STRUCTURE OF A COLD FRONT NEAR THE CENTER OF AN EXTRATROPICAL DEPRESSION

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ABSTRACT

Hourly aerological soundings released at Bedford, Mass. in April 1960, together with all available conventional synoptic observations, are used for a detailed analysis of a cold front south of the center of a northeastward-moving depression. It is shown that in this case the dynamically important change from one air mass to another throughout the entire troposphere occurred several hours after the passage of a weak surface cold front. The major cooling over Bedford presented itself in a nearly vertical column from about 500 m. above the ground to heights of around 8 km., with the main decrease of temperature at the surface occurring during the following two hours.

It is suggested that this case is not an exceptional one, but that it rather may reflect the typical structure of a strong cold air invasion near the center of a depression in which the horizontal wind component normal to the front increases with height. Under such conditions an analysis according to the classical Norwegian scheme, with the wedge-shaped cold air mass, would be unrealistic.

1. INTRODUCTION

In the descriptions of the fronts of extratropical depressions, much emphasis has been given to the wedge-shaped characteristic of an advancing cold air mass. This concept, introduced into synoptic-meteorological thinking with the Norwegian "Polar Front Theory" (Bjerknes [2] [3], Bergeron [1]) has been used as the standard in textbooks during the past 35 yr. Even the most recent manual of the National Weather Analysis Center [10] states: ". . . the primary synoptic tool is the Norwegian cyclone and frontal model, ..." In the following study we want to analyze a case in which there is no indication of a wedge-shaped cold air mass and to point out under what conditions this absence may be considered a typical feature of cold fronts in a tropospheric cyclonic stream field.

With the information available through war-time weather reconnaissance flights, and later through the radiosonde network in the middle and higher latitudes of the Northern Hemisphere, much research has been done on fronts. Only a selection of the extensive literature, relevant to the present study, will be mentioned here. Taljaard, Schmitt, and Van Loon [18] emphasized the need for a revision of concepts and standardization of analysis of weather maps, with regard to the idealized frontal models mentioned above, in the light of these new and better data. Schwerdtfeger [16], [17] showed that it is important to take into account the wind component perpendicular to the front. Where this component increases with height, the wedge form of the advancing cold air mass cannot persist. This frequently is true in the southwest sector and not far from the center of an extratropical depression. In such cases there appears a transition zone, which is 100 to 300 km. wide with strong instability and corresponding irregular vertical motions produced by a faster advance of the cold air aloft than of the cold air in the lower layers of the troposphere. On the other hand, where the wind component perpendicular to the front generally does not increase with height (that is, in the frontal region farther from the center of an extratropical depression where the cold air mass has spread out over a larger area and the horizontal thickness gradient is less pronounced), the wedge-shaped cold air body can persist for several days. Some supporting evidence for these notions can be found in a study of 50 cold fronts, by Sansom [15]. Miles [9], in an extensive study of cold fronts over southeastern England, arrives at two basic types of cold fronts. He pays particular note to the fact that in the majority of fronts studied the wedge shape was either absent, or, if present, was poorly defined. A few, however, did really have the characteristic wedge shape which is presented as the typical feature in most textbooks.

A series of studies of fronts was made at the University of Washington (Kreitzberg and Reed [6], [7], [8], Reed [13], [14]) employing the results of a detailed analysis of continuous radar echoes and short-interval soundings. These studies also indicate, in several cases, the existence of a progressing transition zone between warm and cold

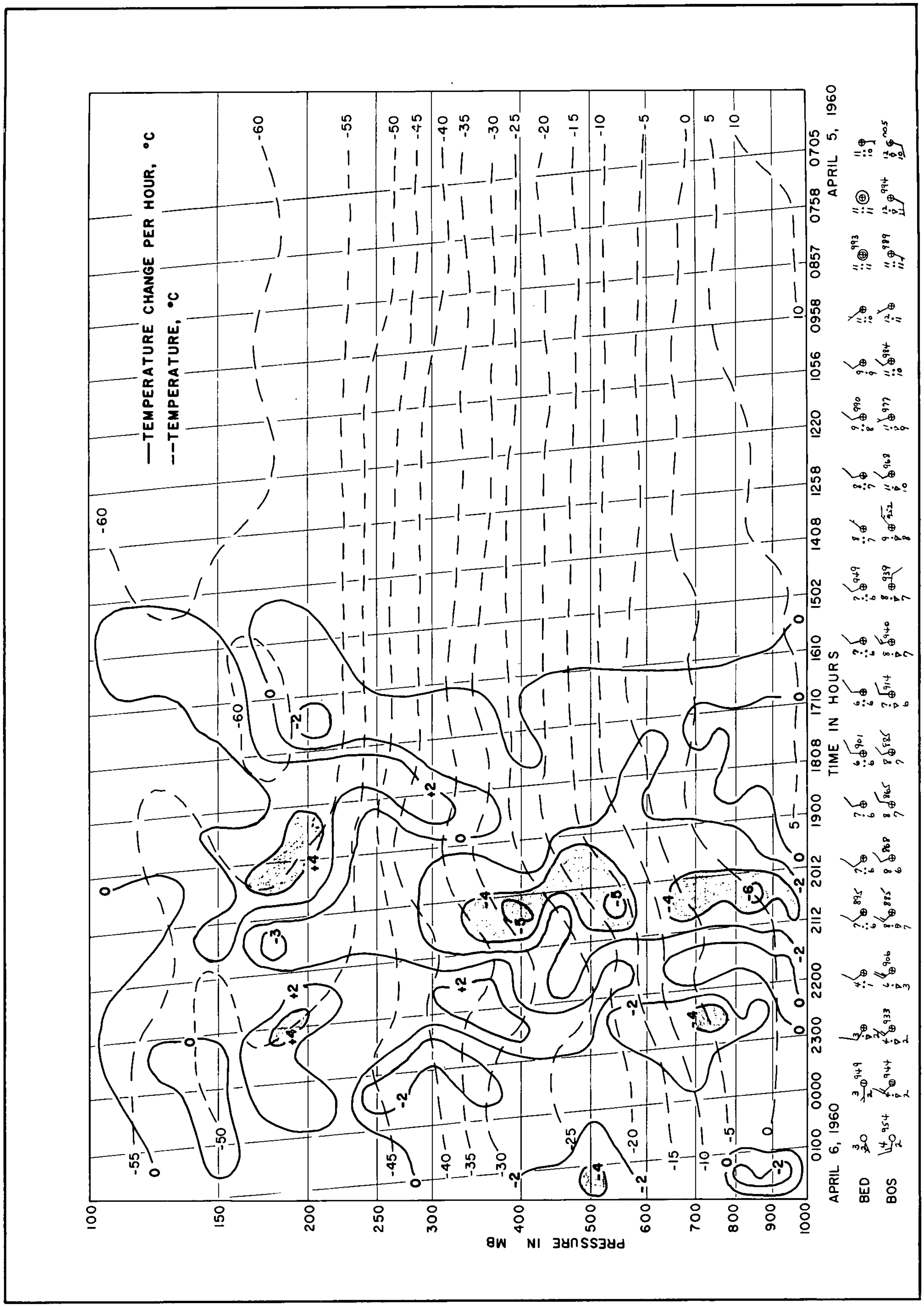


FIGURE 1.—Vertical cross section of hourly temperatures (dashed lines) and hourly temperature changes (solid lines) at Bedford. Below: Hourly surface weather observations made at Bedford (BED) and Boston (BOS).

air with the decrease of temperature occurring simultaneously through most of the troposphere rather than the advance of a wedge-shaped cold air mass in which the decrease of temperature over a given station occurs later at the higher levels.

Because of the need for detailed, short-interval data in conducting studies of the cold front's structure through the troposphere, we are obviously fortunate to have available, for the present study, the results of the series of hourly soundings released at Bedford, Mass., the first week of April 1960, published in extenso by Court and Salmela [5].

2. SOURCES OF DATA AND METHODS OF ANALYSIS

In addition to the rawinsondes from Bedford, results of regularly scheduled soundings for the eastern half of the North American continent were taken from the U.S. Weather Bureau's [19] published data tabulations. Copies of the surface weather observations (WBAN's), barograph charts, some thermograph charts, pibals, and rawinsondes for April 4–5–6 were obtained from the National Weather Records Center at Asheville, N.C., for 19 stations in the New England area.

These observations enabled us to prepare hourly surface maps and 12-hr. upper air maps (absolute topographies for the standard pressure levels and the tropopause, and relative topographies for the standard layers); only a small selection is included in this paper. Additionally, the hourly soundings were used for a vertical time cross-section of temperature and temperature changes per hour, over Bedford, presented in figure 1. For this graph, the values were plotted at 50-mb. intervals up to 200 mb. and at 25-mb. intervals between 200 and 100 mb., according to the exact times at which the pressure levels were reached by the sounding balloons.

It may be appropriate to state that all temperature values were accepted as published by Court and Salmela [5] with one exception: If there was, at an isolated level, a temperature change followed by a change of opposite sign and equal magnitude in the next 1-hr. interval, both these changes were disregarded. Without entering into discussion whether it may be a printing, computing, or instrumental error or even a real phenomenon, we may say that such variations are irrelevant for the question we are dealing with.

In the original publication (Court and Salmela [5]) the u and v components of the wind are given as displacements, in meters, for 1 to 4-min. intervals; u and v are the conventional notations of the W-E and S-N components of the wind. From these data, the 1-min. interval wind components were computed and reduced to meters per second. Since we want to call attention to the importance of the motion of air perpendicular to the front and its variation with height, this component was determined as $V_n=\sqrt{u^2+v^2}\,(\sin\alpha)$, where V_n is the component normal to the front and α is the angle between the wind vector and the surface front's orientation at the time of its passage at Bedford, Mass. (see fig. 2); this orientation is practically the same as that of the isotherms between 850 and 500 mb. (see figs. 3 through 6). A correction to the original wind data is indicated by a dot on the wind profiles in figures 7 and 8. Most errors could be traced directly to a large fluctuation in either wind speed or direction.

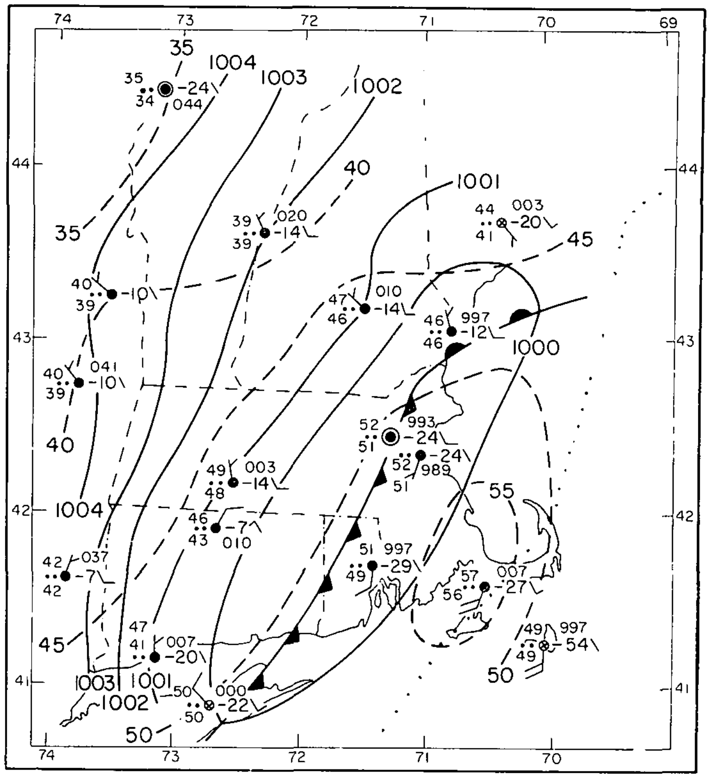
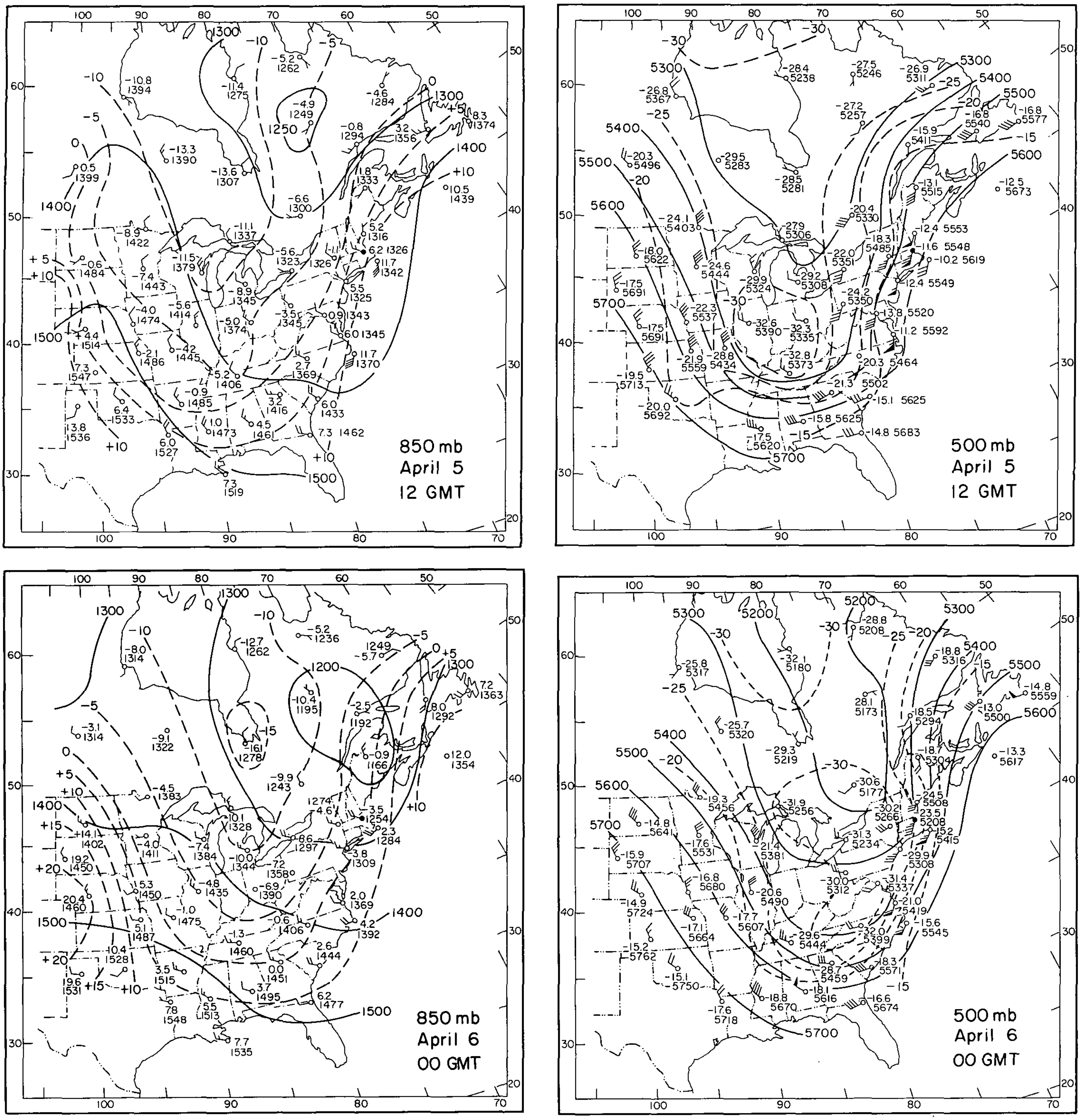


FIGURE 2.—Surface map April 5, 0900 GMT. Sea level isobars (mb.), surface temperature isotherms (°F.), and surface frontal system are shown according to conventional analysis. The dotted line indicates the approximate position of the surface front, 6 hr. later.

The final V_n values in figures 7 and 8 were smoothed vertically, taking $\overline{V}_{n2}=(V_{n1}+V_{n2}+V_{n3})/3$. In these graphs the speed of the front at the time of its passage is assumed to be the average speed as determined by surface charts for the 6-hr. period 0300 to 0900 GMT April 5, and is indicated by a solid line. The dashed line is the fastest observed motion of the surface front during the period of study. Some acceleration in the speed of the surface front as it moved out over the Atlantic can be noted for the 3-hr. period following its passage through Bedford.

An estimate of the accuracy of the Bedford wind data can be derived from a study by Rapp [12] who analyzed a series of special tests conducted with the AN/GMD–1, the type of receiver used at Bedford. Results of these tests for the u and v components of the wind indicate a probable instrumental error of less than 1 m./sec. at low elevations increasing irregularly to about 3.5 m./sec. at the 8-km. level. Rapp also states that larger errors



FIGURES 3–6.—Absolute topographies of the 850- and 500-mb. surface, for April 5, 1200 GMT and April 6, 0000 GMT. Continuous lines are contours in gpm.; dashed lines, isotherms in °C.

are to be expected under normal operating conditions in the field.

3. SYNOPTIC FEATURES

The storm system on which we shall concentrate our interest was one in a series moving from the west coast across the United States and out over the Atlantic. This particular system was located on the west coast on March 29, 1960. It was a well developed storm, centered near Wichita, Kans., at 0600 GMT on April 1, and one of the

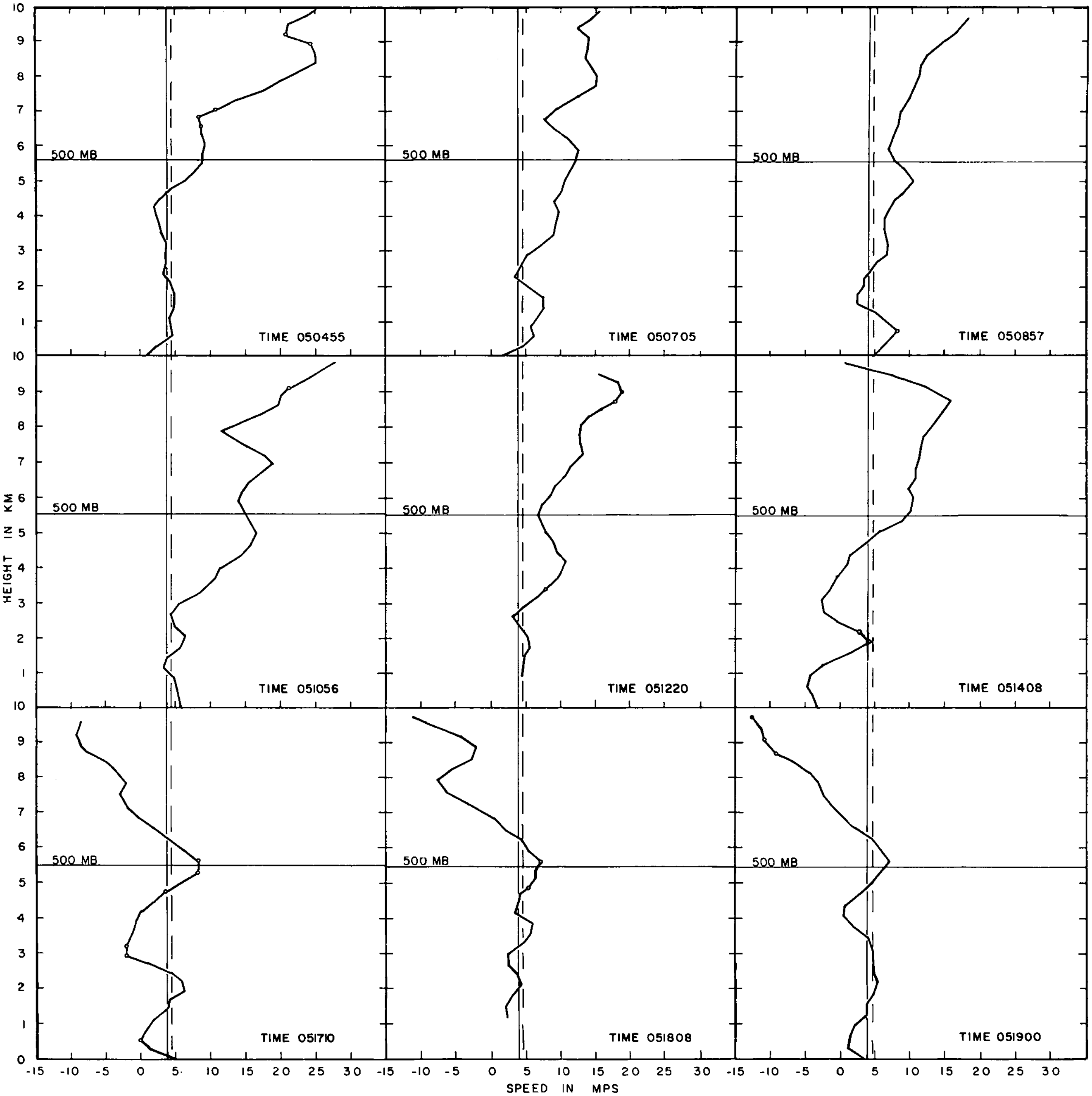


FIGURE 7.—Profiles of the wind component normal to the surface fronts and the mean isotherms, V_n. The straight vertical lines, solid and dashed lines (at about 5 m.p.s.) represent the average and the fastest speed of the surface front.

first storms photographed by TIROS I (Bristor and Ruzecki [4]). By 1800 GMT on April 4 we find the main air mass boundary was lying in a trough along the east coast with weak open waves and low pressure centers located over eastern Pennsylvania and northeastern Alabama. A distinct surface warm frontal passage is indicated at 1830 GMT on April 4 at Bedford. Bedford remained in the warm sector until about 0900 GMT on April 5, when a weak surface cold front passage appears to have occurred. Winds then veered to a northwesterly direction, but the temperature decreased only slowly and did so only in the lowest 2000 m. of the troposphere (see fig. 1). Looking at the sea level pressure field only (fig. 2), one could get the impression that this surface front

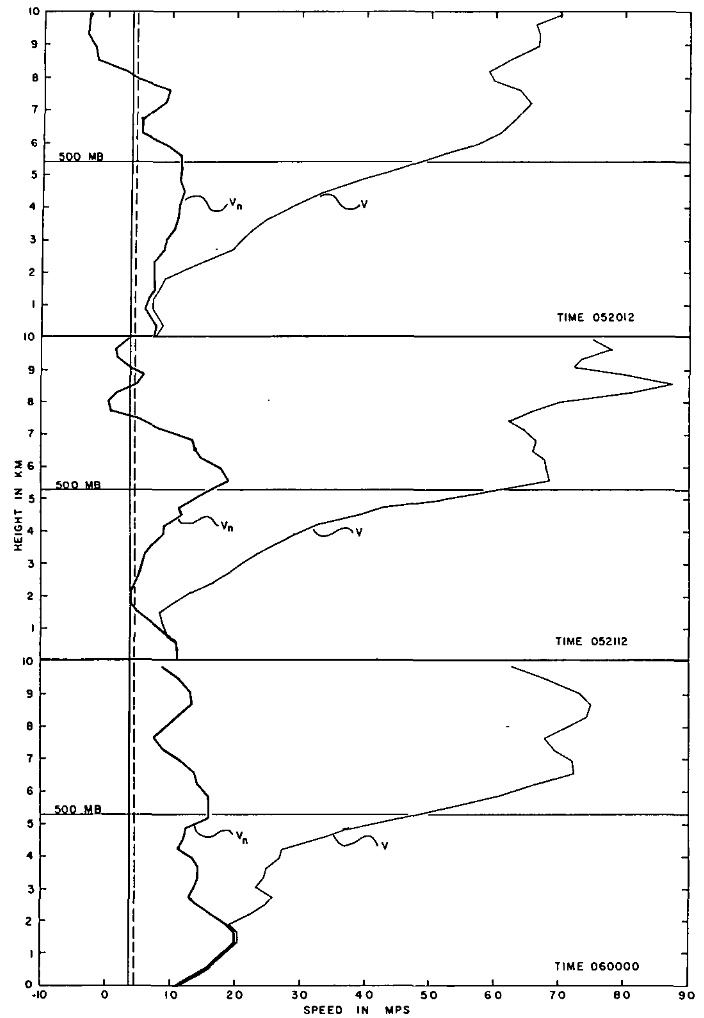


FIGURE 8.—Same as figure 7, for the soundings nearest in time to the passage of the "main front". V=total wind speed.

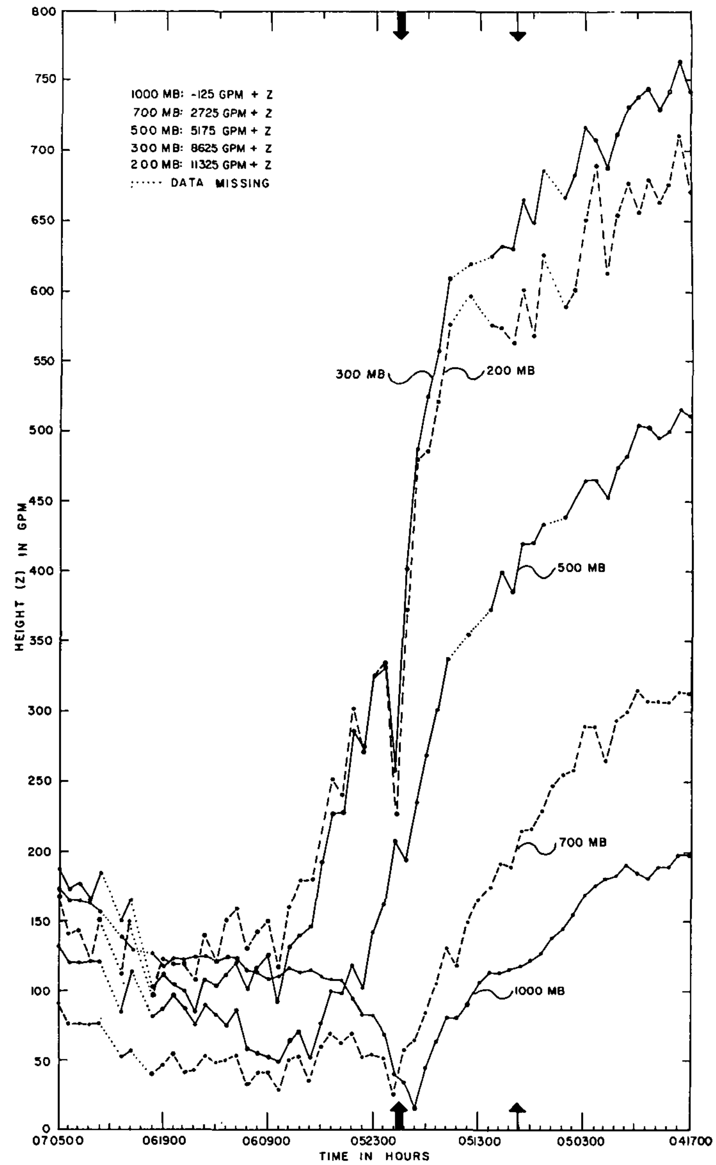


FIGURE 9.—Time-variation of the height of standard pressure surfaces over Bedford. Time of passage of the surface front and of the main front marked by small thin and heavy arrows, respectively.

represented the principal frontal zone. However, the "real thing", the dynamically important change from one air mass to another throughout the whole troposphere, occurred about 10 hr. later. This is clearly seen in figure 1 in which the heavy lines indicate the hourly rate of change of temperature over Bedford. On an average over the lower 8 km. of the troposphere, this temperature change amounted to 7°C. for the 2-hr. interval from 1900 to 2100 GMT. In the following discussion, this front will be referred to as the main front and the former as the surface front.

4. RESULTS AND THEIR INTERPRETATION

The analysis of the tropospheric temperature field in the boundary region of a cold air mass on the rear side of a depression over New England, April 5, 1960, gives no evidence for the existence of a sloping discontinuity surface. Instead we find that, as the whole system moved northeastward and the fronts passed over Bedford, the major cooling occurred in a nearly vertical column from about 500 m. above the ground to heights of around 8 km., and the major decrease of the surface temperature itself occurred during the following 2 hr. (see figs. 1, 9, and 10). All this happened about 10 hr. (equivalent to a distance of approximately 250 km.) after the sea level pressure field, the temperature near the surface, and the winds in the lowest 1500 m. indicated the passage of a weak cold front.

If only two daily aerological ascents and not the hourly soundings of Bedford had been available, it would have been easy, and of course quite unrealistic, to accept the familiar notion of the advancing cold air wedge, in this case with a relatively steep slope of about 1 : 40.

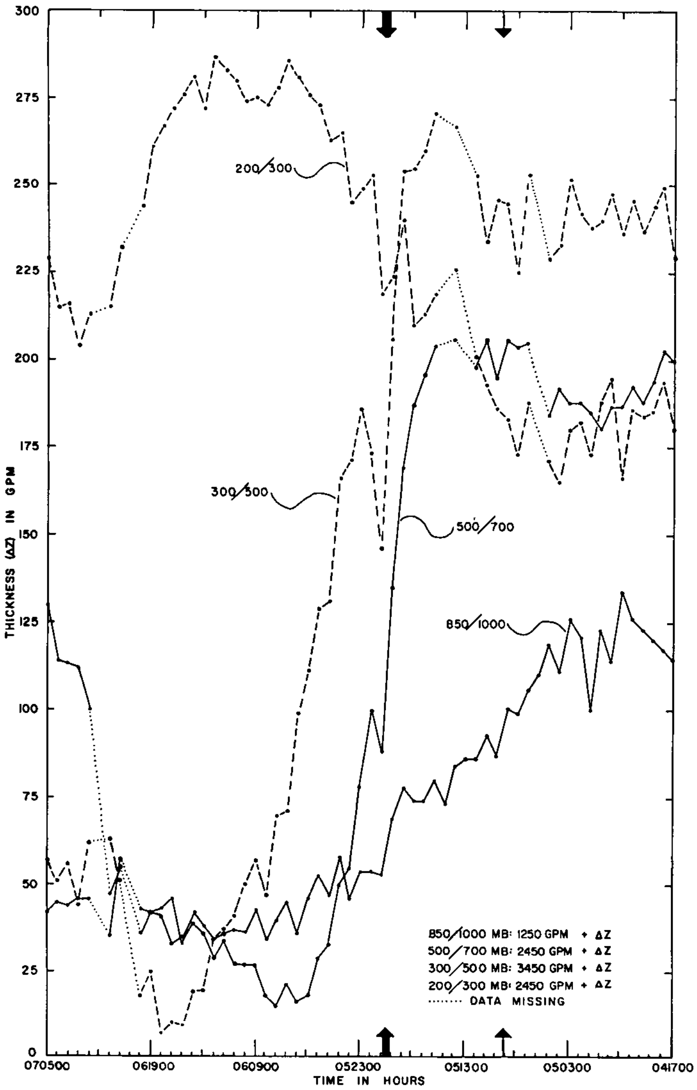


FIGURE 10.—Time-variation of thickness of standard layers over Bedford.

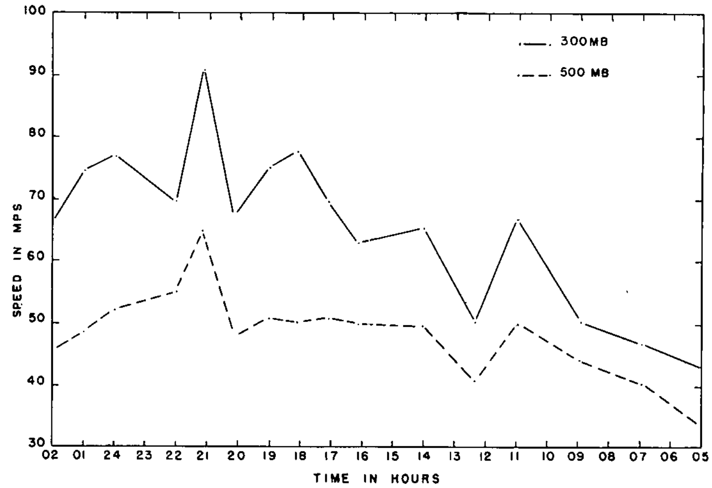


FIGURE 11.—Time-variation of wind speed at 500 and 300 mb. over Bedford, April 5, 0500 to April 6, 0200 GMT.

The wind components in the direction normal to the surface front and approximately normal to the isotherms in the lower half of the troposphere on the front side of the major cooling, show that the colder air aloft was moving at a speed in excess of the surface front movement (see figs. 7 and 8). In particular, this was so at the time of the advance of the main cold air mass over Bedford. In the soundings of 2012, 2112, and 2400 GMT (the 2200 and 2300 ascents are without winds for the lower half of the troposphere) the wind component normal to the surface front and to the lower troposphere isotherms exceeds the surface front speed at practically all levels up to about 8 km. In the crucial sounding at 2112 GMT, the increase of this component with height between 2 and 6 km. appears highly significant (see fig. 8).

It is obvious that the strong contrast between the warmer air ahead of the main front and the cold air behind it, with an almost vertically oriented transition zone through the troposphere, must be related to an increase (with height) of the slope of the isobaric surfaces; thus a pronounced horizontal pressure gradient must exist in the upper layers. A well developed jet stream is then to be found in the upper troposphere. This suggests that in such cases the approximately evenly spaced contour lines conventionally drawn on the upper air maps can not give an adequate picture of the true stream pattern, and that the wind measurements of stations at a distance of several hundred kilometers may significantly differ from those which a pilot would find when flying through such a region.

This is actually true in our case. Figure 11 shows the observed wind speed over Bedford at the 500 and 300-mb. levels for the period from April 5 0500 to April 6 0200 GMT. These winds were computed from the 2-min. interval displacements, corresponding to a layer of about 700 m. thickness, centered at the height of the 500- and 300-mb. levels. We find a maximum value of 91 m./sec. (177 kt.) at 300 mb., well above the geostrophic wind of 72 m./sec. (139 kt.) computed from the contour lines in the corresponding upper air maps, and far in excess of the maximum value reported, 41 m./sec. (80 kt.), by neighboring aerological stations at April 6 0000 GMT. Thus we may say that a realistic appraisal of the vertical structure of fronts is not a superfluous refinement of the generally useful concept of the wedge-shaped cold air mass; it rather has immediate practical implications and applications.

5. GENERAL CONCLUSIONS

Such a statement leads us to the question already mentioned in the introduction, whether we are here "confronted" with a unique or, at least, exceptional phenomenon, or with a general characteristic of cold fronts in cases of considerable vertical extent of the cold air mass, a characteristic which up to now has not found much attention in the American meteorological literature. This lack of attention may be partly because of the large distance in space and time between the regular aerological soundings, and perhaps partly because more unquestioning faith than critical research has been put into the dominating Norwegian scheme.

In the form which might be considered an oversimplification of complicated atmospheric processes, and which certainly can not give full justice to all the details of the particular case analyzed in the preceding paragraphs, this question can be answered by means of figure 12. We make the realistic assumption that the thickness field (thickness of the 1000 to 500-mb. layer, for instance) can be schematically represented by the T-lines and the height of the 1000-mb. surface by the h-lines. This means that we have a low pressure system in the northeast corner of the graph, and a pronounced "cold front," or transition zone, between warmer air in the southeast and colder air in the northwest, extending toward the southwest. Hence the H-lines represent the height of the upper pressure level (500 mb. in our example). The strong upper wind field in the transition zone, as we found it documented by the hourly soundings in figure 11, becomes evident. Under such circumstances, the slope of the upper isobaric surface (equivalent to the horizontal pressure gradient) at point B is less than at point A (fig. 12). Assuming now that the path of the air moving through the transition zone is long enough that the wind can approach geostrophic equilibrium, an air parcel moving northeastward from A will arrive in the B region with a speed which is greater than the geostrophic equilibrium speed. Consequently, as the Coriolis force is greater than the pressure gradient force, there will be a deviation from the direction of the geostrophic wind toward the right; this ageostrophic component (in the direction of the little arrow at B), must tend to increase the wind component perpendicular to the thickness lines. The fact that such ageostrophic components are really important, has been proven by many transosonde flights (Neiburger and Angell [11]). In a case like that illustrated by our schematic figure 12, this effect can increase with height as long as the wind in the transition zone increases with height. That can happen, practically, throughout most of the troposphere.

An ageostrophic wind component of the same type, but opposite sign, must be expected in the "entrance" region where the air moves slowly into the transition zone and becomes subjected to a much stronger horizontal pressure gradient (for instance: between $H+4$ and $H+5$ along $T+1$ in fig. 12). There, however, the resultant wind will tend to become parallel to the thickness lines, thus decreasing the cold air advection.

The pressure variations related to such ageostrophic wind components, negative in the eastern border region of the transition zone (in the prefrontal region) and positive behind it, have at least the right sign. The

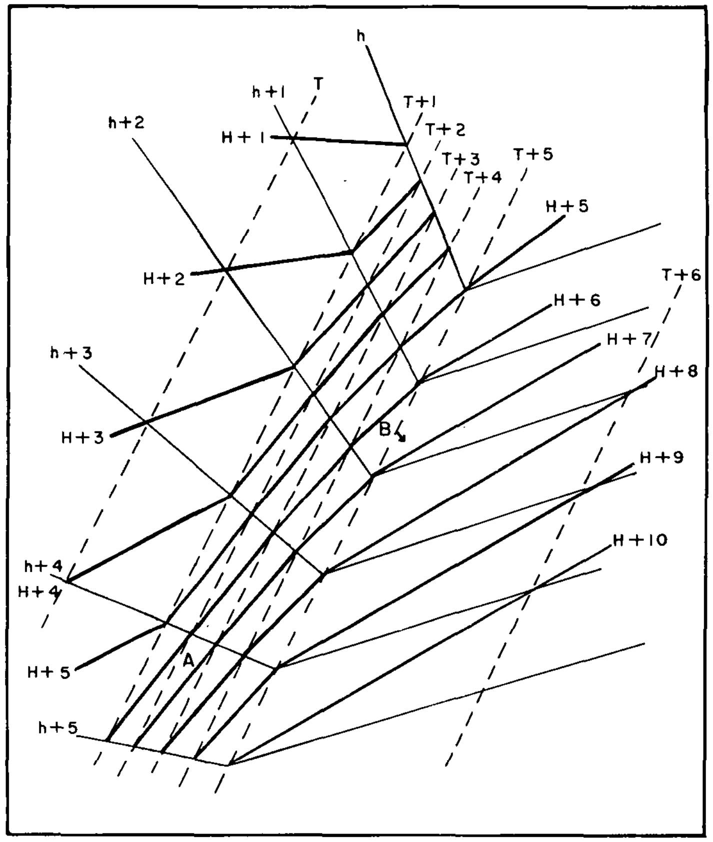


FIGURE 12.—Schematic illustration of the stream field in the region of a pronounced and high-reaching cold front. Thickness field given by T-lines, lower pressure surface (near sea level) by h-lines, and upper pressure surface by H-lines.

question whether also their magnitude is in agreement with the observed wind and temperature field, shall be examined in a later study. With this reservation, our considerations give us the possibility of making a general statement about the stream field in the region of cold fronts: *If and where the height and thickness field is such that there is cold air advection, an increase of the wind with height, and a decrease (downstream) of the slope of the isobaric surfaces, one must expect the wind component perpendicular to the mean isotherms to increase with height.*

Conditions of this kind are realized in many cold fronts of well developed extratropical depressions (see Kreitzberg and Reed [6]; Miles [9]; Reed [13], [14]; Sansom [15]). It may mainly be due to the large distance between sounding stations and/or the long time intervals between regular soundings that the features described in this study are not more frequently recognized. Nevertheless, they are important for realistic analysis and forecasting.

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[*Received February 18, 1964: revised June 22, 1964*]