A Boundary Layer Interpretation of the Low-level Jet

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Abstract

The inertial boundary layer theory developed by Charney and Morgan to explain the intensification of currents near the western boundary of an ocean is applied to the northward flowing atmospheric jet stream observed in the lowest two kilometers over the central United States; here the mountainous spine of Central America serves as a barrier to the westward flowing trade-wind current and deflects it to the north. Because of diurnal frictional changes caused by surface heating and cooling, the jet undergoes a marked diurnal variation with a noctural maximum so strong that the current becomes super-geostrophic. The dynamic consequences of the unbalanced current are examined in the light of theories by Tepper and Veronis.

The thin, narrow concentration of momentum in the lowest 2 kilometers of the atmosphere known as the "low-level jet stream" has attracted surprisingly little attention, apart from work by WAGNER (1939), MEANS (1952, 1954), GIFFORD (1953), BLACKADAR et al. (1957), ESTOQUE (1959). This jet, only a few hundred kilometers wide, may move twice as fast as layers several hundred meters below or above (figs. 1, 11). Although its direction is fairly constant from the south or southwest, its speed has a large diurnal variation with an early morning maximum double or more the afternoon minimum (fig. 2). In the United

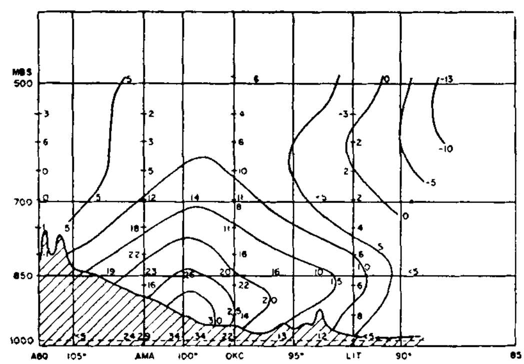

Fig. 1. Mean southerly wind components along 35 degrees latitude, 1500Z July 10th through 0300Z July12 th, 1951, speed in knots (after Means, 1954),

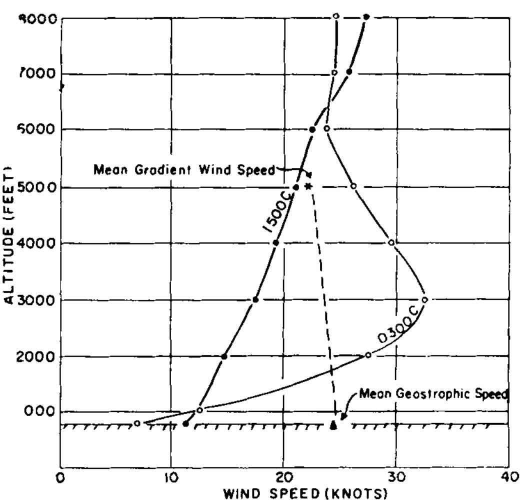

Fig. 2. Average wind speed profile for 16 significant boundary layer jets at San Antonio, Texas, during January 1953 (after Blackadar et al., 1957).

States, it appears most prevalent in the strip between 95° W and 100° W, from southern Texas to Nebraska, a region also noted for its high frequency of severe nocturnal thunder- and wind-storms. Pilots are grateful for its 30 knots or higher "sunrise" tail-winds when flying north and avoid its opposing winds on return flights (Polson, 1958). On the other hand, the strong vertical wind shears accompanying the jet are often dangerous to aircraft descending for nocturnal landings at near-stall speeds; practical procedures for forecasting this phenomenon have been devised by Blackadar and Reiter (1958).

Wagner (1939) attempted to explain this nocturnal acceleration of the low-level winds by drainage of cold air from the high ground to the west, as a result of radiational cooling. Newton (1959) has suggested that the low-level jet is quite similar to the Gulf Stream in its narrow concentration of momentum and location just to the east of a meridional barrier—the slopes leading up to the Rocky Mountains in the case of the atmospheric jet and the continental shelf for the Gulf Stream.

In this paper we shall examine further this analogy by applying to the atmospheric jet the inertial boundary layer theories developed and applied so successfully to the Gulf Stream by Charney (1955) and Morgan (1956) and shall utilize the unbalanced current models of Tepper (1955) and Veronis (1956) to study the dynamic consequences of the observed large diurnal variation of the jet from sub- to super-geostrophic speeds.

Western Boundary Currents of Oceans

The intensification of currents near western boundaries of oceans such as the Gulf Stream and the Kuroshio has been the subject of intensive research in recent years. A good review is given by Stommel (1958), to whom credit is given for first pointing out the critical role played by the latitudinal variation of the Coriolis parameter in the formation of such jet-like currents (1948). The linear viscous theories of Stommel (1948), Munk (1950), Hidaka (1949) explained the principal features of the mean circulation but left unexplained the details of the individual boundary layer. This was examined by Charney (1955) and Morgan (1956) who retained in a narrow layer next to the western wall the important non-linear inertial terms neglected in the earlier models, but omitted the frictional terms which were presumed to be important only in the immediate vicinity of the wall.

A simple qualitative description of their results based on absolute vorticity conservation has often been used to explain variations in shear and curvature of currents moving in north-south directions. If $\frac{f+\zeta}{D}$ is constant for individual columns moving northward, then as f, the Coriolis parameter, increases with increasing latitude, ζ, the relative vorticity and/or D, the column thickness, must decrease. If D is kept constant, the column must acquire more anticyclone vorticity relative to the earth. If this is converted mainly into anticyclonic shear, then there must develop a high-speed current at the western boundary of the flow.

Both Charney and Morgan divide the oceanic flow into two regions—a broad exterior current flowing westward and a narrow boundary layer adjacent to the vertical western wall. In the exterior region the flow is geostrophic but in the boundary layer where only pressure and inertial forces are present, the meridional component, v, of flow is geostrophic but not the zonal component, u, because of presence of the large terms, $u\frac{\partial v}{\partial x}$ and $v\frac{\partial v}{\partial y}$, where x points to the east and y to the north. By transforming the primitive equations into those expressing the conservation of potential vorticity and the Bernouilli equation, Charney and Morgan found the volume transport streamlines and surface contours both for the case of a single homogeneous layer and a double layer, where zero horizontal pressure gradient was assumed in the lower thicker layer. Charney assumed that in the interior region, the volume transport of the westward wind-driven surface layer decreased linearly to the north over a distance of 1,400 km, from the Florida Straits to Cape Hatteras; however, Morgan assumed a volume transport which did not vary northward in 2,000 km. Despite this difference in the meridional distribution of the westward transport of water and the difference in magnitude of transport, $8 \cdot 10^{13}$ cm^3 sec^{-1} assumed by

Charney and $2 \cdot 10^{13}$ cm^3 sec^{-1} by Morgan, their results are quite similar.

Application to the Atmosphere

In applying to the atmosphere the inertial boundary layer theory derived for the ocean, the continental slope off the east coast of the United States is replaced as the western wall by the high escarpment of Central America with an average height of about two kilometers. The wind-driven surface water transport is replaced by the trade-wind transport itself as shown schematically in figure 3. The trade-winds entering the western Caribbean Sea and the Gulf of Mexico impinge on the southeast-northwest oriented Isthmus of Panama; the northern portion is deflected northwestward and northward and takes on the characteristics of an inertial boundary layer because of the northward increase of the Coriolis parameter, and enters the Mississippi Valley. In the computations below, it is assumed that that portion of the trade-winds south of Jamaica crosses the Isthmus and does not enter the Mississippi Valley.

In Table 1 are shown the average monthly sea-level geostrophic wind speeds perpendicular to the 80° W meridian and also the average annual values for 1941—50 for the Colon—Key West section, the Kingston, Jamaica—Key West section, and the Kingston-Jacksonville section, computed from pressure

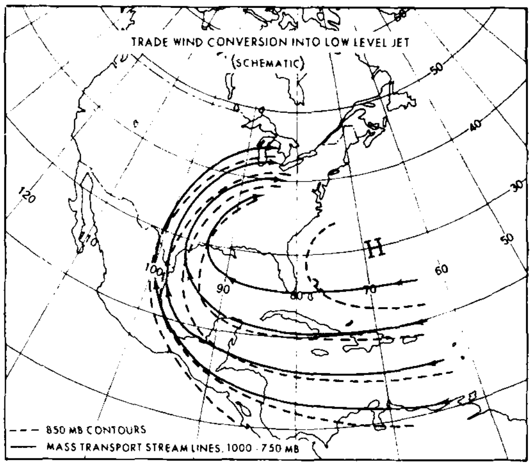

Fig. 3. Schematic sketch of trade-winds entering the Caribbean Sea, Gulf of Mexico and southern United States.

data published in World Weather Records, 1941—50 (Anonymous, 1959).

The average monthly geostrophic wind speeds are about the same across the line from Colon to Key West, and from Kingston to Key West, but smaller from Kingston to Jacksonville. During the Midwest "severe-storm" season, March to May, there is little variation in wind speed, and surprisingly little year-to-year variation in the average annual wind speeds.

Morgan's analysis is applied to a 2-layer atmosphere for two different widths of that

Table 1

Average sea-level geostrophic trade-wind speeds (meter sec^{-1}) across the 80th West Meridian from Colon (90°21′ N) to Key West (24°35′ N); from Kingston (18°15′ N) to Key West; and from Kingston to Jacksonville (30°20′ N).

Average 1941—50				Year			
	CO-KW	*KI-KW*	*KI-JAX*		*CO-KW*	*KI-KW*	*KI-JAX*
Jan.	8.7	9.3	6.3	1941	7.0	7.3	4.6
Feb.	7.9	8.2	5.1	1942	6.3	7.7	5.1
Mar.	6.9	7.3	4.0	1943	6.8	7.3	4.7
Apr.	6.1	6.6	4.6	1944	6.8	7.3	4.7
May	6.0	6.1	3.8	1945	6.4	6.4	4.3
June	6.3	5.2	2.7	1946	6.8	6.6	4.8
July	7.0	5.4	2.9	1947	6.8	6.6	5.1
Aug.	5.6	5.2	3.2	1948	6.4	6.8	4.3
Sept.	3.8	3.6	4.4	1949	6.5	7.3	4.8
Oct.	4.2	4.3	6.1	1950	6.9	8.2	5.5
Nov.	7.1	9.8	6.9	Annual	6.5	6.8	4.7
Dec.	9.2	10.2	6.8				
Annual	6.5	6.8	4.7				

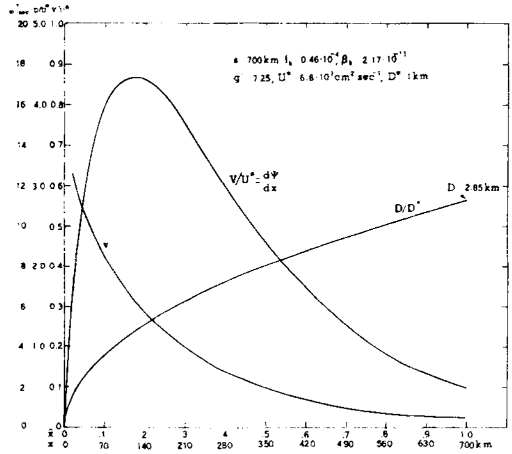

Fig. 4. Computed northward volume transport, (V), relative to trade-wind transport ($U^\star$), velocity (v) and ratio of depth of lower layer (D) to its initial depth ($D^\star$), at the northern boundary of the region where the theory applies, for Case I.

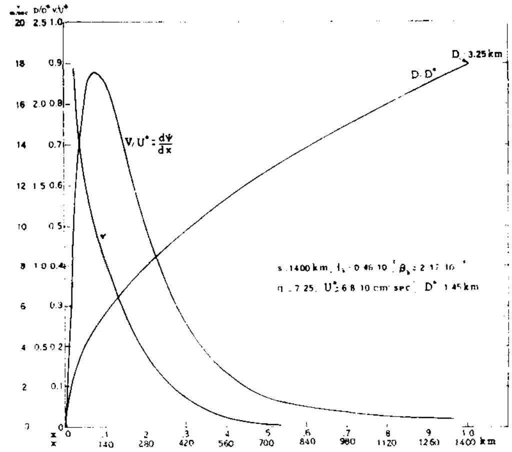

Fig. 5. Computed northward volume transport, (V), relative to trade-wind transport ($U^\star$), velocity (v) and ratio of depth of lower layer to its initial depth ($D^\star$), at the northern boundary of the region where the theory applies, for Case II.

portion of the trade-wind current (assumed to be north of Jamaica) which turns northward into the Mississippi Valley: Kingston, Jamaica to Key West, Florida (700 kilometers) and Kingston to Jacksonville, Florida (1,400 kilometers). In each case the westward transport of trade-wind air across a 1 cm wide strip is taken as $6.8 \cdot 10^7$ cm^2 sec^{-1}, a value corresponding to (I) an average annual sea-level geostrophic velocity perpendicular to a line from Kingston to Key West of 6.8 m sec^{-1} (see Table 1) and an assumed average depth of the trade-wind layer of 1 km, and (II) an average annual geostrophic velocity of 4.7 m sec^{-1} from Kingston to Jacksonville and an assumed depth of the trade-wind layer of 1.45 km. The total transport thus differs by a factor of 2.

When the deeper upper layer is assumed to be motionless, the same equations used for the Gulf Stream apply. The results for the two different transports are shown in figure 4 (Case I) and figure 5 (Case II) for the special condition of the lower layer vanishing at the northwestern corner of the region of applicability of the boundary layer theory.

For this special case, in Morgan's notation:

$$\varepsilon = \frac{g' D^{\star 2}}{\beta_k s^2 U^\star} = 1$$

where $\beta_k = 2.17 \cdot 10^{-13}$ cm^{-1} sec^{-1}, the gradient of the Coriolis parameter at the latitude of Kingston, 18°15′ N,

$U^\star = 6.8 \cdot 10^7$ cm^2 sec^{-1} the average westward volume transport of trade-wind air across a strip 1 cm wide,

s = 700 km or 1,400 km, and

$D^\star$ = 1 km or 1.45 km;

whence g' = 7.25 cm sec^{-2} or 14.50 cm sec^{-2}, corresponding to a temperature inversion of 2.3° C or 4.6° C, respectively, between the two layers.

In figure 4 are plotted three curves against the abscissa $\bar{x} = x/s$, which apply to the northern boundary of the flow ($\bar{y} = y/s = 1$) for the smaller trade-wind transport (Case I):

i) The northward volume transport, V, relative to the westward trade-wind transport $U^\star$. This transport, zero at the wall, reaches a maximum at 118 km east of the wall and approaches zero asymptotically as x increases.

ii) The absolute value of the northward velocity, v. This approaches infinity at $x = 0$, declines rapidly and asymptotically eastward, becoming less than 2 meters sec^{-1} at 350 km. The anticyclonic shears are quite

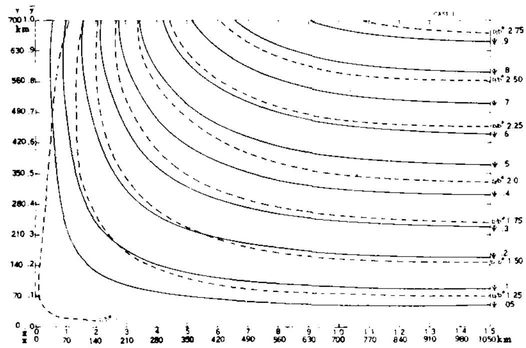

Fig. 6. Computed volume transport streamlines (full) and lines of equal $D/D^{\star}$ (dashed), Case I.

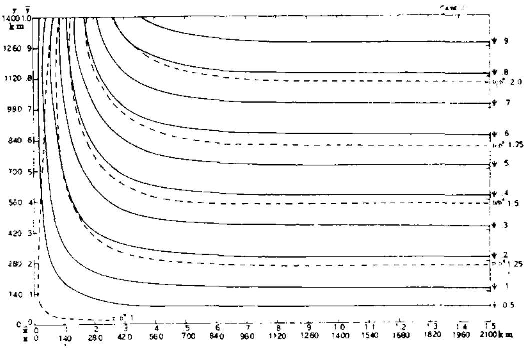

Fig. 7. Computed volume transport streamlines (full) and lines of equal $D/D^{\star}$ (dashed), Case II.

large—greater than or equal to the Coriolis parameter, f, for a distance 100 km east of the wall.

iii) The ratio of the depth of the lower layer, D, to the undisturbed depth of the trade-wind layer $D^{\star}$. This increases rapidly from 0 at the wall to 2.85 at 700 km from the wall, where $D = 2.85$ km.

In figure 5 the corresponding curves are applied to a trade-wind belt twice as wide (Case II), $s = 1{,}400$ km, but having the same average unit volume transport as in the previous case, $U^{\star} = 6.8 \cdot 10^{7}$ cm^2 sec^{-1}, and a larger inversion of 4.6° C.

For the case of the wider trade-wind belt the maximum meridional transport in the boundary layer is 2.6 times larger than in the narrower case; it reaches a maximum at 105 km east of the wall, and is somewhat wider. The northward velocity is also larger than for the narrower trade-wind case, becoming less than 2 m sec^{-1} at 600 km from the wall. The anticyclonic shears are greater than or equal to f for a distance of about 150 km east of the wall. The $D/D^{\star}$ curve increases from zero at the wall to 2.25 at 1,400 km east of the wall where $D = 3.25$ km.

In figures 6 and 7 are shown the volume transport streamlines (full) and lines of equal $D/D^{\star}$ (dashed) for the 700 km wide (Case I) and 1,400 km wide (Case II) trade-wind belts. The streamlines and contour lines are parallel and the motion is geostrophic beyond about 500 km east of the wall (outside of the inertial boundary layer). As the exterior current approaches the wall ($x = 0$) from the right and is deflected northward, it turns not as a uniform current but as a strongly anticyclonic shearing current, increasingly so with distance westward to and northward along the wall. The volume transport streamlines cut across the contour lines towards lower height values and both sets of lines turn anticyclonically so that near the northern limit of domain of validity of the solution, they are directed slightly towards the east.

Computed and Observed Frictional Effects

It thus appears from the close agreement of observation and theory that the northward flow of air in the Great Plains can be interpreted as an inertial boundary layer similar to the Gulf Stream, characterized by strong lateral anticyclonic shears on the western flank. If lateral friction with the western wall is taken into account, as Neumann has done in a linear analysis of the Gulf Stream (1960), then there would be strong shears on the western side of the boundary flow so that the flow would have the appearance of a true jet in the horizontal, as indicated schematically by the dotted lines joining the "v" curves in figures 4 and 5. Vertical frictional stresses originating at the ground would create a vertical wind shear in the surface layer; above the surface layer, where the boundary layer acceleration does not occur, the speed is zero so that the flow also has the appearance of a jet in the vertical. As radiation operates alternately to create and destroy the surface nocturnal inversion, the vertical frictional stresses would undergo a strong diurnal varia-

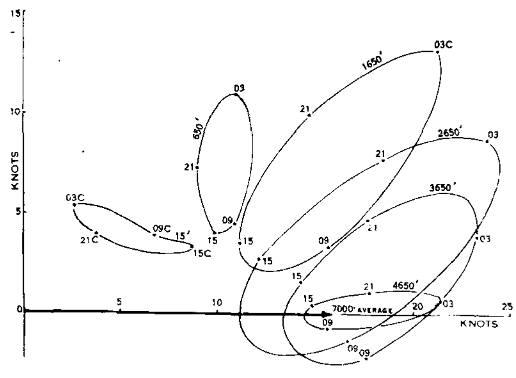

Fig. 8. Composite diagram of diurnal wind variation at Wichita, Kansas and Oklahoma City, Oklahoma (after Blackadar et al., 1957).

tion and so change greatly the profile of the vertical jet.

BLACKADAR et al. (1957) have constructed several mathematical models based on eddy coefficients varying with time of the day and height above the ground in attempts to reproduce the marked observed diurnal variation in the vertical jet. Despite ingenious combinations of time and height variations, the authors admit their inability to reproduce the magnitude of the diurnal change. It is believed that this deficiency in large measure is due to neglect of the mechanism which in the first place actually causes formation of the low-level jet itself. It is not the radiative and frictional effects which occur *locally* that create the highly concentrated flow in the western boundary layer, but the *bulk* properties of the flow caused by large scale inertial effects. This is not to deny the strong influence of large pressure systems moving in and out of the Midwest which might disrupt completely the flow of air from the south. But when southerly flow is present, as in the case when the western end of the Bermuda anticyclone covers the Mississippi Valley, then the inertial boundary layer influences enter in an important manner. Without this westward intensification of the southerly flow up the Mississippi Valley, there would not be a "basic flow" on which the frictional stresses could operate diurnally and so give the strong diurnal low-level jet so characteristic of that region.

During the warmest part of the day when frictional coupling of the low-level jet with the ground is greatest, the jet speed is considerably reduced from its value called for by the frictionless boundary layer theory. But at night, as the growing surface inversion increasingly shields the jet from surface friction, the jet speeds tend to be restored to their boundary layer values. During this interval, the wind speeds in the jet become super-geostrophic. This diurnal variation is illustrated in figure 8, taken from BLACKADAR et al. (1957), who made a composite diagram showing the average of the diurnal wind variations at different levels above the ground at Wichita, Kansas and Oklahoma City. The winds were taken from 6-hourly pilot balloon observations made during 29 days during the summer of 1951, selected to avoid marked changes of surface pressure gradient. The orientation of each wind was measured relative to the wind direction at 7,000 ft above sea-level, at which height the diurnal variation vanished. The diurnal variations are small at and near the ground, increase to a maximum at about 2,000 ft above the ground, at which level the wind speed is sub-geostrophic from about 0900 to about 1800 hours, local time, and super-geostrophic from 1800 to 0900 hours, with the maximum at about 0300 hours. In the sub-geostrophic phase, the wind velocity is such that there is balance between pressure gradient, Coriolis and frictional forces originating at the ground. In the nocturnal super-geostrophic phase, the winds are not in balance since the frictional forces are greatly diminished. Hence, during this phase the vector difference between actual and geostrophic wind describes a portion of an inertial oscillation of period 12 pendulum hours, which at latitude 37.5° is 19.7 hours. This oscillation moves the wind vector arrowheads in an anticyclonic path, theoretically a circle, but actually observed to be an ellipse, probably because of the presence of sizeable frictional forces during the warmer part of the day. The unbalanced flow which exists during the night hours, when the winds are released from the braking influence of ground friction, occurs over the entire jet-stream. The dynamic effects cannot be examined, therefore, merely as a problem involving only the wind changes over a single station, but must be studied from the view-point of the entire jet-stream.

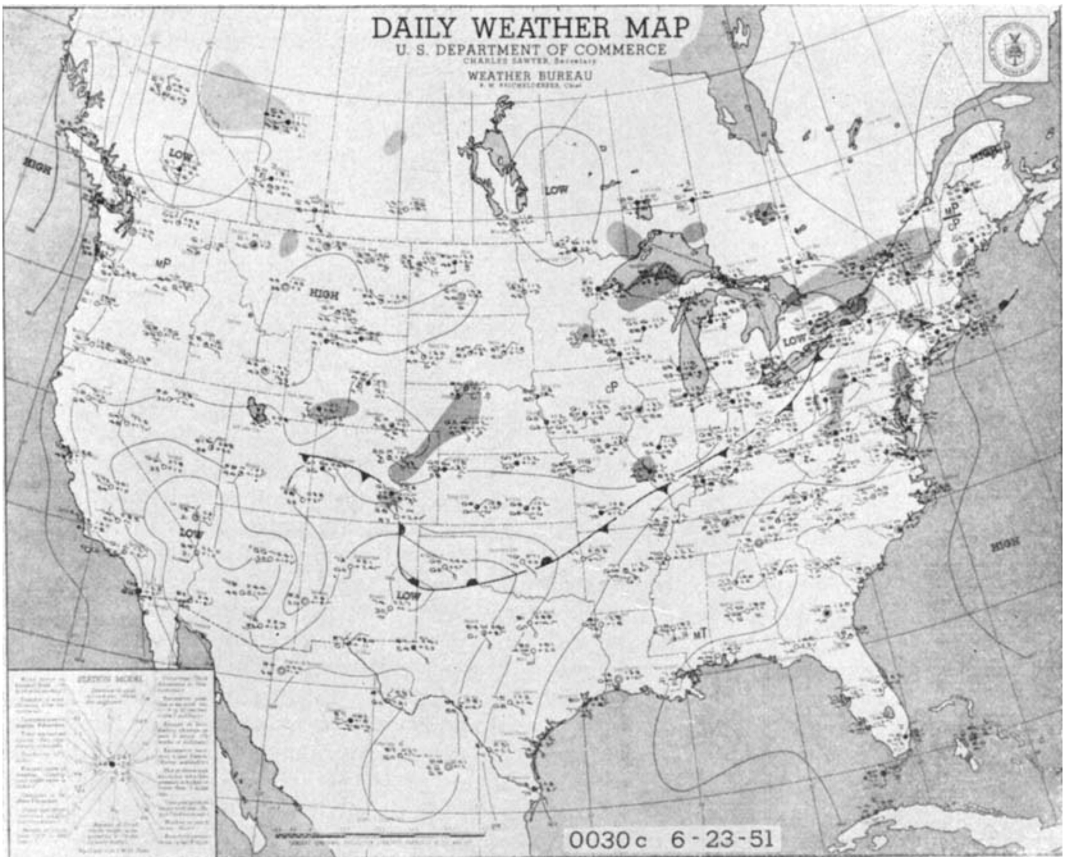

Fig. 9. Sea-level weather map for 0030 C.S.T., June 23, 1951.

Low-level Jet of June 23—24, 1951

A good example of the diurnal variation of the low-level jet occurred on June 23—24, 1951. The synoptic charts for 0030 C.S.T. for June 23 and 24 (figures 9 and 10) show the geostrophic flow of trade air around the western end of the Bermuda anticyclone which extends as far west as west Texas and as far south as latitude 10° N. The isobars indicate a stronger northward flow over central Texas than to the east and the air impinges on a stationary front extending from the Texas Panhandle east-northeast to northern Ohio. About 300 km south of this, five vertical cross-sections are drawn along the line from Albuquerque, New Mexico to Lake Charles, Louisiana (fig. 11). These sections, which reveal a low-level jet centered near Abilene, Texas, are approximately perpendicular to the jet axis and are drawn for 03, 09, 15, 21 and 03 hours C.S.T. The diurnal variation in the thermal stability of the surface layer is shown by the changes in the potential temperature isotherms (full or heavy dashed lines), while the strong variations in jet speed are shown by the isotachs (thin dashed lines). At 03 h when the stability is largest, the main jet axis is over Abilene, Texas, maximum speed 55 knots; at 09 h when the surface layer becomes unstable, this decreases to 35 knots; at 15 h, at maximum instability, the jet maximum decreases to 25 knots; at 21 h when the surface layer becomes more stable the jet maximum increases to 35 knots; and at 03 h the next morning the jet maximum increases to the 55 knot value observed 24 hours earlier. Presumably a major portion of the jet-stream becomes super-geostrophic during the night hours when frictional contact with the ground is curtailed.

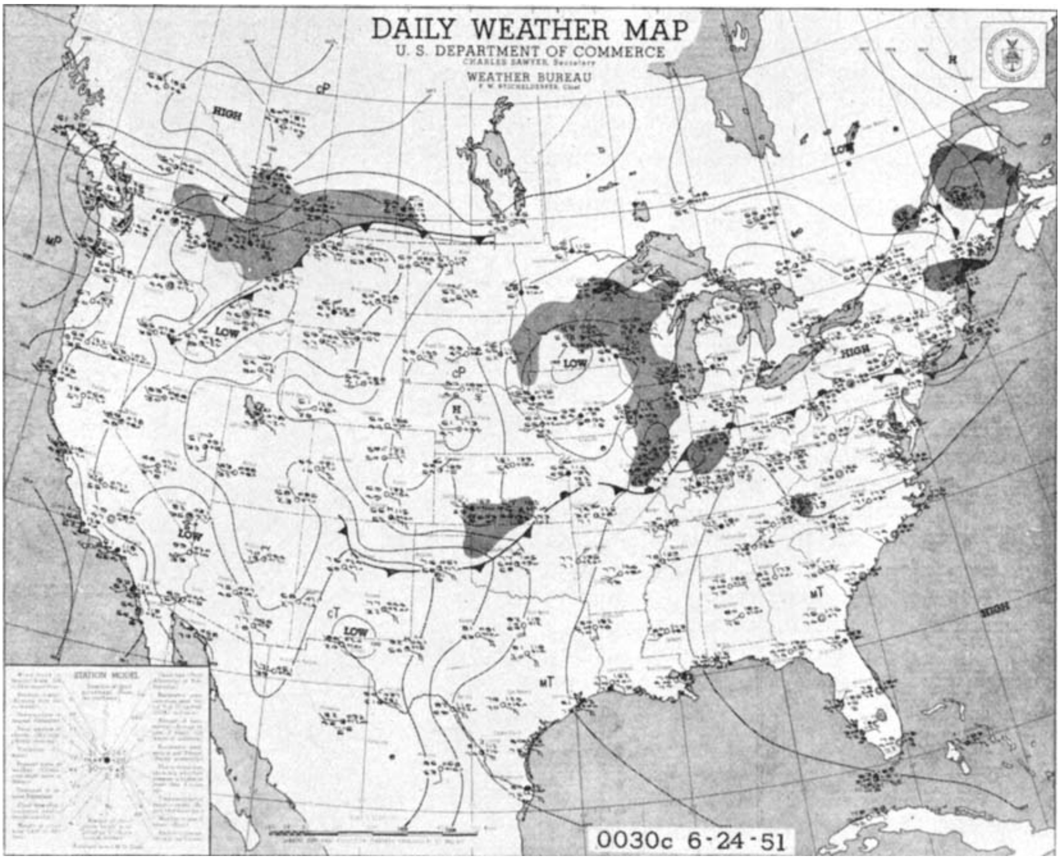

Fig. 10. Sea-level weather map for 0030 C.S.T., June 24, 1951.

There is fairly good agreement between some of the observed features of the low-level jet, illustrated in figure 11, and the theoretical features shown in figures 4 and 5.

First, we can define the top of the lower layer as the 305° Θ-isotherm, which is seen to intersect the surface just west of Big Spring, Texas in the bottom cross-section of figure 11 and rises eastward to a height of 1.7 km over Shreveport, Louisiana, 650 km to the east. This is smaller than 2.75 km at the same distance in Case I (figure 4) and close to the 1.5 km in Case II (figure 5).

Second, the observed jet-stream axis (center of the isotach maximum in the lower section of figure 11) is about 190 km east of the 305° Θ-isotherm intersection with the surface. If we define the computed jet-axis as the volume transport maximum, this is located 110 to 140 km east of the wall in figure 4 (Case I) and about 105 km east of the wall in figure 5 (Case II).

Third, the "width" of the observed jet, defined as the distance from the isotach maximum to the 5 knot (or 2.5 m sec^{-1}) isotach, is about 480 km in figure 11, bottom section. If we define the computed jet-width as the distance from the wall ($v = \infty$) to $v =$ 2.5 m sec^{-1}, the width is 300 km for Case I (figure 4) and 350 km for Case II (figure 5).

Dynamic Consequences of Unbalanced Flow in the Jet

We are thus lead to consider the dynamic consequences of an unbalanced jet-stream directed to the north. In 1955, at the writer's suggestion, such a problem was solved numerically for both a linear and non-linear model by TEPPER (1955). Tepper assumed a

Gaussian-type profile jet-stream of width about 500 km and maximum speed of about 20 meters $\sec^{-1}$ in a 2.5 km thick surface layer separated by a 5° C temperature discontinuity from an upper layer where horizontal pressure gradients were assumed to vanish. With a value of $g' = g\ \frac{\varrho' - \varrho}{\varrho} = 17$, Tepper and Newstein also obtained numerical solutions for Gaussian-type profiles skewed to the east and west (1956).

A marked disturbance of the interface occurred as the Coriolis force caused the unbalanced current to move eastward in the initial phase of an inertial oscillation. The interface rose to the east of the jet axis and lowered to the west, with a maximum displacement upward at the end of 6 hours of about one kilometer at a distance of 500 km from the original jet axis. Results of both the linear and non-linear cases were in close agreement but with important exceptions: for the non-linear case the slope of the leading edge of the interface wave at its maximum amplitude was 1 : 275 compared to 1 : 450 for the linear case. Even more important, in the non-linear case, the slope steepened with time and became essentially vertical after 8 hours giving rise to a "pressure-jump", the atmospheric analog to the well-known hydraulic jump or shock wave.

In comprehensive meso-scale synoptic analyses, TEPPER et al. (1954) showed that in northern Texas and southwestern Oklahoma, pressure jumps of lengths a few hundred kilometers and average duration of 3 to 4 hours, moved perpendicularly away from the basic current and to the right, at average speeds of about 20 meters $\sec^{-1}$, a value close to $\sqrt{g'D}$, the theoretical propagation speed of gravity waves of wave-length large compared to the depth D. They were also found to be three times more frequent during the night hours (from 6 pm to 6 am local time) than during the daytime. This latter point is particularly important since, as noted earlier, the low-level jet stream is most strongly developed at night and in fact becomes super-geostrophic.

During the 8 month period from January to August 1951, 247 pressure jump lines swept through the dense network of 134 observing stations in southeastern Kansas and northern

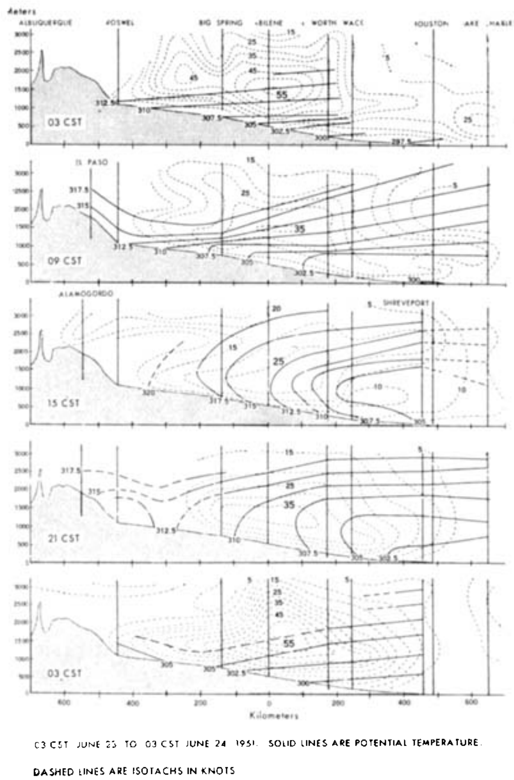

Fig. 11. Five west-east vertical cross-sections at 6-hourly intervals through the low-level jet stream showing the strong diurnal variation of thermal stability in the surface layer and in the speed of the jet, June 23—24, 1951.

Oklahoma and 725 or 86 % of all severe local storms reported were located in the areas swept out by some pressure jump line. It would appear that strong vertical stretching of the lower layer of air accompanied passage of the interface wave and with the air sufficiently humid, as was usually the case, there resulted condensation and liberation of latent heat of condensation which furnished much of the energy of the storm.

The presence of gravity waves on the interface and the steepening tendency of its leading edge are difficult to observe directly since a much denser network of aerological stations is required; but their existence is surmised from surface reports. However, the chain of evidence points to a definite connection between these nocturnal events: the formation of the surface inversion, the quenching of surface friction, the re-establishment of the strong lew-lovel jet and the high frequency of line phenomena (pressure jumps and severe storms) propagating to the right (east or southeast) of the jet.

VERONIS (1956) pointed out two objections to Tepper's analysis of the unbalanced southerly jet: the assumption of an initially unbalanced current as large as 20 meters per second and the neglect of possible non-linear interactions between the baroclinic (or internal) oscillation mode which Tepper examined and the barotropic (or free surface) mode which he neglected. Veronis attempted to remove the first of these weaknesses by allowing the momentum to be given to the current over time intervals of 6 and 12 pendulum hours, respectively, but only analyzed both modes in a linear model so that interactions between internal and external modes were neglected.

Westerly momentum of amount $6\pi \cdot 10^4$ gm cm^{-1} sec^{-1} per unit volume was fed into a surface layer of ocean 200 km wide and the resulting axial and transverse motions, and perturbations of the free surface and the interface computed for $g' = 2$, a 500 meter thick top layer and a motionless bottom layer 3,500 meters thick. We may apply the numerical results of this model to the atmosphere where the 500 meter bottom layer is supplied with the momentum and the upper layer is motionless. The value of $g' = 2$ corresponds to a temperature inversion of about 0.6° C. Since the unit momentum given to the system occurs as a factor in each of the solutions, we can increase the oceanographic results to correspond with meteorological conditions. To do this, we decrease the density by a factor of 10^3 and thus increase the resulting wind speeds and interface and free surface displacements by the same factor. Also, we assume the model is applicable to a southerly current where variation of the Coriolis force with latitude is neglected; according to Veronis, this appears to be acceptable for the short time scales considered here.

Interpreting Veronis' results for a 500 meter thick lower atmospheric layer to which a momentum of amount $2.7 \cdot 10^7$ gm cm^{-1} sec^{-1} per unit volume is given over a period of 6 pendulum hours (nearly 10 hours at latitude 37.5°), we find that there are still evident rather violent lateral oscillations of 10 to 20 m sec^{-1} at the current center, with periods of about 12 pendulum hours. The surface pressure changes at the right edge of the current are of the order of 0.1 mb, or much smaller than observed pressure jumps. These changes are not as large as those found by Tepper who added momentum greater than the amount assumed here and did so instantaneously instead of over an interval of 6 pendulum hours. Also Tepper did not allow free surface oscillations to occur which would have compensated to some extent the pressure changes caused by variations of the interface.

Thus, the question as to whether some pressure jump lines derive their energy from the accelerating low-level jet as ground friction diminishes nocturnally cannot be answered now.

Acknowledgments

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