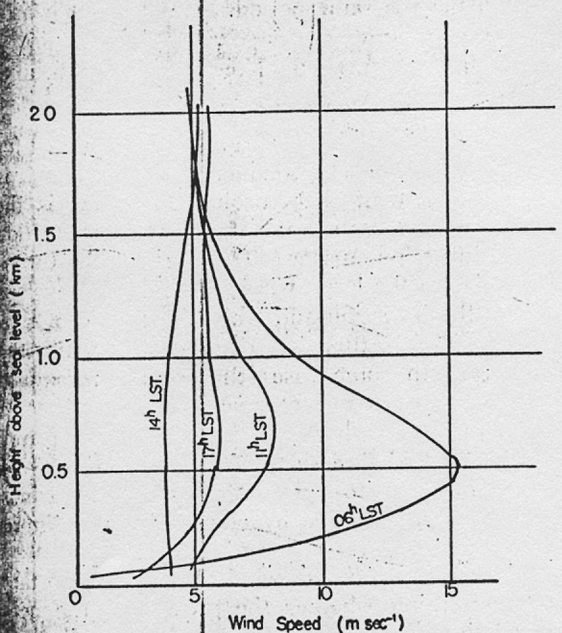

Fig. 5. Mean wind speed profiles in winter at Fort Lamy, after Goualt [9] at various local standard times (LST). The direction is from the northeast.

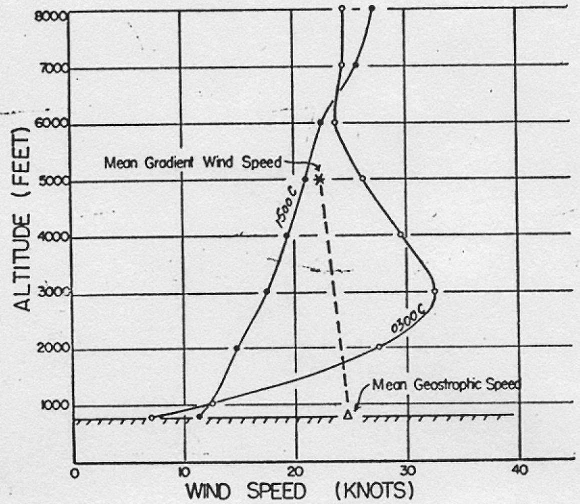

Fig. 6. Average wind speed profile for 16 significant boundary layer jets at San Antonio, Tex., during January 1953.

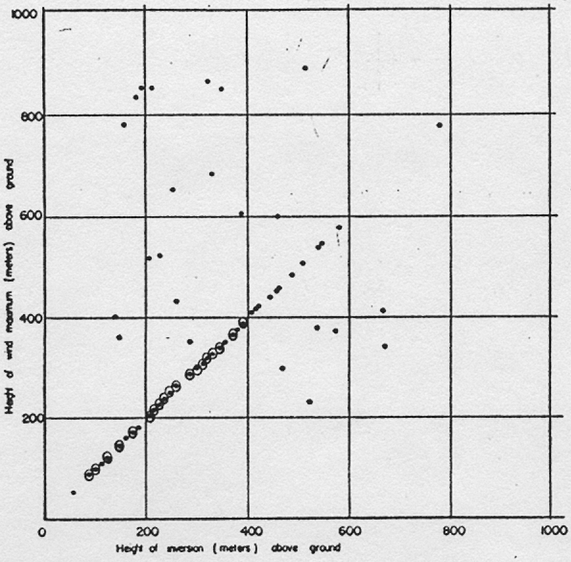

FIG. 7. Comparison of the height of the wind maximum and the height of the top of the nocturnal inversion of 88 selected cases from a series of kite observations at Drexel, Nebraska, between 1916 and 1918.

knots at 0030Z, 0630Z, and 1830Z, respectively, as measured from ozalid copies of surface maps for each of the 16 days prepared at the WBAN Analysis Center. At 850 mb the average geostrophic wind speed was determined from similar charts at 0300Z and 1500Z to be 21.5 and 20.9 knots, respectively. A gradient wind correction was determined for each day from a 12-hr trajectory centered at San Antonio. The geostrophic wind speed averages, including the average gradient wind correction of 0.7 knot at 850 mb, are plotted in FIGURE 6. A comparison with the actual wind speeds shows that the wind speed in the jet maximum is supergeostrophic. Generally, where the wind is most strongly retarded during the afternoon it becomes most strongly supergradient during the night.

RELATION TO THE NOCTURNAL INVERSION

During the period from 8 p.m. to 8 a.m., the height of the wind speed maximum usually coincides with the top of the nocturnal inversion. FIGURE 7 is a scatter diagram of the height of the wind speed maximum *vs.* the height of the temperature maximum obtained from the kite observations at Drexel, Nebraska. In order to restrict the study to nocturnal type inversions, only cases which conformed to all of the following conditions were selected.

1. Local time between 8 p.m. and 8 a.m.
2. Temperature increasing with height up to a single maximum within a distance of 1000 m of the surface (1396 m above sea level).
3. A significant wind speed maximum below 1000 m above the surface.
4. Temperature decreasing with height up to 2000 m above sea level at some time during the previous afternoon.
5. No obvious air mass changes during the night or during the previous afternoon.

Most of the night-time inversions during the summer satisfy these restrictions. During this season the association between the heights of the two phenomena is striking. During the winter, when persistent daytime inversions often occur, the association between the wind speed maxima and temperature maxima generally becomes much more casual. The association which appears in FIGURE 7 is thus characteristic of conditions in which strong diurnal variations are displayed. The reasons for this association will be discussed in the next two sections.

There appears to be a slight increase of the height of the nocturnal inversion with an increase of the wind speed at the height of the maximum. This tendency is no more than would be expected from the fact that the nocturnal inversion is rising during the night while the wind speed is increasing during the same period.

GROWTH STABILITY OF NOCTURNAL INVERSIONS

Above a height of about a meter the rate of nocturnal cooling is too large to be accounted for by the radiational or conduction fluxes, and it is therefore evident that turbulent transfer is the chief control on the rate of upward propagation of the inversion surface. The cause of the turbulence lies in the large wind shear which develops within the inversion and which is capable of supplying sufficient turbulent energy to overcome the stability. In some cases the inversion may be destroyed in the surface layers, or never become established; in other cases the inversion grows slowly upward despite an extremely large wind shear and despite episodes of turbulent winds such as were observed by Durst [5] and Gifford [8].

The character of the wind profile is important in determining whether the growth of the nocturnal inversion is orderly and controlled or whether it is chaotic or perhaps entirely absent. The following discussion indicates that when a jet-like profile exists with the wind maximum at the top of the inversion, the generation of turbulence within the inversion is under automatic control and thus the upward growth tends to be a gradual and orderly one.

Consider a simple nocturnal inversion, based at the ground, of thickness h as shown in FIGURE 8. In the absence of any removal of heat from the layer, a small amount of turbulent mixing increases the potential temperature θ_o at the surface by an amount $d\theta_o$ and increases the height of the top of the inversion by an increment dh in such a manner as to transform the line AB into the line CD. The height m at which the potential temperature is not affected by mixing is found by equating the areas AOC and DOB. The following relations are valid when dh is small compared to h and when $d\theta/dz$ above the inversion is small compared to the value in the inversion:

$$m = h/2 \qquad d\theta_o/dh = \frac{\theta_h - \theta_o}{h}, \tag{1}$$

where θ_h is the potential temperature at the level h.

The question now arises whether the new distributions of temperature and wind resulting from the above processes give rise to a greater or smaller tendency for further turbulent breakdown. This question is resolved by determining whether the Richardson number

$$Ri = \frac{g}{\theta}\frac{\partial\theta/\partial z}{(\partial V/\partial z)^2} = \frac{gh}{\theta_m}\frac{\theta_h - \theta_o}{(V_h - V_o)^2} \tag{2}$$

is decreased or increased as a result of the change. In this equation, V is the wind speed and g is gravity. Thus if

$$dRi/dh \geq 0 \tag{3}$$

the new Richardson number is greater than that originally present, and so it may be expected that the change has decreased the likelihood of further turbulence, at least in the absence of other processes.

The evolution of the inequality, treating θ_o, θ_h, and V_h as variables, leads, after substitution of equations (1), to the relation:

$$\frac{dRi}{dh} = \frac{gh}{\theta_m(V_h - V_o)^2} \times \left[\frac{d\theta_h}{dh} - 2\frac{(\theta_h - \theta_o)}{(V_h - V_o)}\frac{dV_h}{dh}\right] \geq 0. \tag{4}$$

Now g, h, θ_m, $(\theta_h - \theta_o)$, and $(V_h - V_o)$ are positive quantities under the conditions being considered here. Therefore, whether or not the inequality is satisfied depends upon the values of $d\theta_h/dh$ and dV_h/dh.

At and above the level D in FIGURE 8, the turbulent processes being considered here do not change the potential temperature or the wind speed. Therefore,

$$\frac{d\theta_h}{dh} = \frac{\theta_D - \theta_B}{dh}$$

which is equal to the value of $\partial\theta/\partial z$ prevailing above the inversion. In a similar manner, the ratio dV_h/dh is equal to the value of $\partial V/\partial z$ which prevails above the point reached by the rising inversion. The ratio $d\theta_h/dh$ is not likely to become significantly negative. Therefore, in order to satisfy the inequality (4) it is sufficient that above the inversion the wind decrease with height, a condition which, of course, exists when the maximum wind occurs at the top of the inversion.

While the wind shear is probably sufficient in most nocturnal inversions to allow a continual generation of some turbulence, the presence of a wind speed maximum at the top of the inversion assures that this generation is kept under control. Then the generation of turbulence continues only so long as other processes tend to increase the wind shear or to change the temperature distribution. Moreover, these considerations show that a jet-like profile with a wind maximum at the top of the inversion is a stable configuration, whereas, the existence of a wind maximum above the level of the inversion is liable to result in a chaotic breakdown.

EXPLANATION OF THE PHENOMENON

The diurnal wind variation which is characteristic over land surfaces was first explained by Espy [6], and later in more detail by Köppen [10], as the result of variations of the downward

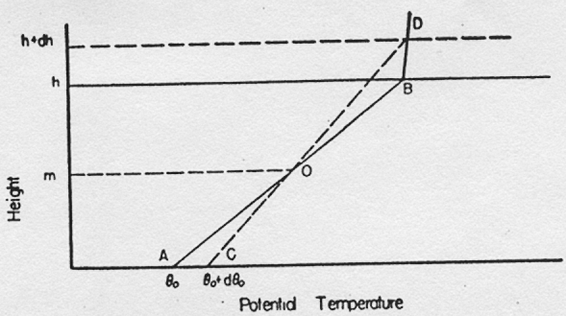

FIG. 8. Potential temperature distribution in an idealized nocturnal inversion (solid line) and the corresponding distribution after a small amount of turbulent mixing (dashed).

turbulent flow of momentum. Wagner [15] showed that the daytime maximum wind speed in the surface layers is caused by the strong increase of eddy viscosity with height in this layer. Inasmuch as the wind was supposed to be in a quasi-equilibrium at all times it was predicted that the wind above the surface layer is retarded during the daytime and approaches the geostrophic value at night. Thus none of these theories explains the strongly supergeostrophic winds which occur at night within the boundary layer.

Wagner [16] made a study of the boundary-layer winds in the midwestern United States, and came to the conclusion that the diurnal variations of the wind could be completely explained by the superposition of three wind systems:

1. A circulation between the dry region in the southwestern United States and its surroundings.
2. A circulation between the plains and the mountains to the west, and
3. A circulation between the sea and the continent.

Wagner's conclusions are open to question because the analyses were made of observations which, being at a constant height above sea level, were therefore at various distances above the ground and at varying local times. Thus Wagner neglected boundary-layer mixing and did not make allowance for these effects in his analysis.

Vertical turbulent mass exchange through the jet level is not favorable for the maintenance of the jet because the tendency for mixing is always to smooth out maxima and minima of the velocity profile. It is postulated, therefore, that after the nocturnal inversion first begins to be established at about the time of sunset, the turbulent mixing very rapidly dies away above the inversion, and ceases to have any important effect on the motion at these levels. Within the deepening nocturnal inversion some turbulence is maintained because of the large wind shear. Heat is transferred downward to the surface where it is lost by radiation. This heat loss which is not compensated in the upper layers results in the continuous upward growth of the inversion during the night. In a like manner momentum is removed from the upper portion of the inversion and is carried downward by the turbulence to the surface where it is dissipated most effectively. Above the level reached by the deepening nocturnal inversion, no appreciable cooling has yet taken place, and it seems reasonable to postulate that likewise there has been no loss of momentum from these levels. It is possible that after a pronounced jet-like profile has become established the negative wind shear above the wind maximum may become sufficient to generate turbulent mixing, especially in view of the relatively small hydrostatic stability. Such mixing would have an important effect upon the subsequent evolution of the wind profile, but is probably not a significant factor during the formation of the wind maximum.

It is assumed for simplicity that the horizontal pressure gradient is constant in time and in each horizontal plane. It is also assumed that the motion is completely horizontal. Then, above the inversion, the equations of motion may be written

$$\frac{\partial}{\partial t}(u - u_g) = f(v - v_g)$$

$$\frac{\partial}{\partial t}(v - v_g) = -f(u - u_g) \tag{5}$$

in which u, v, u_g, v_g are components of the wind and geostrophic wind and f the coriolis parameter. The solution and geometric interpretation of these equations are facilitated by introducing the complex number

$$W = (u - u_g) + i(v - v_g) \tag{6}$$

which, when plotted in the complex plane, gives a vector representing the deviation from the geostrophic wind. The equations (5) then become

$$\frac{\partial W}{\partial t} = -ifW \tag{7}$$

which may be integrated to give the solution

$$W = W_o e^{-ift} \tag{8}$$

where W_o is the deviation at the initial time, which is taken to be at about sunset. The type of motion which this solution represents is shown in FIGURE 9. It is seen that the deviation from the geostrophic wind remains constant in magnitude but rotates to the right, its period for one complete revolution being one-half pendulum day. It may be seen from FIGURE 9 that if W_o is a typical geostrophic deviation at the time of sunset, a supergeostrophic maximum of the wind speed is reached about six pendulum hours later—about 12 hours at San Antonio and 9½ hours at Washington, D. C.

Above the top of the inversion where this solution is valid, the solution at each level may be considered independently of the other levels above or below, except that at the initial time, W_o is a determined function of height. From the initial

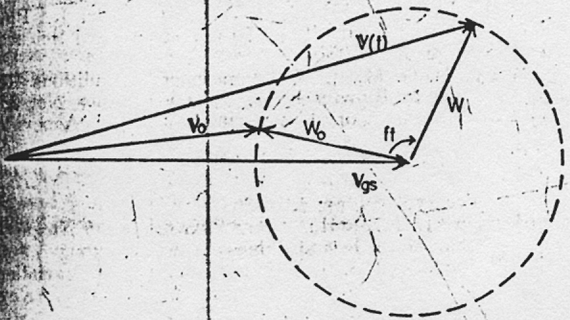

FIG. 9. Relation of the complex number W and the wind vector $\mathbf{V}(t)$ to the initial values W_o, $\mathbf{V}_o$, and the geostrophic wind vector $\mathbf{V}_{gs}$ during a frictionally initiated inertia oscillation.

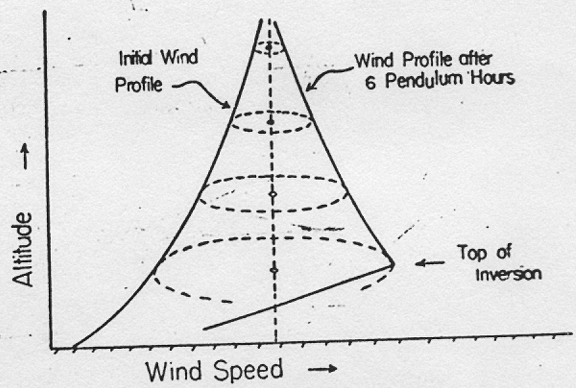

FIG. 10. Schematic illustration explaining the evolution of a boundary layer jet profile.

distribution of wind with height, one may predict the entire wind distribution at a later time, subject to the limitations imposed by the assumptions. The manner in which the solutions at each level may be combined to explain the wind profile above the inversion is indicated in FIGURE 10. In many cases the initial deviation is not exactly opposite to the geostrophic wind; in such cases the time of maximum will be reached after a somewhat different interval of time and will not necessarily occur simultaneously at all levels. It has not been necessary to assume that the geostrophic wind is independent of height. Therefore, FIGURE 10 could easily be modified to apply in cases where the geostrophic wind changes with elevation. The sharpness of the wind maximum tends to be enhanced when the geostrophic wind decreases with height. If the decrease is sufficiently rapid, a jet-like profile may occur even in the daytime. Conversely, the jet profile may not occur at all during the night if the geostrophic wind increases too rapidly upward.

Below the level of the wind maximum, the velocity distribution must be explained by supposing that varying amounts of momentum have been extracted by the downward turbulent flow and subsequently dissipated at the surface. After consideration of the difference between the observed velocity and that which would have prevailed if the motion had been frictionless, it becomes apparent that the greatest losses of momentum have been suffered near the surface, the least at levels just below the wind maximum. Since the loss of momentum at each level may be expected to be proportional to the loss of heat, the profiles of temperature and wind speed may be expected to be similar within the nocturnal inversion.

FIGURE 11 shows the result of an analysis of the deviations from the geostrophic wind which occurred at O'Neill, Nebraska, on 31 Aug.–1 Sept. 1953. The winds, plotted at 2 hourly intervals, represent in most cases averages of several determinations using smoke puff and pibal swarm techniques [1, 2].

The geostrophic winds were measured at 50 mb intervals above and below 850 mb by the following procedures. First, 850 mb charts were plotted from regular radiosonde observations at 1500Z 31 Aug., 0300Z, and 1500Z 1 Sept. These charts were analyzed independently by four analysts and independent geostrophic wind measurements were made over a 300 km line segment centered at O'Neill. The averages of these determinations at

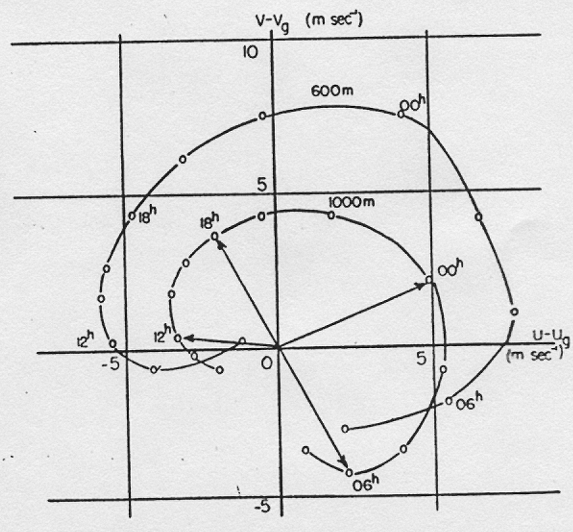

FIG. 11. Geostrophic wind deviation hodographs at 600 m and 1000 m at O'Neill, Nebraska, 31 Aug. to 1 Sept. 1953. Times are CST.

each time were extended to the other levels by the use of thickness charts constructed for each 50 mb layer from mean temperatures measured from the radiosonde observations. The geostrophic wind at 800 m averaged about 13 m sec^{-1} from the southwesterly direction.

During the period following 1800 CST, the deviation vector remained nearly constant in magnitude, and swept out a path which is roughly circular. Between 2000 and 0400 the angular velocity of the deviation vector averaged 13.5° hr^{-1} at 600 m and 17.0° hr^{-1} at 1000 m, compared to the predicted value of 20.2° hr^{-1}.

It is probably possible to improve the accuracy of the predictions by refining the simplified explanation given above. The frictional restraint on the upper layers has been assumed to die away instantly at the initial time, whereas it probably decreases more slowly. As mentioned previously, the large negative shear that appears in the not so stable layer above the inversion at around midnight makes it reasonable to expect that there is some mixing in this region. Studies are now in progress to determine the periodic variations which result from various distributions of eddy viscosity in time and height.

REFERENCES

[1] Barad, M. L., 1954a: "Wind Data Collected by Pibal-Swarm Method at O'Neill, Nebraska." Air Force Cambridge Research Center, unpublished report.

[2] Barad, M. L., 1954b: "Wind Data Obtained by Smoke-Puff Technique at O'Neill, Nebraska." Air Force Cambridge Research Center, unpublished report.

[3] Blackadar, A. K., 1955: "The Low Level Jet Phenomenon." Institute of the Aeronautical Sciences Preprint No. 519.

[4] Byram, G. M., 1954: "Atmospheric Conditions Related to Blowup Fires." Station Paper No. 35; Southeast Forest Experiment Station, Asheville, N. C.

[5] Durst, C. S., 1933: "The Breakdown of Steep Wind Gradients in Inversions." *Quart. J. Roy. Meteor. Soc.*, vol. 59, pp. 131–136.

[6] Espy, J. P., 1841: *"The Philosophy of Storms."* Charles Little and James Brown, Boston.

[7] Farquharson, S. J., 1939: "The Diurnal Variation of Wind over Tropical Africa." *Quart. J. Roy. Meteor. Soc.*, vol. 65, pp. 165–183.

[8] Gifford, F. A., 1952: "The Breakdown of a Low-Level Inversion Studied by Means of Detailed Soundings with a Modified Radiosonde." *Bull. Amer. Meteor. Soc.*, vol. 33, pp. 373–379.

[9] Goualt, J., 1938: "Vents en Altitude a Fort Lamy (Tchad)." *Annales de Physique du Globe de la France d'Outre-Mer*, vol. 5, pp. 70–91.

[10] Köppen, W., 1883: "Über die tägliche Änderung des Windes über dem Lande und dem Meere." *Ann. Hydro. Marit. Meteor.*, vol. 11, pp. 625–643.

[11] Lettau, H., 1954: "Graphs and Illustrations of Diverse Atmospheric States and Processes Observed During the Seventh Test Period of the Great Plains Turbulence Field Program." Occasional Report No. 1, Atmospheric Analysis Laboratory, Air Force Cambridge Research Center.

[12] Means, L. L., 1944: "The Nocturnal Maximum Occurrence of Thunderstorms in the Midwestern States. *Miscellaneous Reports*, No. 16, University of Chicago Press.

[13] Means, L. L., 1952: "On Thunderstorm Forecasting in the Central United States." *Monthly Weather Review*, vol. 80, pp. 165–189.

[14] Neyland, L. J., 1956: "Change Without Notice." *Flying Safety*, April 1956, pp. 16–20.

[15] Wagner, A., 1936: "Zur Theorie des täglichen Ganges der Windverhältnisse." *Gerlands beiträge zur Geophysik*, vol. 47, pp. 172–202.

[16] Wagner, A., 1939: "Über die Tageswinde in der freien Atmosphäre." *Beiträge Physik fr. Atm.*, vol. 25, pp. 145–170.

MEETINGS (Cont'd from page 282)

Sessions are tentatively planned on Descriptive Meteorology, Theoretical Meteorology, Climatology, Long Range Forecasting, and Numerical Weather Prediction.

Meeting headquarters for the Conference will be the Sheraton-Fontenelle Hotel, 18th and Douglas Streets, and a block of rooms has been reserved for those making reservations.

Meeting in College Station, Texas, 13–15 November

The 161st national meeting of the Society will be held on the campus of the A. & M. College of Texas, 13–15 November. Dr. Myron G. H. Ligda, President of the College Station Branch, is Program Chairman. Deadline for receipt of titles, abstracts, presentation requirements is 15 July. Dr. Ligda is also interested in receiving names and addresses of students who attended the meteorology training program of the Signal Corps during World War I at College Station to enlist support for a class reunion. Information should be directed to the Department of Oceanography and Meteorology, A. & M. College of Texas, College Station, Texas.

Papers are especially invited on the following subjects: Agricultural Meteorology, Weather Modification and Cloud Physics; Severe Weather (observation, analysis and forecasting); Trajectories; Interaction between Sea and Atmosphere; IGY observations or studies. Tentatively, special symposia are planned on Education; Professional Standards and Objectives; Encouragement of Research; Public, Commercial and Military Meteorology Requirements.

Programs of the fall meeting programs will be published in the September *Bulletin*.