CASE STUDY OF THUNDERSTORM ACTIVITY IN RELATION TO THE LOW-LEVEL JET

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ABSTRACT

The relationship between the low-level jet and thunderstorm activity in the south-central United States is examined through mesoanalysis of surface data from Weather Bureau and NSSP stations. Separate squall systems moved through Kansas and Oklahoma during the night; the systems in Kansas persisted while those in Oklahoma died out. The most likely explanation for this is the synoptic-scale vertical velocities in the vicinity of the low-level jet.

1. INTRODUCTION

Heavy rainstorms or thunderstorms are microscale or mesoscale phenomena which cannot be adequately described by standard methods of analysis. The techniques of mesoanalysis applied to Weather Bureau and to NSSP α and β network reports make it possible to place the mesoscale convective systems accurately in space and time (cf. Fujita [4], [5]). Prediction of thunderstorm activity, however, is necessarily based upon synoptic-scale analysis of upper-air data, upon conditions of moisture, stability, and wind as determined by widely spaced radiosonde and pibal observations made at 6- or 12-hr. intervals. This study attempts to locate the thunderstorms through mesoanalysis at the surface, and then to correlate the thunderstorm activity with the position and the properties, wind, temperature, and moisture, of the low-level jet. A single 24-hr. period was chosen during which a low pressure system moved across the NSSP (National Severe Storms Project) network of stations in Oklahoma and Kansas (fig. 1), preceded by heavy thunderstorm activity in Kansas, Oklahoma, and Missouri. During this period there was a typical, strong southerly jet near the ground. With the existing network of upper-air stations, it is impossible to analyze the low-level winds on a scale comparable to the surface analyses; however, it is possible to relate the broadscale features of the jet to the thunderstorm activity.

2. DATA SOURCES AND ANALYSIS OF DATA

The following types of data were used:

1. Pressure and hygrothermograph traces from Weather Bureau and from NSSP α and β network stations.
2. Surface weather observations (WBAN 10 Forms) from Weather Bureau stations.
3. Hourly precipitation data published by the U.S. Weather Bureau.
4. Hourly radar summaries (WB Form 610–3) from radar stations in the Midwest and Southwest.
5. Winds aloft computation sheets (WBAN 20) from Weather Bureau, Air Force, and Navy stations.
6. Adiabatic charts (WBAN 31A, B, C).

All winds aloft used in this study were taken from the original minute wind calculations on the WBAN 20 Forms. Sea level pressures have been corrected to better

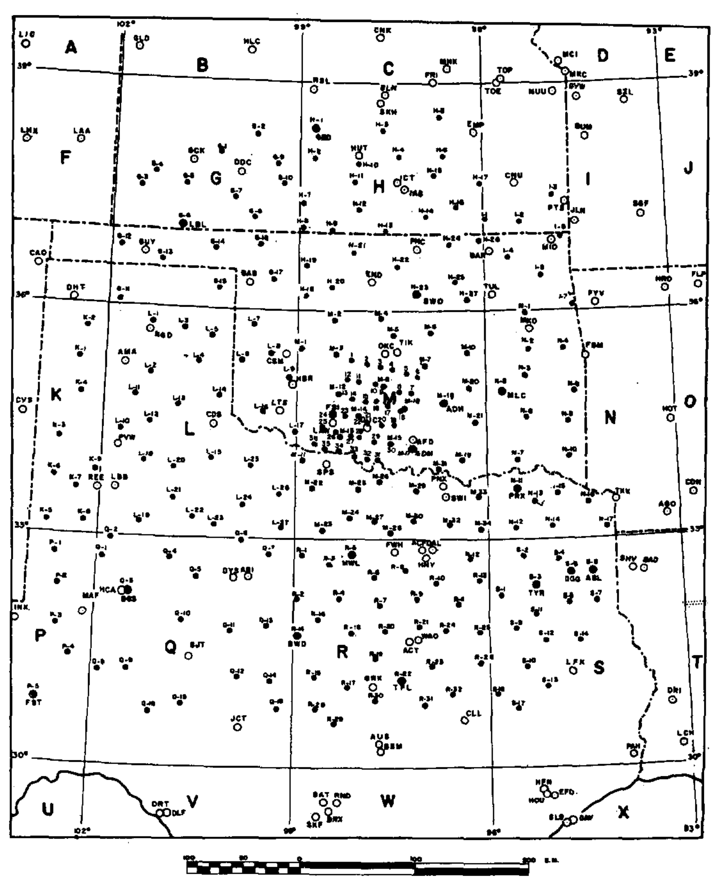

FIGURE 1.—NSSP network stations.

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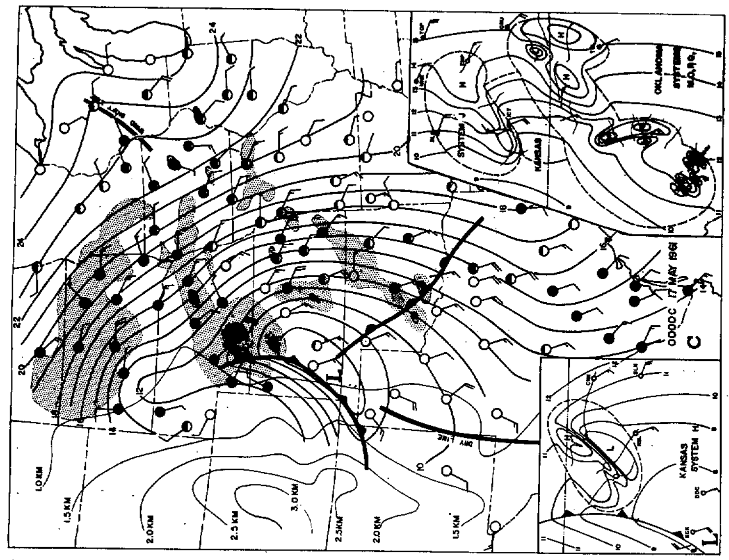

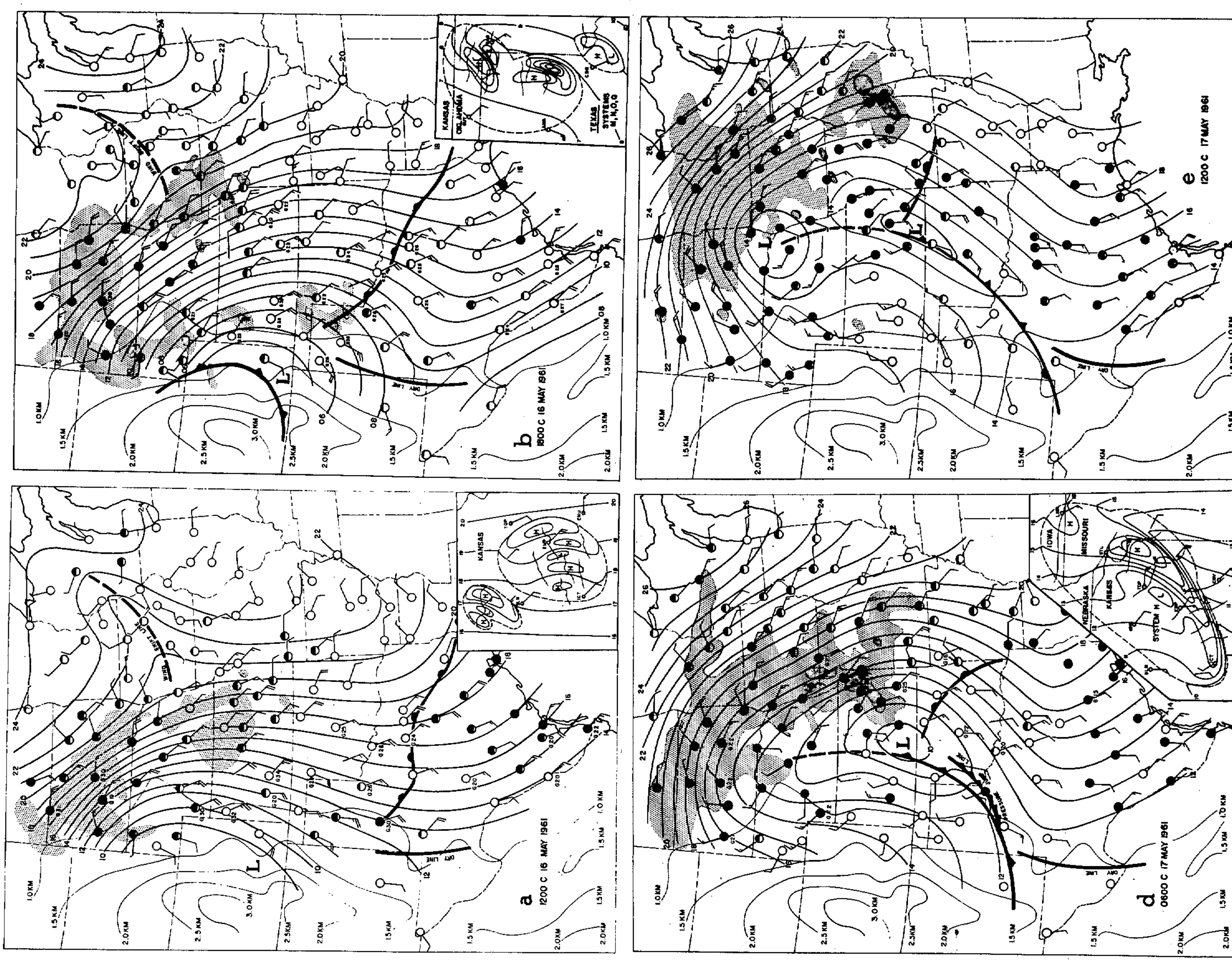

FIGURE 2.—Surface maps. Precipitation areas are shaded. Insets show significant mesosystems. (a) 1200 CST, May 16, 1961; (b) 1800 CST May 16, 1961; (c) 0000 CST May 17, 1961; (d) 0600 CST May 17, 1961; (e) 1200 CST May 17, 1961.

approximate the horizontal pressure gradient at the level of the terrain using a method developed by Fujita and Brown and described by Fujita [5].

3. SURFACE PRESSURE FIELDS, 1200 CST MAY 16 TO 1200 CST MAY 17, 1961

The surface synoptic charts for the period are shown in figure 2. Winds are plotted in standard synoptic form; a full barb represents a 10-kt. wind. Station circles which are not blacked in at all indicate no ceiling (broken or overcast clouds) reported at any altitude at the observation time. At the other stations, the amount of the circle filled in is inversely proportional to the ceiling height; a completely filled-in circle indicates a ceiling at or below 2,500 ft.; a cirrus ceiling is indicated by a thin line at the top. The precipitation areas shown on the map are based on the reports for the hour immediately *preceding* the map time. The pressures shown on the large maps are "undisturbed" pressures, the pressures associated with the mesoscale disturbances having been removed by subtraction. The original analyses in the regions of the mesoscale disturbances are shown in the insets at double the scale of the larger maps.

A low-pressure system moved across the northern edge of the NSSP network on May 17 (fig. 2 c–e). The frontal system associated with the Low was relatively weak, and pressure and temperature fields were much more disturbed by the development of the mesosystems in Oklahoma and Kansas than by the passage of the cold front. The warm front, especially, seemed not to be reflected at all in the pressure traces. The 62° F. dew-point line was used to locate the warm front throughout the period. This line seemed to mark a weak discontinuity in the dew-point gradient and also approximated the northern boundary of the region within which there was very little thickness gradient at low levels. The dry line was drawn as a 55° dew-point line within the warm air. It remained quasi-stationary except for the final map time (fig. 2e).

The cold front was accompanied by both wind shifts and pressure jumps at a number of stations. The pressure jumps were small, 1 to 2 mb. at most. Isochrones of the wind-shift line and of the cold-frontal pressure jump are shown in figure 3. The pressure jump line appeared to move steadily southeastward at 25 to 30 kt. until it disappeared near 0800 CST in central Oklahoma and northern Texas. The last discernible jump was at Wichita Falls, Tex. The wind shift line, however, decelerated rapidly in this area after about 0400 CST, not reaching Wichita Falls at all on the 17th. The temperature gradient behind the cold front was relatively weak, and temperature changes with the frontal passage were often obscured within the diurnal changes at individual stations. Breaks did occur at some stations, however, and the existence of a front was obvious both in the 24-hr. temperature changes (for example, −8° F. at Amarillo from 1300 CST, May 16, to 1300 CST, May 17) and in vertical cross sections. The placing of the front

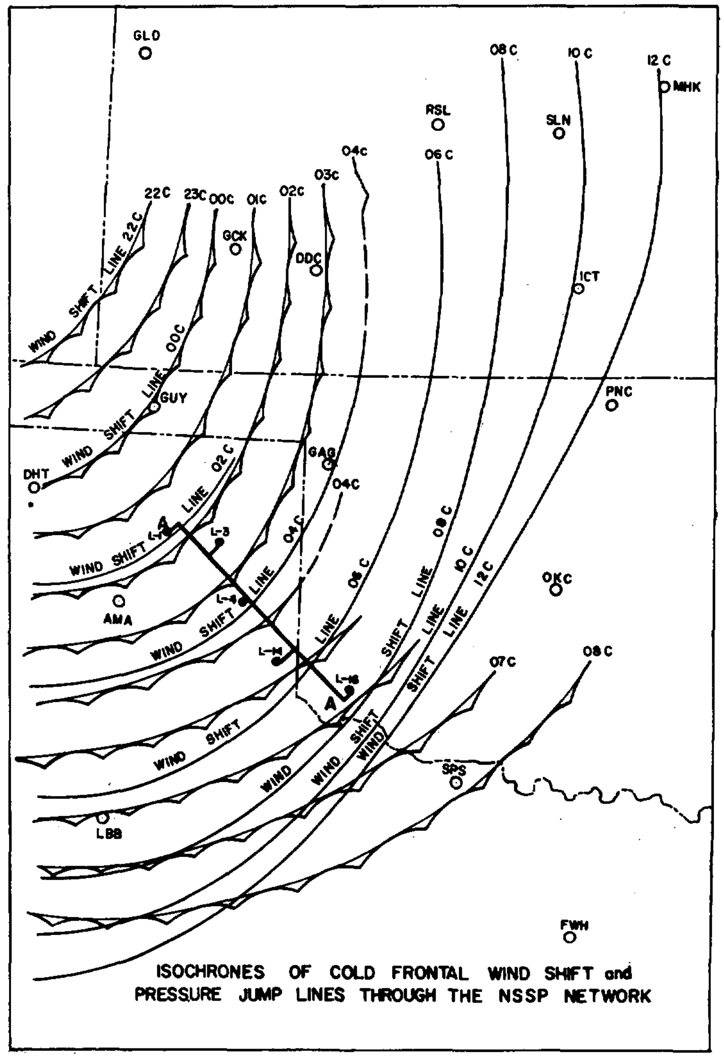

FIGURE 3.—Times of cold frontal passage.

in northern Texas was aided by radar reports of a thin line from the radar station at Amarillo.

The mesosystems analyzed in the inserts were all recognizable systems which could be traced across the map. System H, for instance, appears at both 0000 and 0600 CST on the 17th (fig. 2c and d).

Figures 4 and 5 show the isochrones of the pressure surge associated with each mesosystem which could be identified as a traveling disturbance on the individual pressure traces. System H passed mostly to the north of the NSSP network (fig. 4) but there was no difficulty in picking out the pressure surge and in establishing continuity from the available Weather Bureau traces in Kansas. The systems moving through the α and β networks in Oklahoma were much more difficult to follow. The activity was very intense, yet more sporadic than that farther to the north.

4. PRECIPITATION PATTERNS

During the 24-hr. period under study, precipitation fell over a large region in the Midwest (fig. 6). The greatest

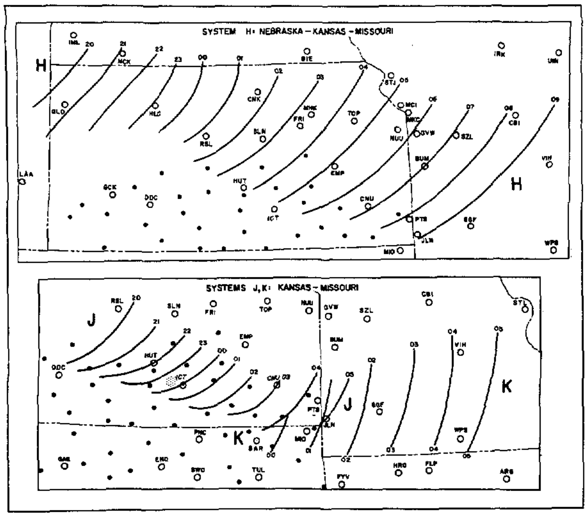

FIGURE 4.—Isochrones of pressure surge lines in Kansas and Missouri.

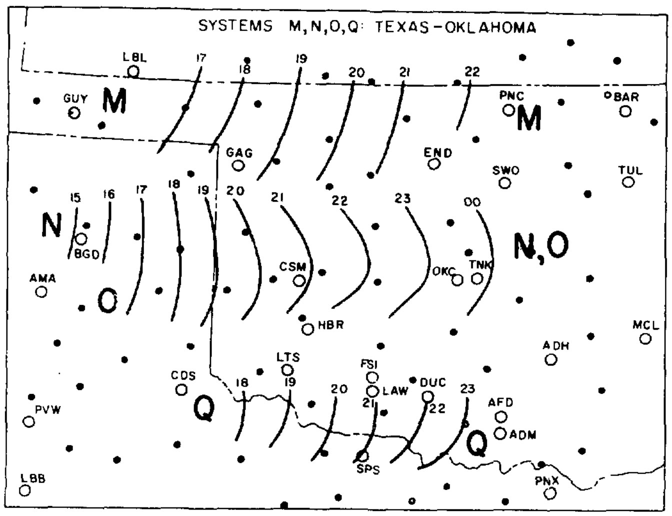

FIGURE 5.—Isochrones of pressure surge lines in Oklahoma and Texas.

amount of rain fell in South Dakota; however, the maximum rainfall intensities occurred in Kansas, Oklahoma, and extreme northwestern Texas. At Aberdeen, S. Dak. for instance, it rained during 22 of the 24 hr. The total rainfall during the period was 1.62 in., and the maximum rainfall in any 1-hr. period was 0.15 in. At Elk City, Okla., the total rainfall of 1.34 in. was accumulated in three consecutive hourly rainfall reports, indicating a rain duration of anywhere from 62 min. to 3 hr.

It is possible to pick out individual rain systems moving across the region by observing the west to east elongation of the regions of maximum rainfall in figure 6. There are at least three systems evident in Oklahoma, and one large system in Kansas and Missouri (cf. figs. 2b and 2d).

Tracks of the individual rainstorms are shown in figure 7. A rainstorm here is defined as any localized center of maximum rainfall which can be traced on successive hourly precipitation maps. Newton and Newton [7] used a similar method of presentation.

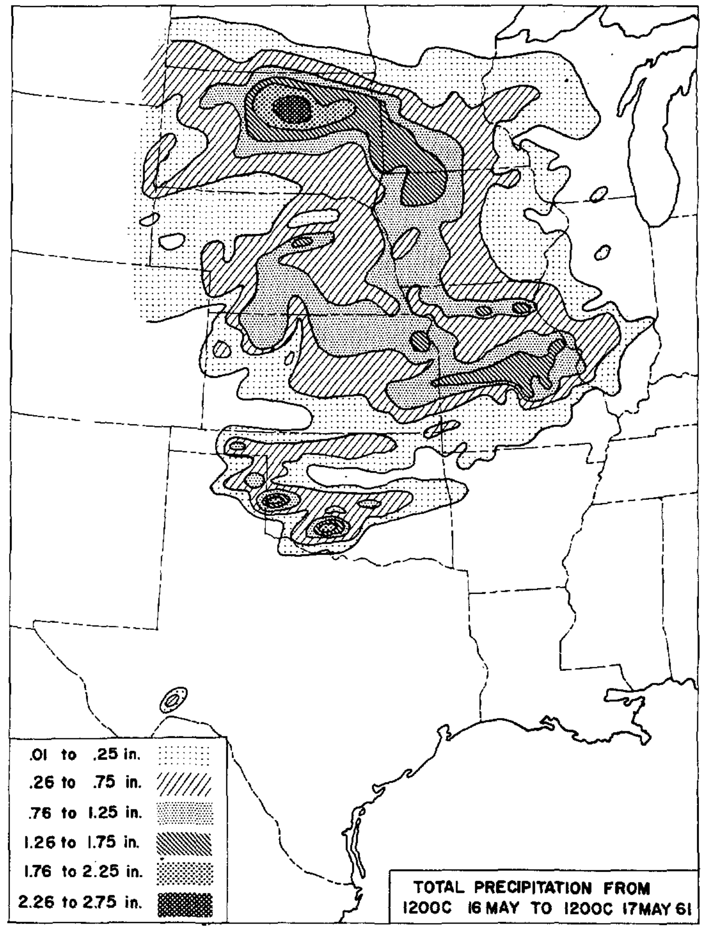

FIGURE 6.—Twenty-four-hour precipitation.

With the exception of the systems moving through Kansas and Missouri (Systems H and J), the duration of the rainstorm centers was relatively short. Newton and Newton [7], for instance, were able to trace an individual rainstorm for as long as 17 hr.; Fujita [5] quotes a figure of 12 hr. for a typical life cycle of a mesoscale disturbance. The maximum *intensity* of precipitation was associated with Systems O and P in Oklahoma; however, their intensity was very short lived compared with System H, which maintained a rain intensity of greater than 0.5 in./hr. for at least 14 hr.

5. FORMATION OF THE THUNDERSTORMS

Most of the activity developed in the middle to late afternoon along the eastern slopes of the Rocky Mountains, well ahead of the approaching cold front. The only system which could have formed along the front was System H. The exact location of the front at 1800 CST (fig. 2b) was not well defined in Colorado and Nebraska; however, it appeared to be about 60 mi. to the west of the center of maximum rainfall at this time. Following the typical pattern, the rainfall maximum moved much more rapidly than the front and was well to the east of the cold front by 1200 CST on the 17th (fig 2e).

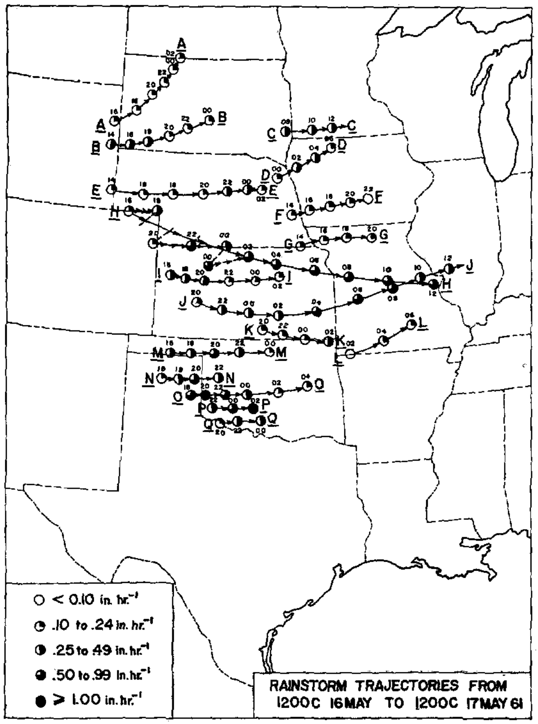

FIGURE 7.—Rainstorm history.

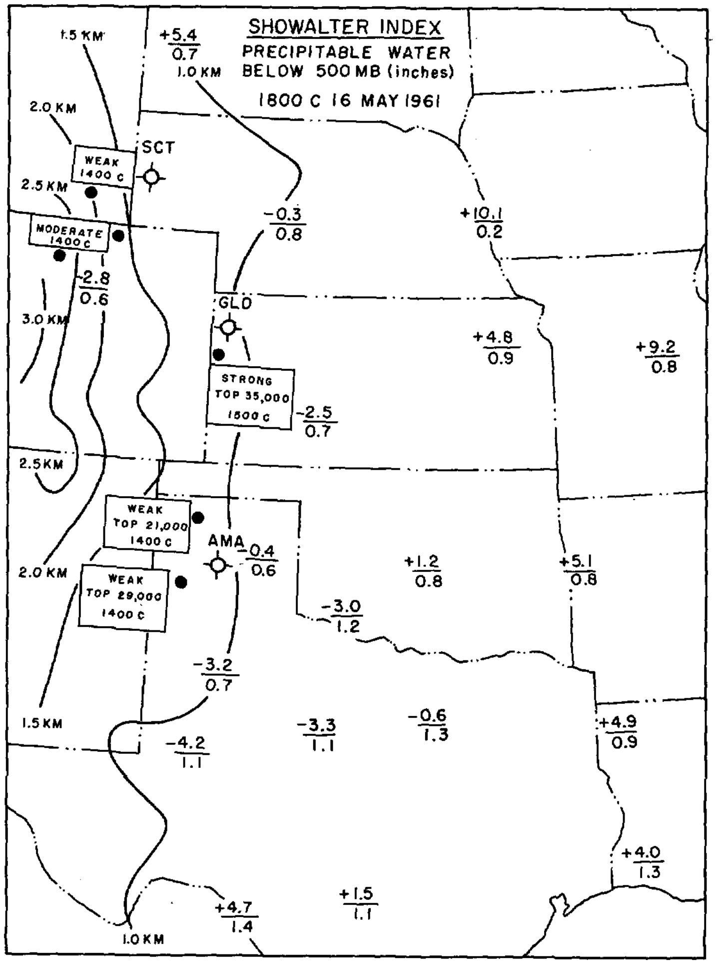

FIGURE 8.—Stability and moisture.

The Showalter Stability indices (cf. Petterssen [9]) and the precipitable water, at 1800 CST, May 16, are plotted in figure 8. Precipitable water was calculated in 50-mb. increments from the surface to 500 mb. Shown as well are the first radar plots of the activity which developed into the intensive systems in Kansas and Oklahoma. It appears that the systems originated over the mountains, in regions where the air was conditionally unstable, but not in the regions of maximum instability or moisture content. The most unstable air, with the highest moisture content, was to be found in west-central Texas where thunderstorms did not develop at all.

6. HORIZONTAL STRUCTURE OF THE JET

Because of small-scale irregularities in the wind speeds near the ground it is often difficult to arrive at a reasonably objective analysis of the low-level jet on constant pressure or constant height charts. Therefore, the horizontal structure of the jet is shown by using vertically averaged winds in the lowest few kilometers. The detailed vertical structure of the jet is examined in the next section.

The vertical integral of the wind from the surface to some fixed height h above sea level may be evaluated directly from balloon displacements on the WBAN 20 forms in the following way (see Panofsky [8]):

$$\int_s^h \mathbf{V}dz=\bar{b}\mathbf{R}=\bar{\mathbf{V}}(h-s) \tag{1}$$

where $\bar{b}$ is the mean rate of ascent of the balloon from ground level to h and $\mathbf{R}$ is the horizontal component of the vector displacement of the balloon. At each station, graphs were plotted of balloon height, scalar horizontal balloon displacement, and azimuth angle of the balloon as a function of time. From these, it was possible to determine the rate of ascent, $\bar{b}$, and the vector displacement, $\mathbf{R}$, at any desired altitude.

The analyzed fields of $\bar{b}\mathbf{R}$ are shown in figure 9 for the two rawinsonde observation times 1800 and 0600 CST. Since the mean density within the warm air mass was practically constant,[2] the vector fields in figure 9 may be considered to represent, to within a constant factor, the mass transport from the ground to the level h.

The mass flux necessarily goes to 0 at the intersection of h with the ground; thus, the apparent wind shear to the left of the jet in Texas, Oklahoma, and Nebraska is partly the result of the increasing terrain elevations. The actual wind shears can be estimated from the cross sections (figs. 10 and 11) or from figures 17 and 18 which

[2] The variation in mean density from the surface to 2.5 km. at 0600 CST May 17 between Oklahoma City and Bismarck, N. Dak., was only 1 percent.

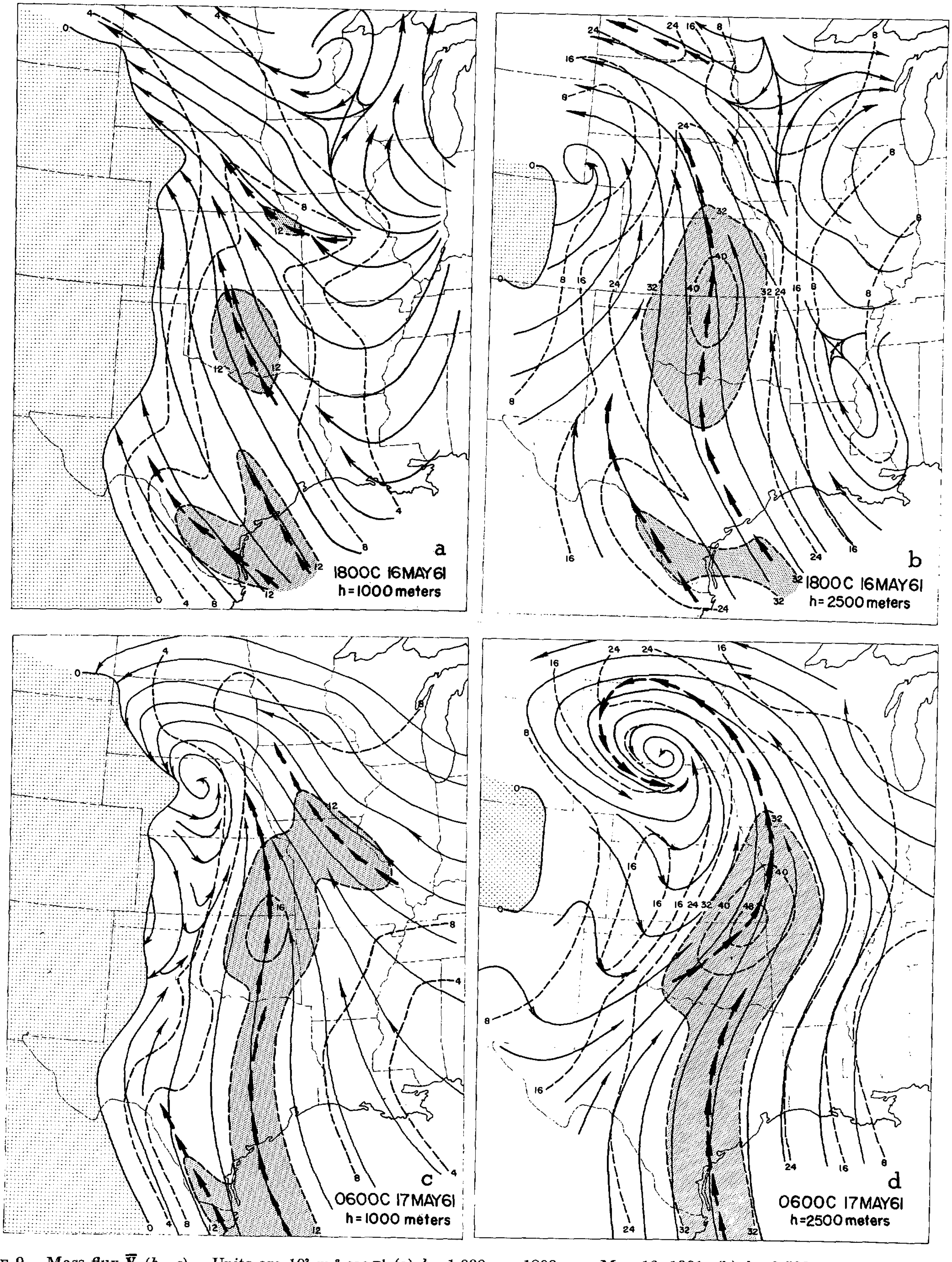

FIGURE 9.—Mass flux $\overline{\mathbf{V}}\ (h-s)$. Units are 10^3 m.^{2} sec.$^{-1}$ (a) h=1,000 m., 1800 CST, May 16, 1961; (b) h=2,500 m., 1800 CST May 16, 1961; (c) h=1,000 m., 0600 CST May 17, 1961; (d) h=2,500 m., 0600 CST, May 17, 1961.

show the mean winds from the surface to h in a jet coordinate system.

There was a definite jet structure to the flow at 1800 CST May 16 as well as at 0600 CST the following morning. At 1000 m., this structure was somewhat incoherent on the evening map with indications of three, or perhaps four, distinct wind maxima. During the night mean winds rotated in a clockwise direction and the zone of maximum transport moved slightly to the east ahead of the advancing front.

7. CROSS SECTIONS

Figures 10 through 13 are cross sections (along the broken line of fig. 14) of potential temperature, mixing ratio, and total wind speed in the vicinity of the jet, at 1800 and 0600 CST.

At 1800 CST, the wind field below 3500 m. was fairly complex, with a double jet structure in the vertical over the entire area west of Little Rock. The lower jet over Texas and Oklahoma, shown as two separate wind maxima, would still not be continuous if the cross sections were taken along a straight line. The level of maximum wind paralleled the ground at an altitude of 500 to 800 m.; however, the wind maximum was apparently higher on the morning chart. Wind speeds in the jet increased during the night, as expected, from roughly 45 to 60 kt.

Lapse rates through the jet at 1800 CST were essentially dry adiabatic. An inversion surface overlay the lower of the two jets. At 0600 CST, the stability had increased markedly throughout the entire layer. The jet, however, did not lie along an inversion surface; it occurred in a zone where the air was stable but where the temperature was still decreasing with height.

Mixing ratios were highest to the west of the jet core. The vertical extent of the moist layer on the left flank indicates upward motion in this region.

The depth of the moist layer and the amount of moisture in Oklahoma and Arkansas increased during the night (compare figs. 12 and 13) even though the thunderstorm activity was dying out in Oklahoma during this period. The dissipation of the thunderstorms cannot, then, be attributed to a drying out process, but must be related to some other stabilizing factor.

8. VERTICAL MOTIONS

Vertical motions were computed from the data on the WBAN 20 Forms and the equation of continuity, using a method outlined by Panofsky [8]. Briefly, the vertical velocity at any height h above sea level can be obtained from the divergence of the vector field $\bar{\delta}\mathbf{R}$ plotted in figure 9.

As pointed out by Panofsky, this method of evaluating the vertical velocities has several advantages over the standard kinematic approach:

1. It makes calculation of the surface vertical velocities unnecessary since the effects of changes in terrain height are automatically incorporated in spatial differentiation of the $\bar{\delta}\mathbf{R}$ field.

2. It eliminates the errors involved in re-integrating the vertical winds, which are obtained by graphical differentiation of the balloon displacements.

The mathematical expression for the vertical velocity at some height h above sea level is

$$w_h = -\frac{\bar{\rho}}{\rho_h} \nabla \cdot \bar{\delta}\mathbf{R}$$

where ρ_h is the density at h, $\bar{\rho}$ the mean density from the surface to h.

The divergence calculations were carried out within overlapping squares, 300 km. on a side. The mean value of the normal component of the wind along each side of the square was determined by averaging over three points, with the wind at the central point double weighted.

The vertical velocities do not, of course, represent point values, but averages over areas of (300 km.)^2 The relationship between small-scale and synoptic-scale vertical velocities is not obvious; however, the general assumption has been that where the synoptic-scale vertical velocities are large, the presence of individual cells of strong upward motion is more likely (Curtis and Panofsky [3]). Computed vertical velocities at 1800 CST May 16 and 0600 CST May 17 are shown in figures 15 and 16.

At 1800 CST vertical velocities at 2500 m. were generally positive and of the order of 1 or 2 cm. sec.$^{-1}$ in the region of continuous precipitation in South Dakota and Iowa. The air was ascending at the rate of 5 cm. sec.$^{-1}$ or more in the vicinity of thunderstorms in Nebraska, Kansas, Oklahoma, and Texas. At 0600 CST the largest positive velocities occurred near the squall line in southern Iowa, Kansas, and Missouri. At both times, the air ahead of the cold front was generally ascending as would be expected; however, within the region of ascent there were pronounced variations associated with the isotach maximum representing the low-level jet.

Figures 17 and 18 show vertical velocities at 1800 and 0600 CST in a jet coordinate system oriented along the axis of maximum mass flux below 2500 m. Vertical velocities are at 2500 m. above sea level. Mean winds from the surface to this level were computed by dividing $\bar{\delta}\mathbf{R}$ by $(h-s)$ (see equation (1)) at coordinate points. Average isotachs were then analyzed in the jet coordinate systems. Maximum ascent occurred downstream from the jet maximum and to the left of the axis of the jet. The right rear quadrant of the jet was a region of descent at 1800 CST and of relatively weak ascent on the morning chart. Rainstorms occurring at the map times were taken from figure 7 and plotted in the jet coordinate system. There was no thunderstorm activity at all in the right rear quadrant of the jet.

The vertical velocity field offers an explanation for the occurrence of nocturnal thunderstorms in Missouri and Kansas, and at the same time an explanation for the dying out of the intense thunderstorm activity which developed

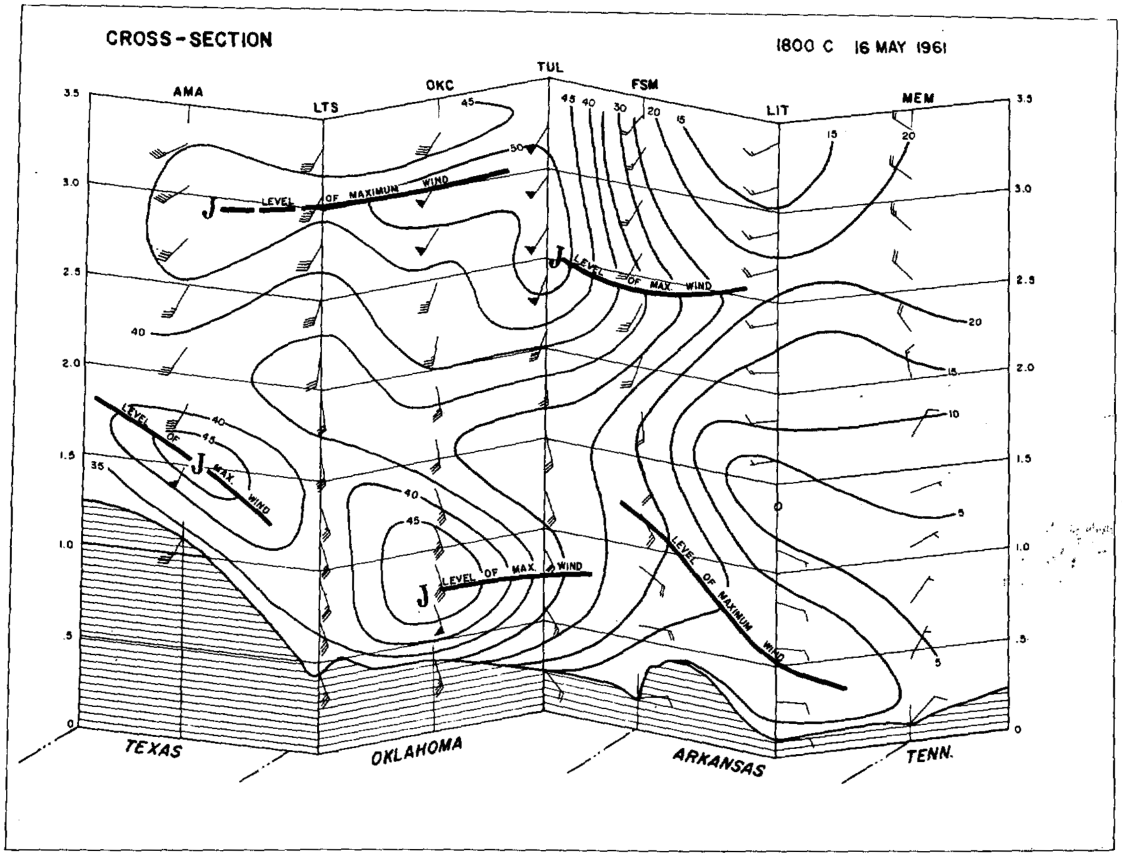

FIGURE 10.—Cross section of wind speed in knots, 1800 CST, May 16, 1961. See figure 14.

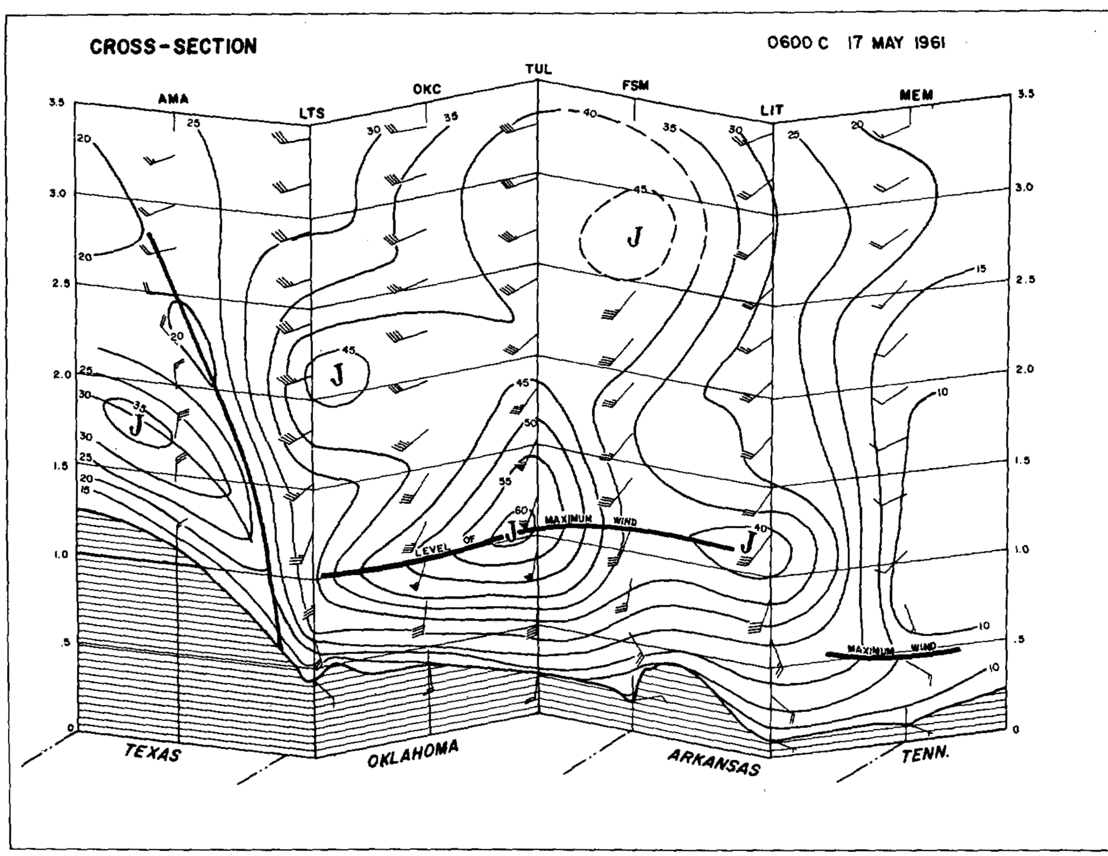

FIGURE 11.—Cross section of wind speed, 0600 CST, May 17, 1961. See figure 14.

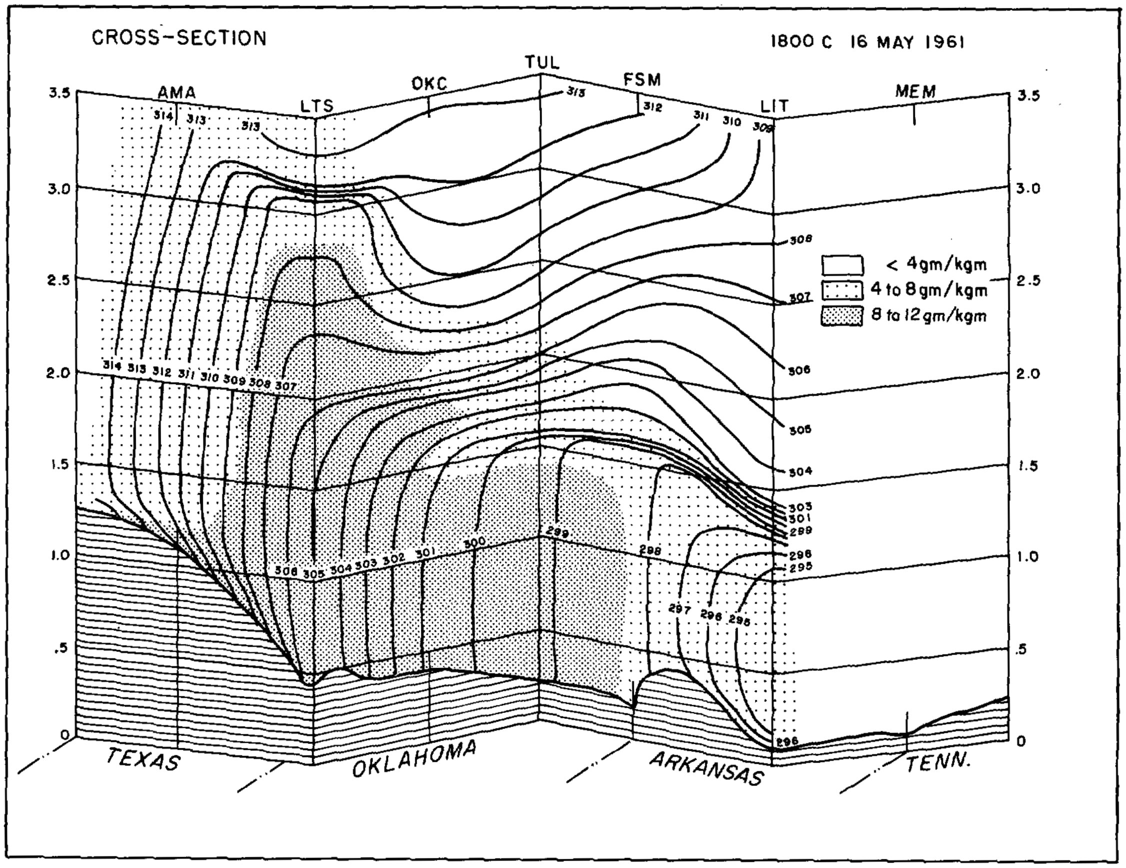

FIGURE 12.—Cross section of potential temperature and mixing ratio, 1800 CST May 16, 1961. See figure 14.

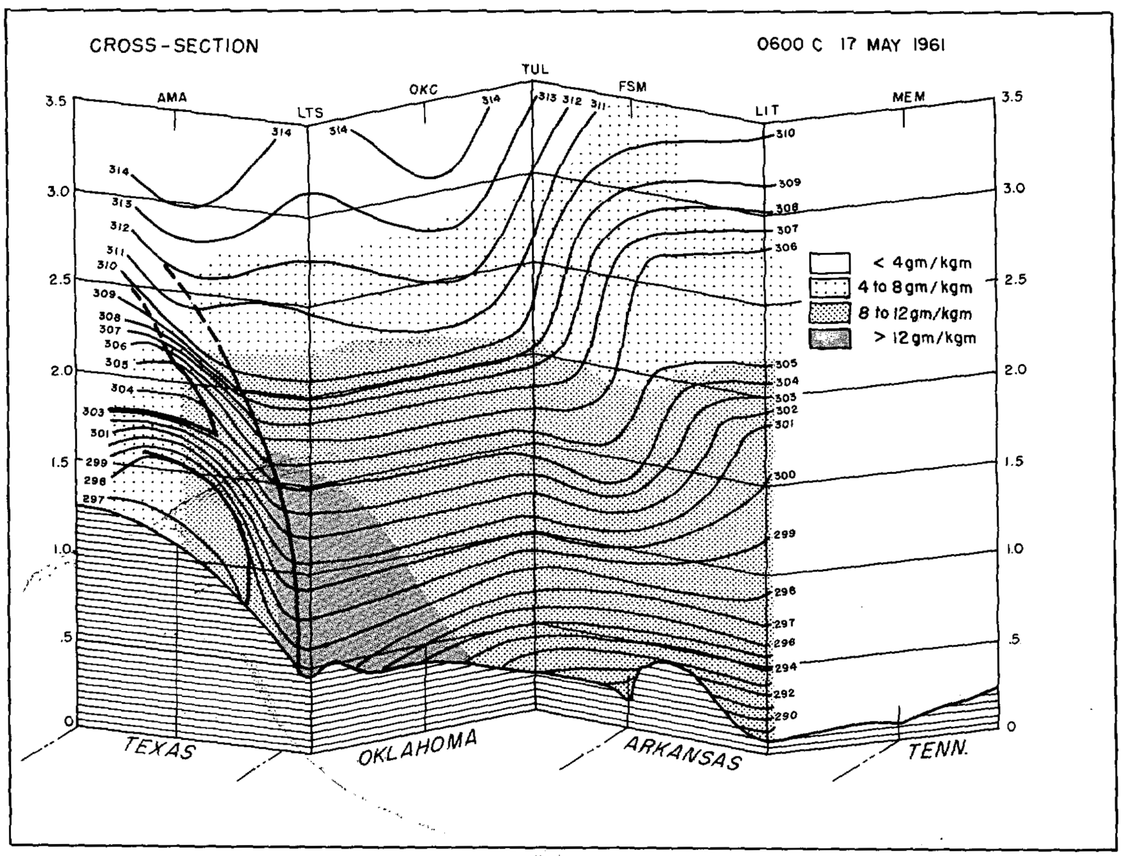

FIGURE 13.—Cross section of potential temperature and mixing ratio, 0600 CST May 17, 1961. See figure 14.

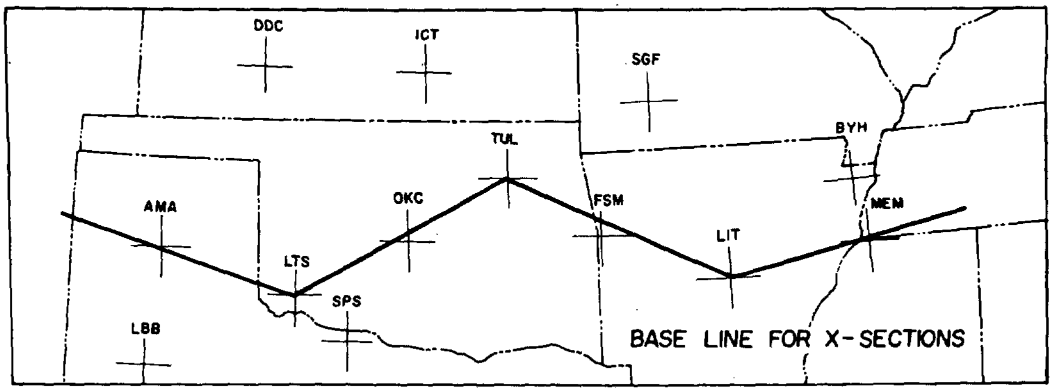

FIGURE 14.—Base line for cross sections in figures 10 through 13. AMA=Amarillo; LTS=Altus; OKC=Oklahoma City; TUL=Tulsa; FSM=Fort Smith; LIT=Little Rock; MEM=Memphis.

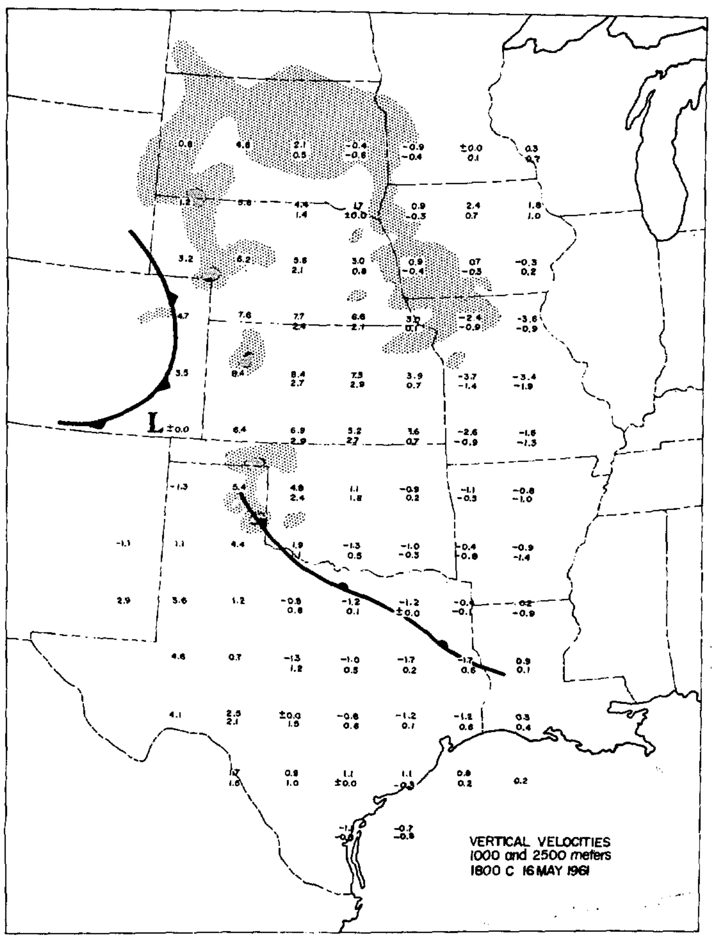

FIGURE 15.—Vertical velocities in cm. sec.$^{-1}$ at 1,000 m. (lower) and 2,500 m. (upper) above sea level at 1800 CST May 16, 1961. Regions of precipitation are shaded.

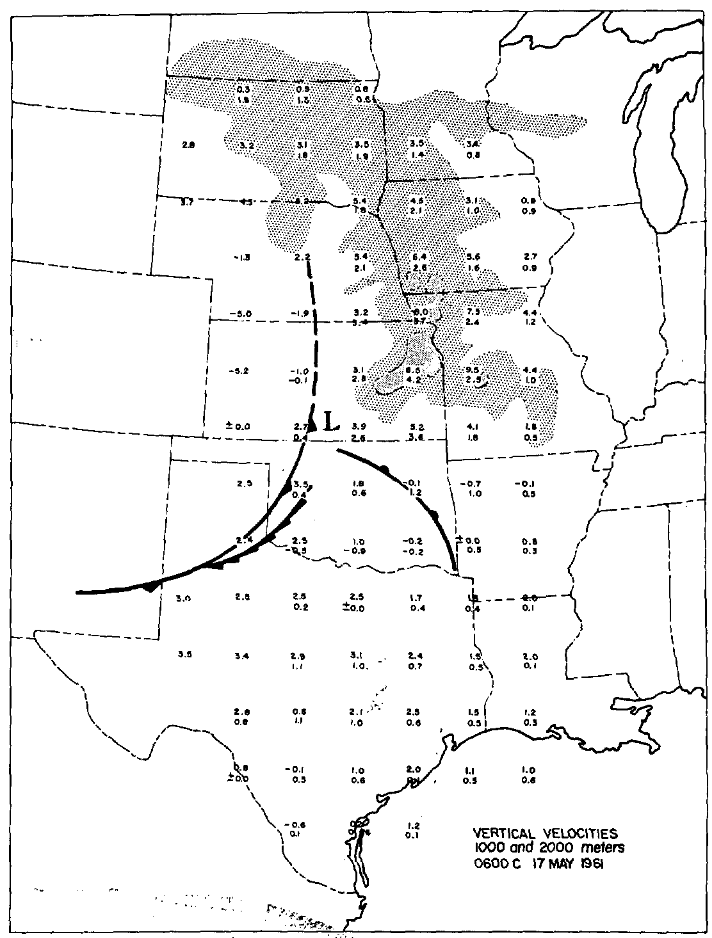

FIGURE 16.—Vertical velocities as in figure 15, 0600 CST May 17 1961.

in northern Texas and moved into Oklahoma during the evening. These thunderstorms formed on the downwind side of the jet but far enough to the left of the axis to be in a region of strong ascent. The axis of the jet in Oklahoma was nearly stationary and the squall line apparently moved through the jet axis and into the general region of subsidence in the right rear quadrant of the jet. Systems forming farther to the north, however, remained within the general region of ascent on the downstream side of the jet and these systems persisted, causing strong nocturnal thunderstorms in Kansas and Missouri. This describes the low-level jet not as a cause

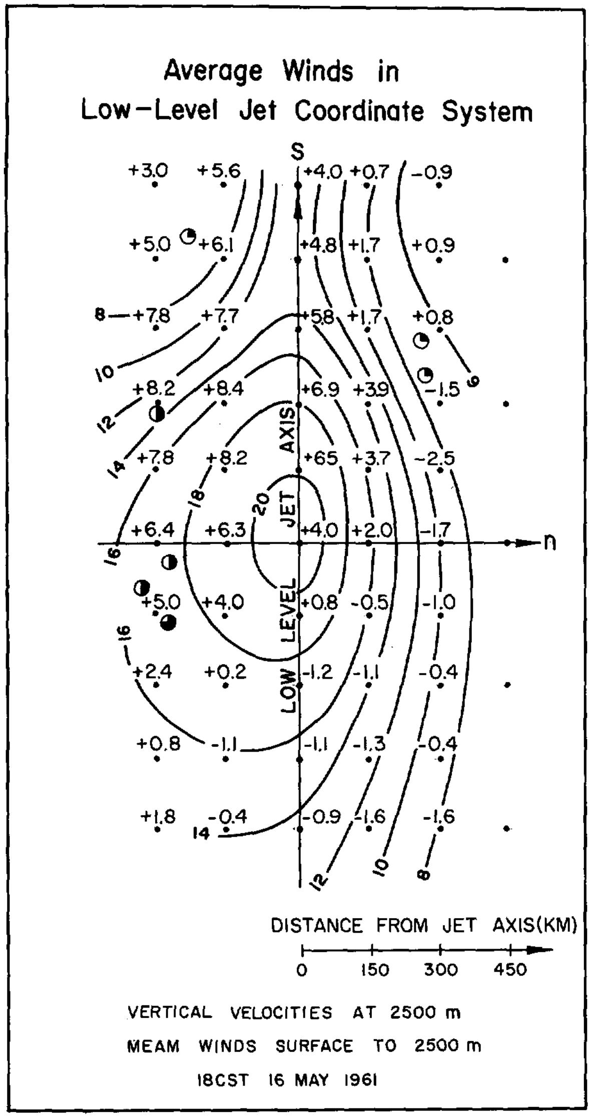

FIGURE 17.—Vertical velocities at 2,500 m. in jet coordinate system at 1800 CST May 16, 1961 and isotachs of mean wind in m. sec.$^{-1}$.

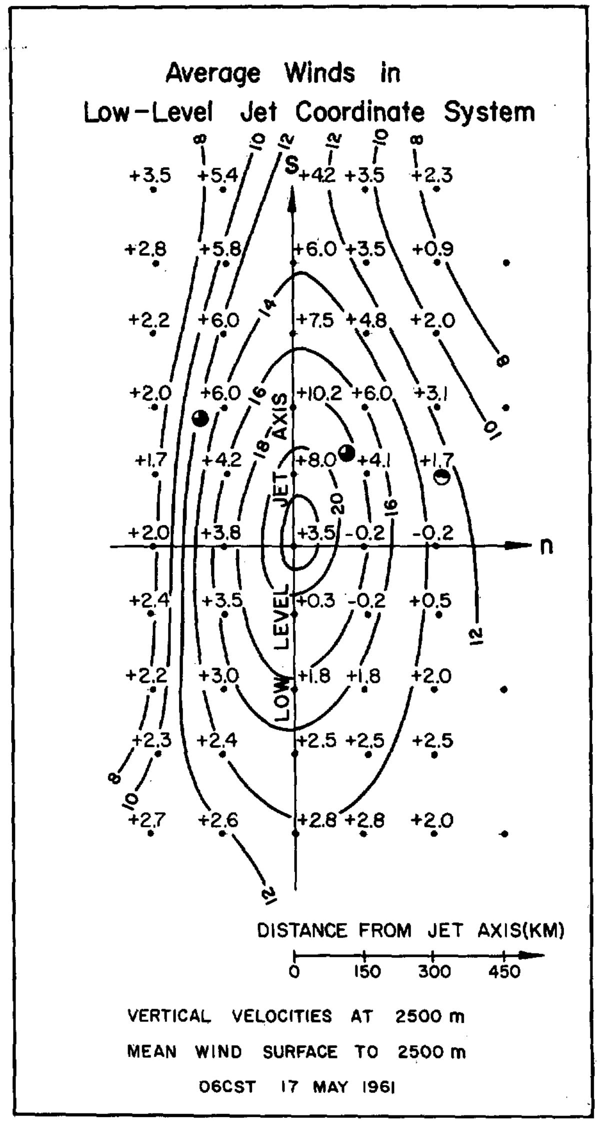

FIGURE 18.—Vertical velocities as in figure 17, 0600 CST May 17, 1961.

for the development in situ of nocturnal thunderstorms over the Great Plains but rather as a mechanism for prolonging the lives of squall lines derived from frontal activity or convection over the mountains.

Pitchford and London [10] have noted the same distribution of vertical velocities along the axis of the low-level jet as indicated here.

The correction coefficient between 1800 and 0600 CST vertical velocities at 2500 m. is approximately zero for points in the same geographical location. In the jet coordinate system, however, the coefficient is 0.70 indicating that, when a large-scale jet of the type described here is present, vertical velocities at particular points are largely determined by their location with respect to the center of maximum wind. This effect apparently overrides any diurnal oscillation of vertical velocities of the type described by Blackadar [1].

9. CONCLUSION

Means [6] demonstrated the importance of the low-level jet in advecting warm, moist air into the Great Plains. This effect is apparent from a comparison of figures 12 and 13. The 850 mb. temperatures in the vicinity of Tulsa, Okla., rose by 4° C., and the mixing ratios increased by 3 gm./kg. during the 12-hr. period from 1800 CST May 16 to 0600 CST May 17. Similar increases in moisture were observed farther to the north in Kansas, Nebraska, and Missouri. Between 1800 CST May 16 and 0600 CST May 17, mean temperatures in the layer from 850 to 900 mb., roughly the zone of maximum low-level wind, rose by an average of 2° C. along a narrow north-south band, the axis of which is approximately the jet axis in figure 9d. Temperature or moisture advection by the jet is not a sufficient condition for the development or persistence of thunderstorms, however, since the intense activity in Oklahoma died out during the night, while the thunderstorm activity in Kansas persisted, even though the maximum 12-hr. increases in temperature and moisture at low levels took place in eastern Oklahoma.

Tepper [12] suggested that nocturnal thunderstorms may form along pressure jump lines generated by the formation of a low-level jet. However, in this case the pressure jump lines developed during late afternoon, before the nocturnal increase in wind speed. Pressure surge lines appear to be a result of intense convection rather than its cause (Fujita [4]).

The large-scale vertical velocity patterns in the vicinity of the low-level jet provide the most likely explanation for the persistence of thunderstorms in Kansas, Nebraska, and Missouri and their dissipation in Oklahoma and Texas. This agrees with Bleeker and Andre [2], Blackadar [1], Sangster [11], and Pitchford and London [10] in attributing the crucial role in nocturnal thunderstorm occurrence over the Midwest to low-level convergence and associated vertical velocity fields. However, these fields may be important not in initiating convection at night but in extending the lifetimes of squall lines which form during the day.

Statistical studies like that of Pitchford and London [10] can only show that nocturnal thunderstorms and low-level jets occur in the same region or at the same time. A series of case studies would help to determine the relative importance to nocturnal thunderstorm occurrence of such factors as in situ development, development over the mountains, and the position and intensity of the low-level jet.

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