Air Mass Modification and Upper-Level Divergence

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Abstract

A procedure for estimating the degree of modification of air mass structure as a result of favorable divergence-convergence patterns is discussed. Its application to the forecasting of severe local storms is discussed.

1. Introduction

FAWBUSH and Miller [1] and Showalter [2] have presented mean or typical soundings representative of the tornado air mass environment. Subsequently, Beebe [3] reported his findings in connection with the analysis of soundings taken in the vicinity of tornadoes. It was this latter finding that led Beebe and Bates [4] to propose a model of jet structures which assists in, or in some cases possibly effects, the release of convective instability through vertical stretching or lifting. In support of this, these writers have shown that by considering certain indications of the vorticity equation, it is possible to analyze configurations of jet axes and jet maxima such that low-level (850 mb) convergence is surmounted by higher-level (500 mb) divergence. The earlier investigations of Fawbush and Miller [5] of many tornado-producing synoptic situations had pointed out the existence of narrow bands of strong winds in the middle troposphere and this parameter was incorporated in their forecasting procedures [6].

Since the publication of these papers, much more attention has been focused upon the problem of forecasting severe local storms and more attention has been given to wind flow patterns at higher and higher levels. In addition, improvement in wind-measuring equipment has resulted in more data in the levels between twenty and forty thousand feet that were not available at the time the aforementioned persons were making their investigations. These data now permit a more complete evaluation of the vertical distribution of divergence and the related vertical motion fields.

Beebe and Bates stress the point that when low-level convergence is surmounted by higher-level divergence, the result is a vertical motion field that will modify air mass structures. The result, in the case of an air mass structure of marked convective instability with dry air above a low-level moist layer, is to permit moisture to penetrate to greater heights and at the same time eliminate any inversion, thus resulting in marked parcel instability. The forecast problem is one of determining how long it will take a favorable convergence-divergence pattern to act to accomplish the foregoing. Proceeding upon the basis that it is possible to measure horizontal divergence and determine its vertical distribution, then it is also possible to determine the strength of the vertical motion field. Given a reasonably accurate prediction of the time interval, the vertical motion field will act over a specified area, then it should be possible to predict the future lapse rate structure and stability.

The work of Beebe and Bates placed emphasis upon certain combinations of low (850 mb) and high (500 mb) jet structures. An investigation of forecast failures resulting from a utilization of these models at the levels suggested revealed that: 1. the 500-mb level did not accurately portray the location of the significant jet stream and its branching structures all seasons of the year, 2. frequently the low-level jet either did not exist or was not clearly defined, and 3. important divergence centers were detectable that were not directly associated with either jet.

In an effort to determine more precisely the divergence patterns at various levels, utilization of the real wind rather than the geostrophic wind was made and, as should be expected, areas of strong divergence aloft were found in wind fields that were not attended by strong speeds but were characterized by large streamline divergence. It was observed that frequently in the warm seasons, severe weather and tornadoes occurred in association with divergence centers that could not have been predicted by the usual methods.

It is the purpose of this paper to compare the modification possible in air mass structure due to vertical motion in the vicinity of severe local storm occurrence under two contrasting synoptic situations.

2. Procedure

A number of investigators concerned with precipitation forecasting have studied procedures for computing vertical motion. In general, these investigators were concerned with procedures that could be utilized with an area and time scale much larger than that associated with the severe storm phenomena, consequently the procedures used filtered out some of the small-scale circulations that are apparently important in severe storm forecasting.

Such a procedure was suggested by Bellamy [7] and applied by Collins and Thompson [8] to a quantitative precipitation forecasting technique. This technique involved the use of a triangular grid, the vertices of which were located at pilot balloon stations.

In applying this technique to the problem of forecasting severe local storms, it was ascertained that, as a general rule, the area over which computations were being made was too large to insure that the magnitude and location of divergence centers were determined with sufficient accuracy. Consequently the evaluation of divergence as expressed by the relationship

$$\operatorname{div}_2 \mathbf{V} = \frac{\partial V}{\partial s} + \frac{V\pi\, \partial\alpha}{180\, \partial n} \tag{1}$$

was utilized. This expression is derived by the use of "natural" coordinates where s is defined as a coordinate parallel to a streamline and positive in the direction of the wind, n a coordinate at right angles and positive to the right of the wind looking downwind, V is wind speed, and a the wind direction in degrees increasing clockwise from the north. The utilization of this expression permits the computation of divergence at any point using any convenient length unit for ∂s and ∂n once an analysis of streamlines and isotachs has been completed. Whereas the technique described by Bellamy tends to underestimate the magnitude of divergence in the vicinity of points where wind directions and/or speeds change rapidly in short distances, this technique provides a procedure that will partially overcome this shortcoming. The technique inherently implies subjectivity in the analysis of streamlines and isotachs but recent studies conducted by the Severe Local Storm Forecast Center indicates that evaluation of divergence is in good agreement with observed occurrences of severe local storms.

Computations of divergence can be made for any convenient height interval. In the cases to be presented later this was done for 2,000-ft intervals. To facilitate the computations, ∂s and ∂n were chosen to be 2.3 deg lat in length.

The equation of continuity, neglecting local and horizontal variation of density, is

$$\frac{\partial \rho w}{\partial z} = -\rho \operatorname{div}_2 \mathbf{V}, \tag{2}$$

where ρ is density and ω is vertical velocity. Integration of (2) between any two levels Z_1 and Z_2 gives:

$$\rho_2 w_2 = \rho_1 w_1 - \overline{\rho \operatorname{div}_2 \mathbf{V}}(z_2 - z_1), \tag{3}$$

where the bar represents the average value in the layer $(Z_2 - Z_1)$. Substituting

$$\overline{\rho \operatorname{div}_2 \mathbf{V}} = \tfrac{1}{2}(\rho_1 \operatorname{div}_2 \mathbf{V}_1 + \rho_2 \operatorname{div}_2 \mathbf{V}_2) \tag{4}$$

into (3) and dividing by ρ_2 gives the vertical velocity at the top of any layer

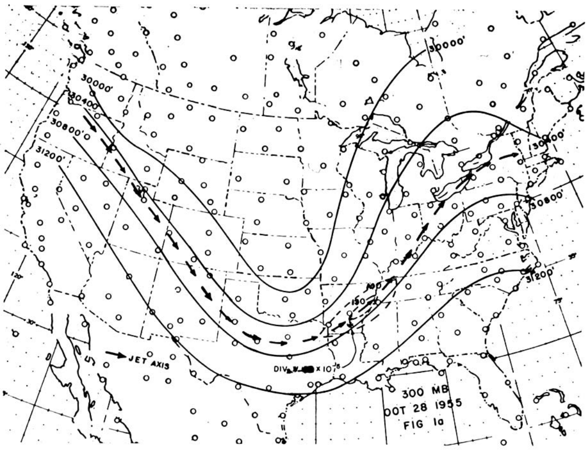
$$w_2 = \frac{\rho_1}{\rho_2} w_1 - \frac{1}{2}\left(\frac{\rho_1}{\rho_2} \operatorname{div}_2 \mathbf{V}_1 + \operatorname{div}_2 \mathbf{V}_2\right)(z_2 - z_1). \tag{5}$$

The ratio of ρ_1/ρ_2 for any layer can be approximated with sufficient accuracy by use of the U. S. Standard Atmosphere densities, and varies from 1.061 for the layer sea level to 2000 ft to 1.078 for the layer 32,000 to 34,000 ft.

Values of divergence determined from equation (1) can be substituted into equation (4) and the vertical velocity distribution determined.

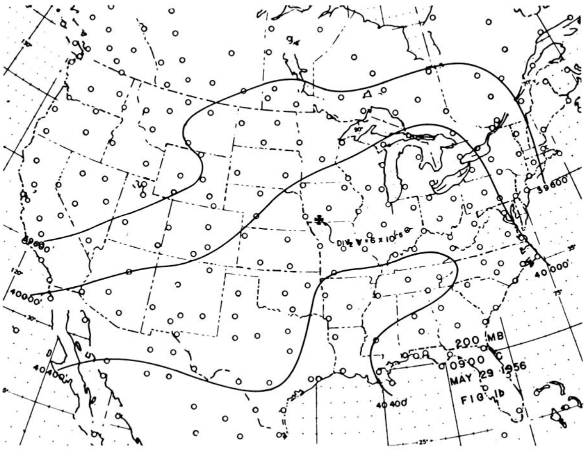
3. Discussion of results

Figs. 1a and 1b represent two contrasting upper-level synoptic situations, both of which are associated with the occurrence of severe weather. In each case the Maltese cross marks the location of the storms which occurred within one hour after the synoptic time of the charts. Fig. 1a is representative of a cold season situation wherein a pronounced jet stream maxima moves out of a trough position. It is typical of one of the upper-level jet stream configurations suggested by Beebe and Bates [4], as being associated with development of severe thunderstorm activity. Fig. 1c is representative of a warm season synoptic situation wherein the major contribution to divergence at the upper levels is from streamline divergence $\left(\frac{V\pi\, \partial\alpha}{80\, \partial n}\right)$ with little or no contribution from speed



(a)

Fig. 1a. 300-mb chart of 0900C, 28 October 1955. Solid lines are height contours. Arrows depict the location of the jet stream with jet maxima indicated by dashed lines. The Maltese cross indicates computational point and location of severe storms within one hour after the synoptic time.



(b)

Fig. 1b. 200-mb chart, 0900C, 29 May 1956. Solid lines are height contours. Maltese cross indicates computational point and location of severe storms, within one hour after synoptic time.

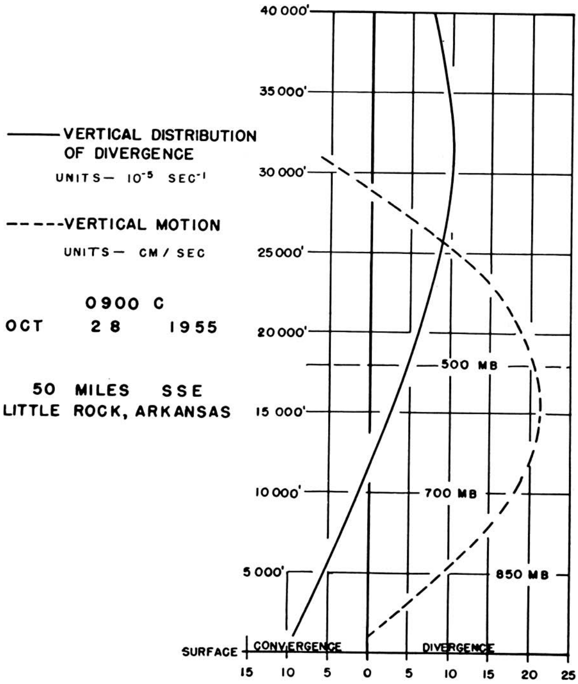


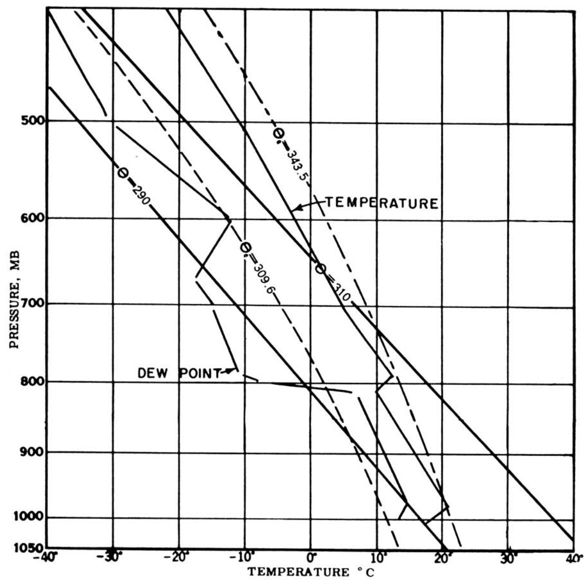
Fig. 2. Vertical distribution of divergence (solid curve) and vertical velocity (dashed curve) at 0900C, 28 October 1956, 50-mi SSE of Little Rock. Units of divergence are in 10^{-5} sec^{-1} positive (divergent) to the right of zero ordinate and negative (convergent) to the left. Vertical motion is in units of cm sec^{-1} positive (upward) to right of zero ordinate and negative (downward) to the left.

divergence $\left(\dfrac{\partial V}{\partial s}\right)$. This latter example is also quite typical of those severe thunderstorm producing synoptic situations that are not associated with well marked jet streams and are actually characterized by low wind speeds (less than 50 kn) at any level over and in the vicinity of the activity.

Computations of divergence and vertical velocities were accomplished for a point fifty miles SSE of Little Rock, Arkansas, from the 0900C upper-air data of 28 October 1955. Several tornadoes were reported within the next three hours in the vicinity of this location.

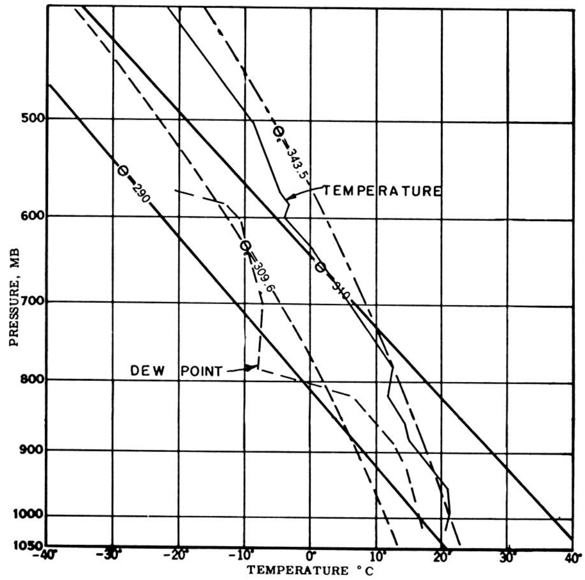
Fig. 2 shows the distribution of divergence with height as well as the computed vertical motion distribution. Fig. 3 shows the Little Rock, Arkansas, soundings for 2100 CST, 27 October and 0900C, 28 October. Fig 4 shows the Shreveport, Louisiana, soundings at 2100 CST, 27 October and 0300 CST and 0900 CST of 28 October. These latter soundings are presented solely to indicate the time period within which the lapse-rate modification at Little Rock took place. From these latter soundings it is concluded that the modification in the air mass over southern Arkansas occurred during the time interval 0300 CST–0900 CST, which coincides with the time of passage of the jet maximum over the area.

In an effort to determine more closely the approximate time interval in which the modification was accomplished, the vertical motion pattern computed as a consequence of the vertical distri-



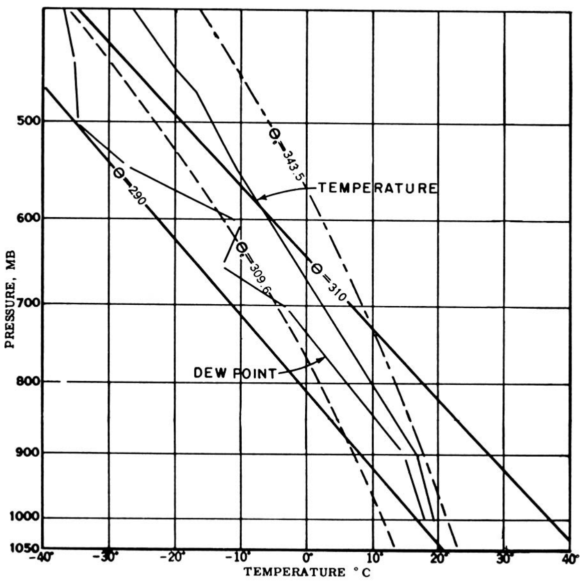
(a)

FIG. 3a. Little Rock, Arkansas, upper air sounding for 2100C, 27 October 1955.



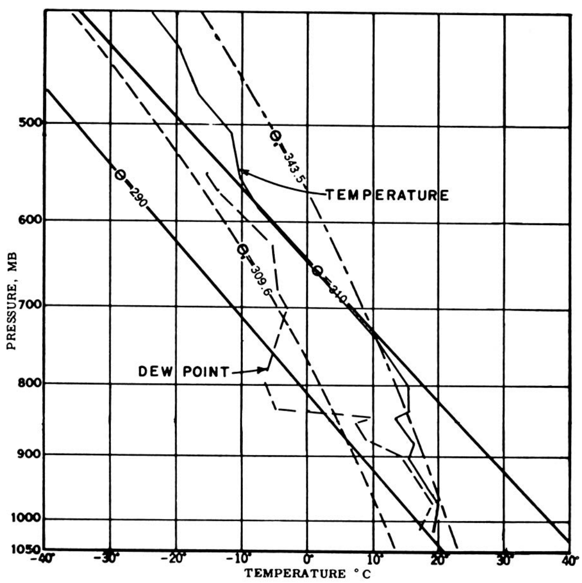
(a)

FIG. 4a. Shreveport, Louisiana, upper air sounding for 2100C, 27 October 1955.



(b)

FIG. 3b. Little Rock, Arkansas, upper air sounding for 0900C, 28 October 1955.

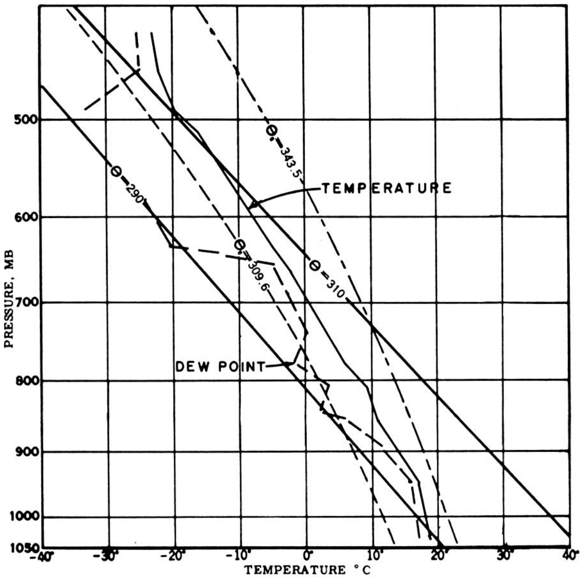


(b)

FIG. 4b. Shreveport, Louisiana, upper air sounding for 0300C, 28 October 1955.

bution of divergence shown in fig. 2 was applied to the 2100C Little Rock sounding and it was determined that with the vertical motion pattern predicated, the inversion could be eliminated in one hour and thirty minutes. Fig. 5 shows the modification in the air mass at a point 50-mi SSE of Little Rock by the action of the vertical motion pattern and it is compared to the actual 0900C Little Rock temperature sounding. Note the similarity between the two.

It is not proposed that the dynamic cooling process was responsible for the total change, however, it is believed in this instance to have contributed the most to the modification. Advective



(c)

FIG. 4c. Shreveport, Louisiana, upper air sounding for 0900C, 28 October 1955.

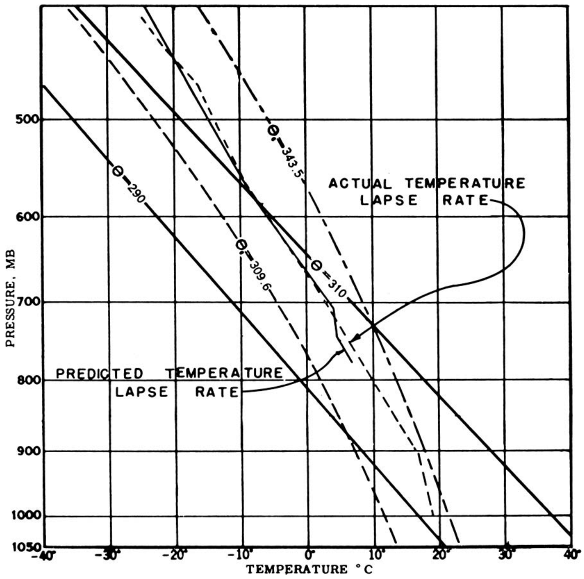


FIG. 5. Comparison of actual temperature lapse rate (dashed curve) at 0900C, 28 October 1955 over Little Rock, Arkansas, with temperature lapse rate (solid curve) predicted for a point 50-mi SSE of Little Rock, Arkansas, as a consequence of the vertical motion distribution indicated in fig. 2, acting for one hour and 30 mins.

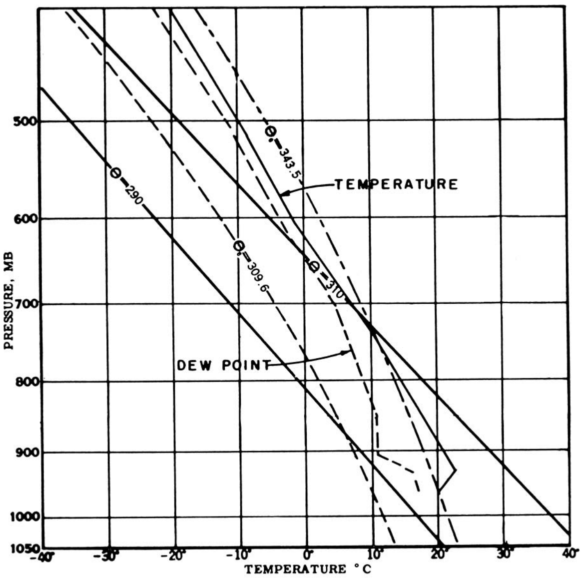
processes at work preceding the 0900 CST observation were towards slight warming (0.2C per hour) at the 850- and 500-mb levels. Neutral to slight cold advection was indicated at the 700-mb level. If only the advective processes were to be considered one would have concluded that little change in stability would have occurred. The Showalter index changed from + 3 at 2100C to a value of − 3 at 0900C.

Similar computations of divergence were made for a point 50-mi SW of Des Moines, Iowa, from 0900 CST data of 29 May 1956. Several tornadoes were reported in this vicinity within the next two hours.

Fig. 6 shows the 0900 CST sounding and a 1045 CST special sounding from Omaha and fig. 7 shows the distribution of horizontal divergence with height as well as the computed vertical motion distribution. Fig. 8 shows the modification that would result in the temperature lapse rate at a point 50-mi SW of Des Moines as a consequence of the vertical motion pattern acting for one hour and forty-five minutes. This modification is compared to the actual 1045 CST Omaha temperature lapse rate. Note that at the 600-, 500- and 400-mb levels, no change in temperature has occurred between the 0900 CST and 1045 CST, while the dynamic modification process indicates that about 1.0C cooling should have occurred. During this time interval, warm advection was indicated at these levels at the approximate rate of 0.5C per hr. The conclusion is drawn that dynamic cooling process and the advective warming processes have canceled each other. At 700 mb and below, considerable warming at 1045 CST is noted when compared to the 0900 CST data. This can be accounted for in part by advection. The temperature gradient at 850 mb and to the south of Omaha at 0900 CST was such that any shift of wind to a more southerly direction would increase the warm advection. Its exact magnitude would be difficult to measure without a more complete knowledge of the changing wind field. However, it should be noted that at the 850-mb level the 0900C wind was from 330° at 17 kn and at 1045C the 850-mb wind was from 200° at 34 kn.

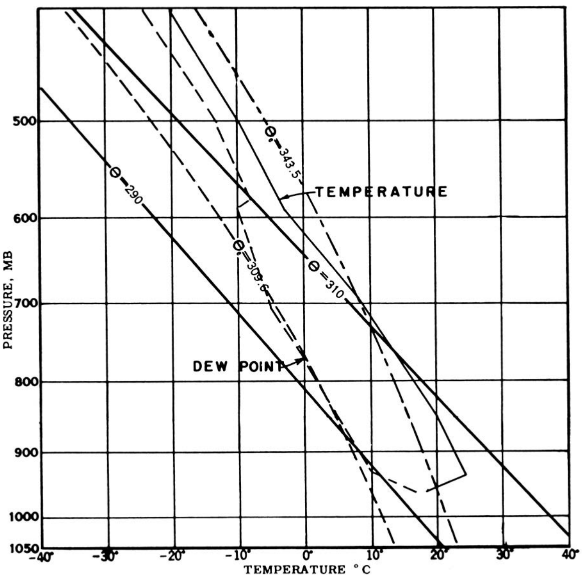
4. Conclusions

While it may be argued that errors in wind measurements as well as those inherent in any subjective analysis may result in computational errors being greater than the value being sought, the comparison between the predicted soundings and actual soundings shown here would seem to indicate that the correct sense of the changes may be detected. Therefore, it is concluded that given reasonable analyses, it is possible to estimate the distribution of vertical motion and this coupled



(a)

FIG. 6a. Omaha, Nebraska, upper air sounding for 0900C, 29 May 1956.



(b)

FIG. 6b. Omaha, Nebraska, upper air sounding for 1045C, 29 May 1956.

with a knowledge of the advective processes makes it possible to determine the direction of changes in air mass structure over selected surface points. This becomes particularly important in dealing with warm season severe storm producing synoptic situations that are not associated with

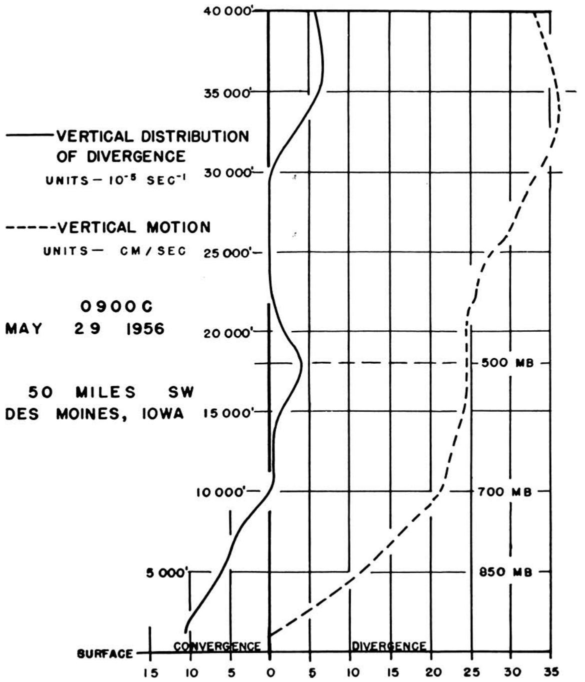


FIG. 7. Vertical distribution of divergence (solid curve) and vertical velocity (dashed curve) at 0900 C, 29 May 1956, 50-mi SW of Des Moines, Iowa. Units of divergence are in 10^{-5} sec^{-1} positive (divergent) to the right of zero ordinate and negative to the left. Vertical motion is in units cm sec^{-1} positive (upward) to right of zero ordinate and negative (downward) to the left.

strong bands of winds indicated as a requirement for the forecasting of such situations detailed by [4] and [5].

Both examples presented show clearly that cooling due to vertical motion can counteract advective warming processes. The second example presented demonstrates that of the two terms the second term on the right of (1) in most warm season situations may make the greatest contribution to divergence and vertical motion.

In these cases it would appear to have been more desirable to have made the computations for points where the soundings were taken rather than at nearby locations of severe storm occurrence. However, one objective of this study was to determine the rate at which air mass modification conceivably could have occurred at the time and site of occurrence rather than at the location of sounding stations removed some distance therefrom. The comparison was made only to show that the direction of the changes were correct.

In the second case, comparison of the predicted sounding for a point 50-mi SW of Des Moines

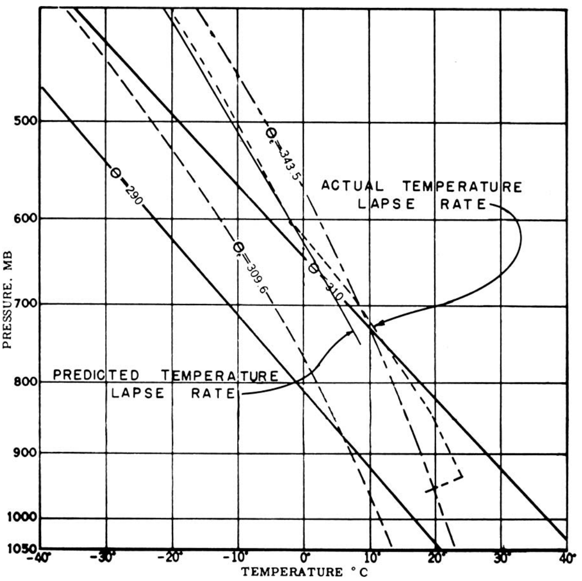


Fig. 8. Comparison of actual temperature lapse rate (dashed curve at 1045C, 29 May 1956 over Omaha, Nebraska, with temperature lapse rate (solid curve) predicted for a point 50-mi SW of Des Moines, Iowa, as a consequence of the vertical motion distribution indicated in fig. 7, acting for one hour and 45 min.

with the Omaha sounding is poor particularly in the lower layers. Therefore, one must conclude that important changes in vertical motion and/or horizontal advection with reversals in their direction may occur closely in time and space behind the instability line.

The primary purpose of this paper was to compare the modification in air mass structure due to vertical motion in the vicinity of severe storm occurrence under two contrasting synoptic situations. A comparison of the vertical motion profiles of figs. 2 and 7 reveals that in the first example the greatest modification due to vertical motion should occur in the 15- to 20-thousand ft layer while the second example indicates the greatest modification in the 30- to 40-thousand ft layer. A further inspection of these vertical motion profiles indicates subsidence is occurring in the layer above 30-thousand ft in the first example, a process that would inhibit the growth of thunderstorm cells beyond that level. The second example indicates positive vertical motion through the 40-thousand ft level which would permit the growth of thunderstorms to greater heights than in the first example. Observational evidence indicates that warm season thunderstorms grow to greater heights than do those in the colder seasons. The distribution of vertical motion with height shown here which is believed to be representative of the two seasons provides at least a partial explanation of why this should be the case. From this it may be concluded that the distribution of divergence with height can be used as an indirect measure of the heights to which thunderstorms can develop.

5. Further research

The problem of relating the strength of the dynamic structure to the time rate of change of stability needs to be studied further. It is known, for example, that the action of strong dynamic structures upon relatively stable air masses can result in severe storms. On the other hand, severe storms are known to occur in unstable air masses that are being acted upon by weaker dynamic structures. Additional research upon the changes in vertical motion over a point as the instability line passes is indicated and such may provide additional clues to the mechanics of the instability line.

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