SOME EFFECTS OF THE EVAPORATION OF WIDESPREAD PRECIPITATION ON THE PRODUCTION OF FRONTS AND ON CHANGES IN FRONTAL SLOPES AND MOTIONS

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INTRODUCTION

Surface weather maps in the United States have undergone a succession of dramatic changes in appearance during the past 15 years as one idea after another has dominated the minds of the analysts. The first major change came with the introduction of air mass analysis. After this type of analysis had been stressed for some five years, analysts not only accepted it wholeheartedly but in their enthusiasm, they reached the "spoked wheel" stage of frontal analysis. Out of every cyclone radiated a multitude of fronts, so many as to resemble the spokes of a wheel. Gradually, this first enthusiasm abated, spurious fronts were eliminated, and frontal analysis reached its stage of logical moderation in the United States.

But the stability of frontal analysis was soon upset by a new enthusiasm—this time for "instability lines," "squall-lines," or whatever the analyst chooses to call organized bands of bad weather ahead of the cold front in the warm sector. Since the concept of organized bands of thunderstorms and showers in the warm sector constitutes quite a departure from the ideas of the Norwegian school of air mass analysis, analysts were at first reluctant to accept it. Many forecasters still introduced surface fronts or cold fronts aloft to account for all organized lines of showers. But as time went on, the pendulum swung the other way; all lines of thunderstorms tended to be designated as instability lines, *even those which were fronts*. Unlike the fronts, which are inevitably tied to their air masses, these new instability lines are unrestrained; they move forward or backward, accelerate, decelerate, dissipate, re-form, cause tornadoes or other severe weather. In fact, as explanatory causes of bad weather they are extremely versatile. Meteorologists now appear to be in a stage of enthusiasm for instability lines similar to the "spoked wheel" stage of frontal analysis (fig. 1).

Especially with summer cold fronts it has become the custom with many analysts to draw an instability line at the forward edge of the transition zone and a cold front at the rear edge of this zone. The cold front is, then, coincident with the rear edge of the rain area, so placed because of the dew point discontinuity which necessarily exists between the rainless and the rain areas.

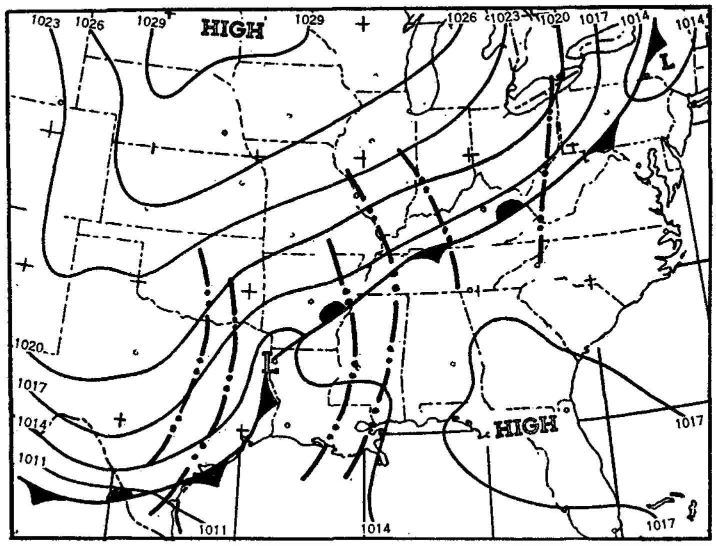

FIGURE 1.—Surface weather chart for 1830 GMT, May 13, 1953, showing what might be termed "over-enthusiasm" for instability lines.

This procedure is tantamount to a complete revision of the definition of fronts, since by definition a cold front lies at the *forward* edge of the transition zone, and such revision nullifies many of the forecasting rules based on the classical frontal theory.

What is needed now is a careful study of the relationship between warm season fronts and instability zones, in order to differentiate between the two. The purpose of this paper is to present a case study of a warm season frontal system, one which might easily be confused with an instability line, and to suggest how rainfall can so modify the temperature difference across a front as to cause an acceleration of its movement toward the warm air and to give it many of the characteristics now attributed to instability lines.

The extent to which rainfall can modify the temperature difference across a front increases with the dryness of the air mass into which the rain falls. In fact, the only limit on the amount of cooling which rain can produce is the wet-bulb temperature. In areas where persistent or heavy rain does occur, this limit is reached (see fig. 2). In the case to be described in detail, the difference between the dry-bulb and wet-bulb temperatures was greatest in Texas. It is here that the greatest temperature drop

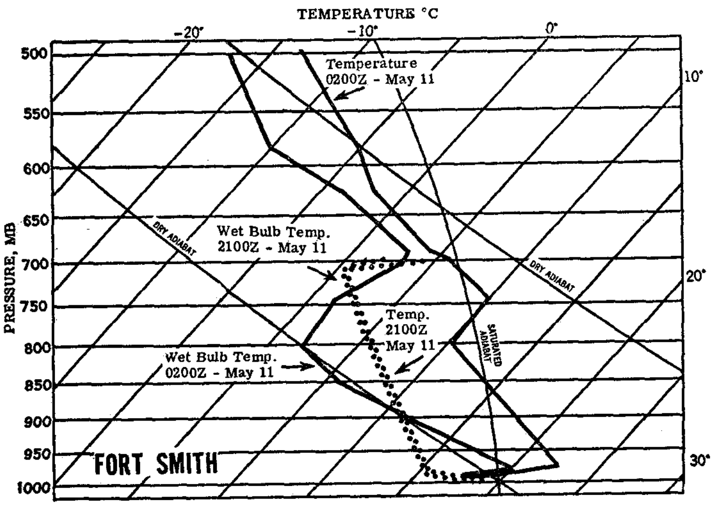

FIGURE 2.—Upper air sounding over Fort Smith, Ark., at 0200 GMT and 2100 GMT, May 11, 1953, plotted on a U. S. A. F. skew *T*, log *p* diagram.

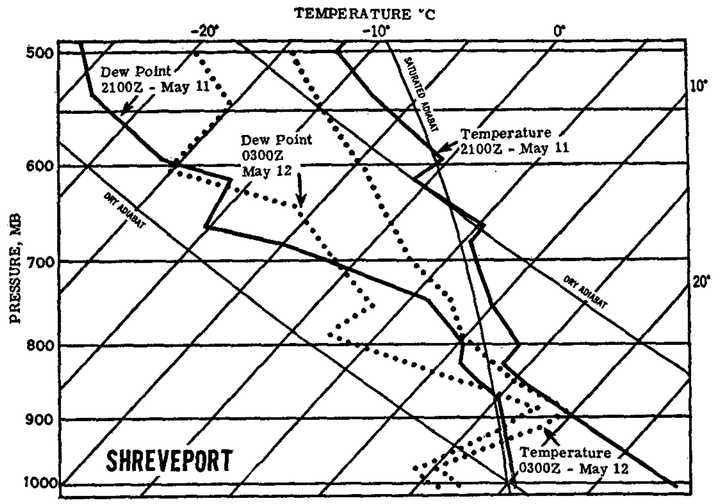

FIGURE 5.—Upper air sounding over Shreveport, La., at 2100 GMT, May 11 and 0300 GMT, May 12, 1953.

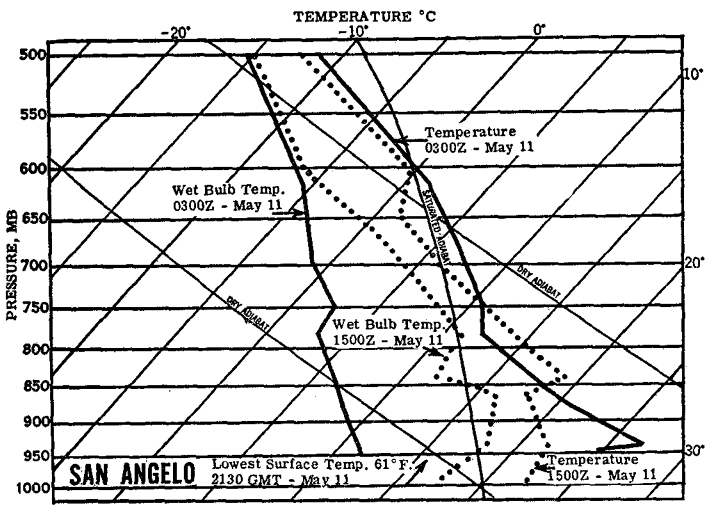

FIGURE 3.—Upper air sounding over San Angelo, Tex., at 0300 GMT and 1500 GMT, May 11, 1953.

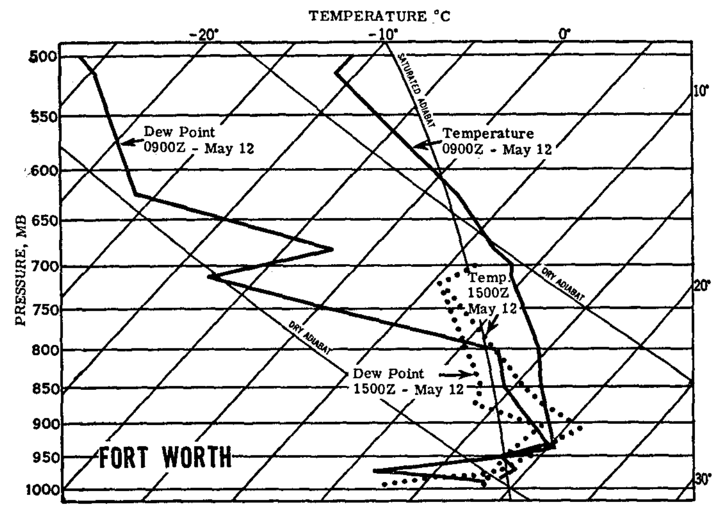

FIGURE 6.—Upper air sounding over Fort Worth, Tex., at 0900 GMT and 1500 GMT, May 12, 1953.

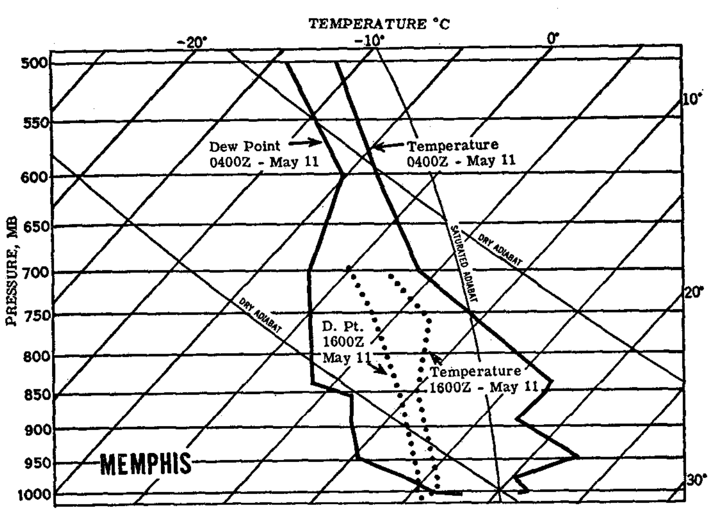

FIGURE 4.—Upper air sounding over Memphis, Tenn., at 0400 GMT and 1600 GMT, May 11, 1953.

occurred at the surface (fig. 3), accompanied by severe rain and hail storms and several tornadoes (including those which wrecked parts of Waco and San Angelo).

ANTECEDENT GENERAL CIRCULATION

Before presenting the specific case study, we shall describe the sequence of events which generally brings into juxtaposition a cool dry air mass and a warm moist one, requisite for producing the type of frontogenesis and frontal accelerations that is dealt with here.

To start with, a rather deep cold air mass enters the West Coast of the United States from the Pacific. As it crosses the western third of the country, the southern portion of this air mass is rather rapidly heated over the southwestern desert region and is further warmed by downslope motion east of the Continental Divide so that when it approaches the tropical maritime air moving northward from the Gulf of Mexico, the temperature contrast between these two air masses has become negligible. This means that the frontal slope will be very steep, approaching the vertical as the two air masses approach the same density. At this stage, there is little evidence of the front in the temperature field on the sea level, 850- and 700-mb. charts, although it is usually possible to find a moderate temperature gradient between the colder mP air and mT air at 500 mb. At 850 and 700 mb. the most obvious difference between these two air masses is the moisture content, which is usually very marked indeed. This moisture difference does account for a difference in the virtual temperature (and density) so that the slope to the front is slightly less than vertical. The front is generally oriented now from northeast to southwest and exhibits little motion, although it may start to move northward as a weak warm front.

HYPOTHESIS

Suppose that at this point, when the frontal motion is slight and the frontal slope is nearly vertical in the lower levels, the warm air overruns the front and heavy thunderstorms break out north of the front. Because of the excessive dryness of the "cold" air mass into which the precipitation falls, evaporation and consequent cooling of this air mass will be very rapid. Within an hour or two, the temperature difference across the front may increase greatly throughout a layer as deep as 10,000 feet at times.

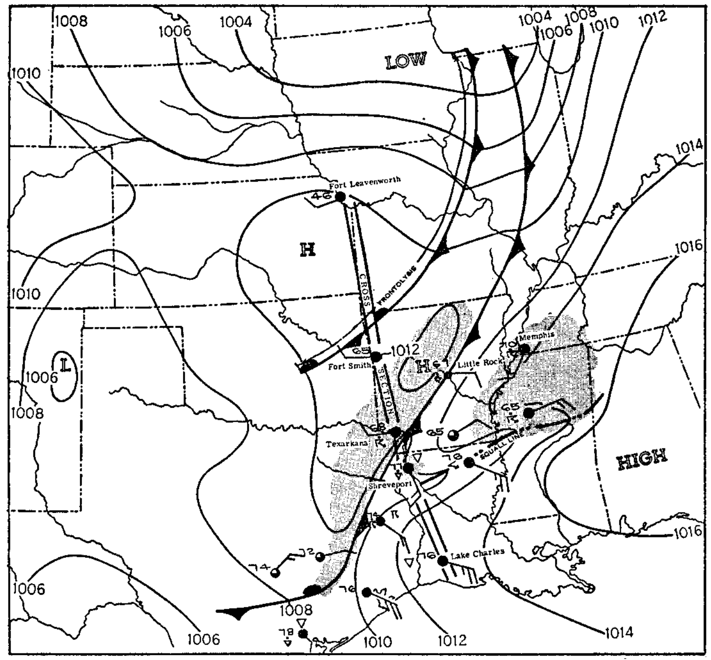

Figure 7.—Surface weather chart for 0330 GMT, May 11, 1953. Shading indicates the areas of active precipitation. Figure 8 is a corresponding cross-section.

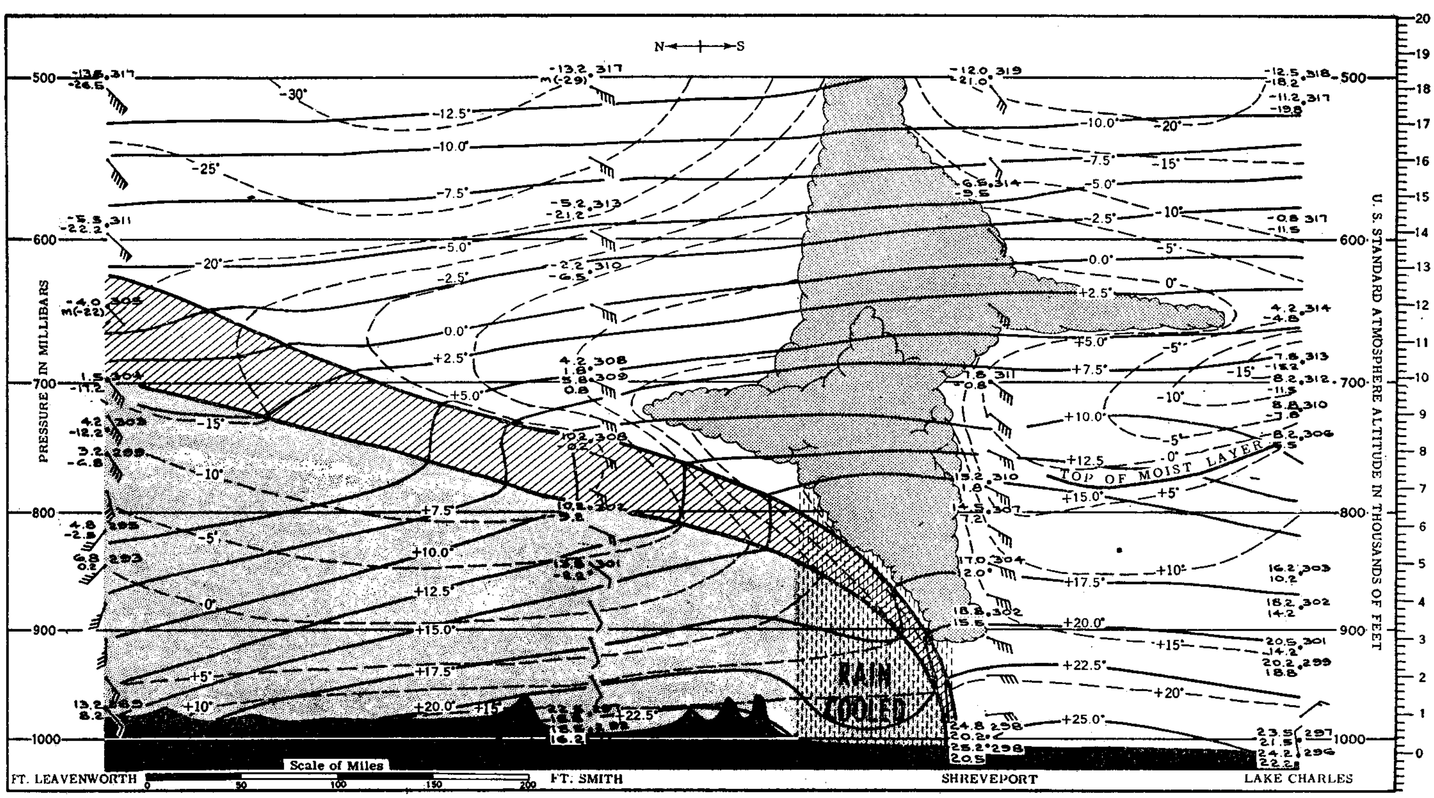

Figure 8.—Cross-section at 0300 GMT, May 11. Isotherms in °C. are the solid lines and isodrosotherms are shown as dashed lines. The plotted data also include potential temperatures (not analyzed). The heavy solid line indicates the transition zone of the front. Winds are plotted with reference to the directions (along the top of the diagram) which show also the orientation of the stations.

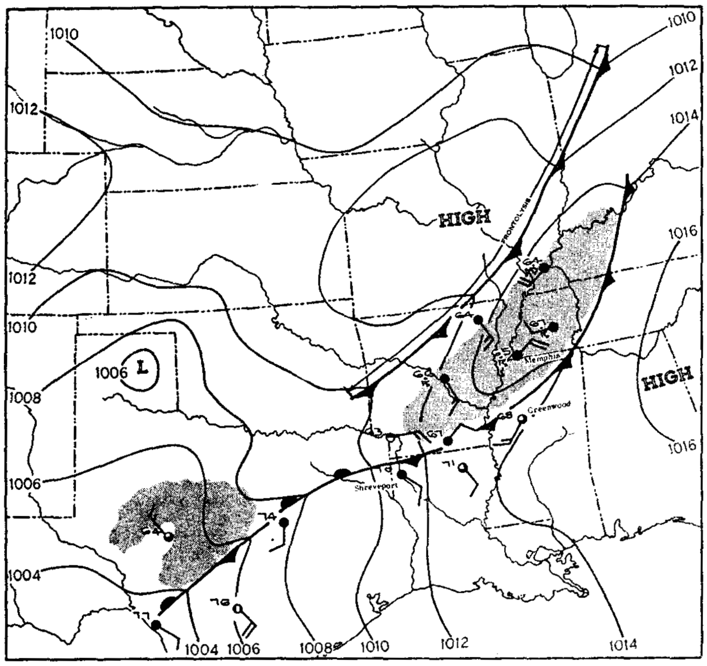

FIGURE 9.—Surface weather chart for 1230 GMT, May 11, 1953.

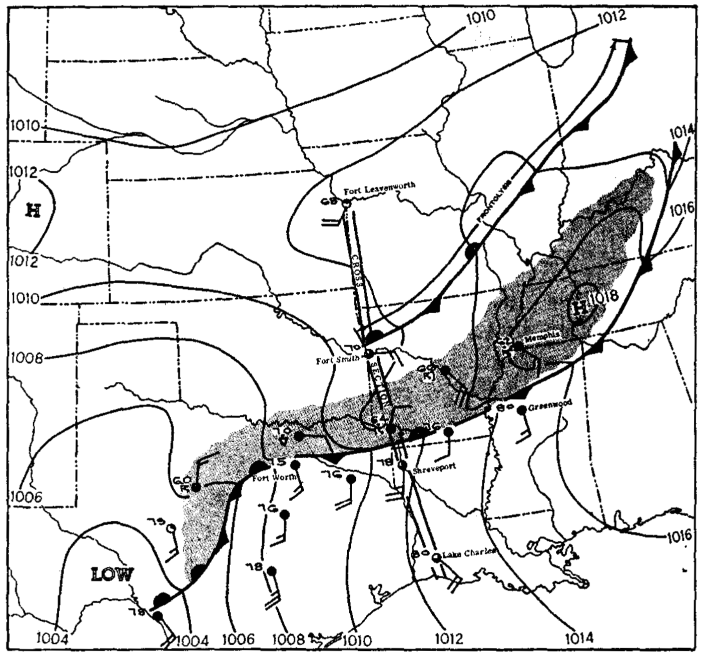

FIGURE 10.—Surface weather chart for 1530 GMT, May 11, 1953. Figure 11 is a corresponding cross-section.

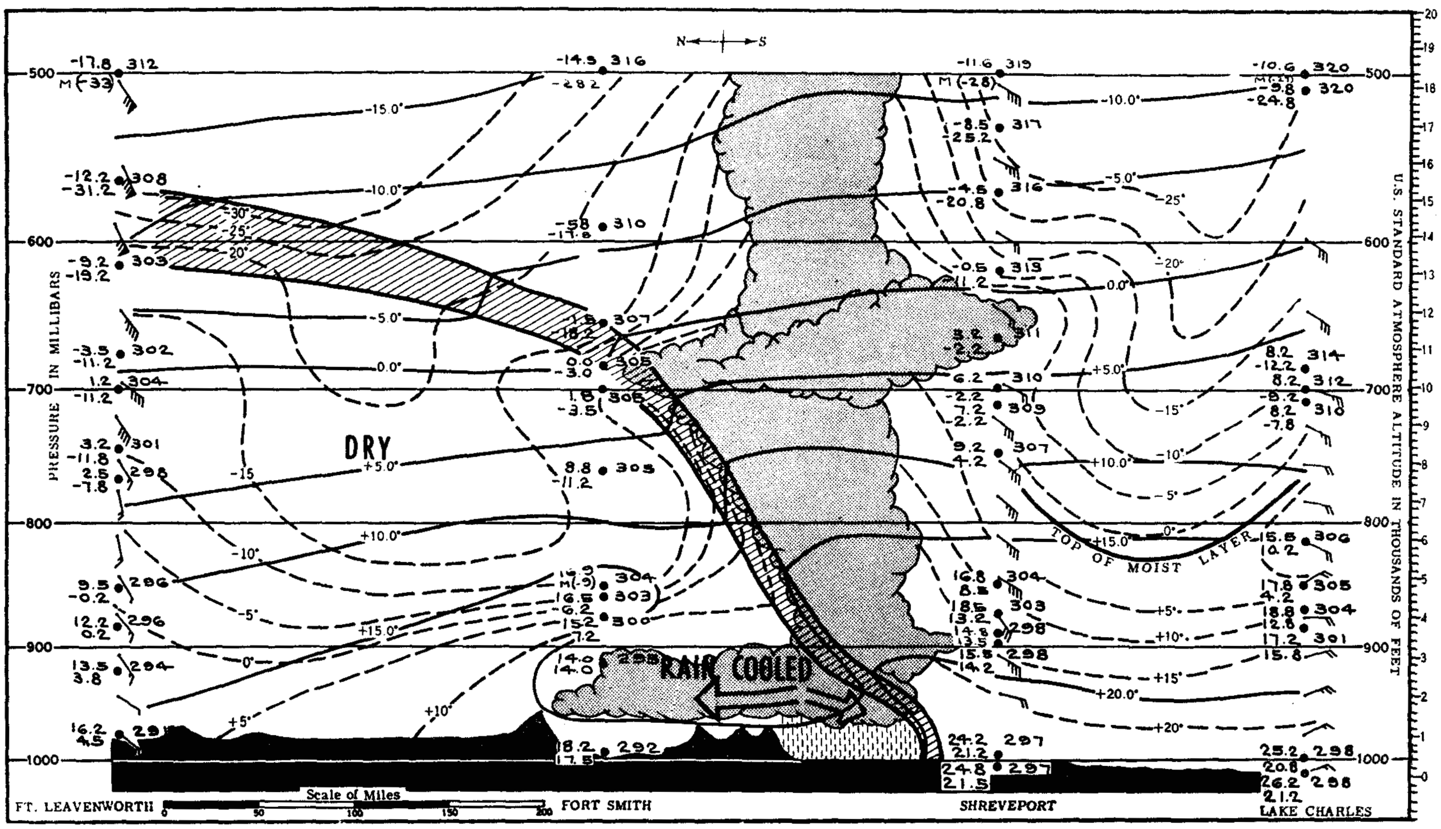

FIGURE 11.—Cross-section at 1500 GMT, May 11, 1953.

This obviously changes the pressure distribution [1, 2]; there is now a ridge forming north of the front in the rain-cooled air. With this rain-induced High and with the pronounced temperature gradient, it is hypothesized that the steep slope of the front can no longer be maintained and the cold air mass must flatten out with the front re-adjusting its slope to the new temperature difference between the air masses (see the following article by Wexler [3]). If it is assumed that at the top of the evaporation-cooled layer there is no appreciable change in temperature, the readjustment of the frontal slope requires that the surface front suddenly move from the rain-induced high-pressure area as the cold air mass spreads out. The effect of evaporational cooling on the frontal motion is most dramatic when a front which has been moving northward as a weak warm front suddenly reverses its motion and surges southward as a strong cold front.

Such a surge of the cold front may prove to be the trigger which frequently sets off pressure jumps. It has been suggested that pressure jumps can be produced by the sudden acceleration of part of a cold front [4, 5]. It seems that just such frontal accelerations may be easily produced by the process of evaporational cooling outlined above.

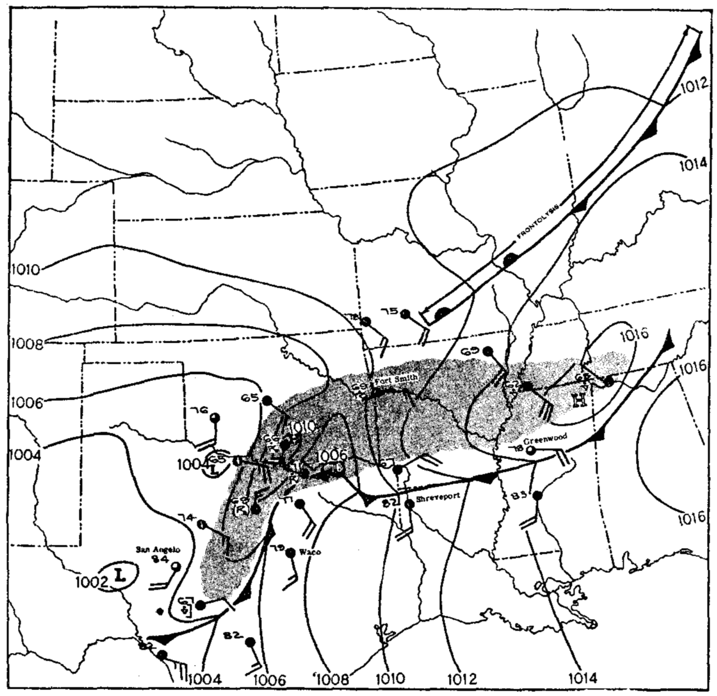

Figure 12.—Surface weather chart for 1830 GMT, May 11, 1953.

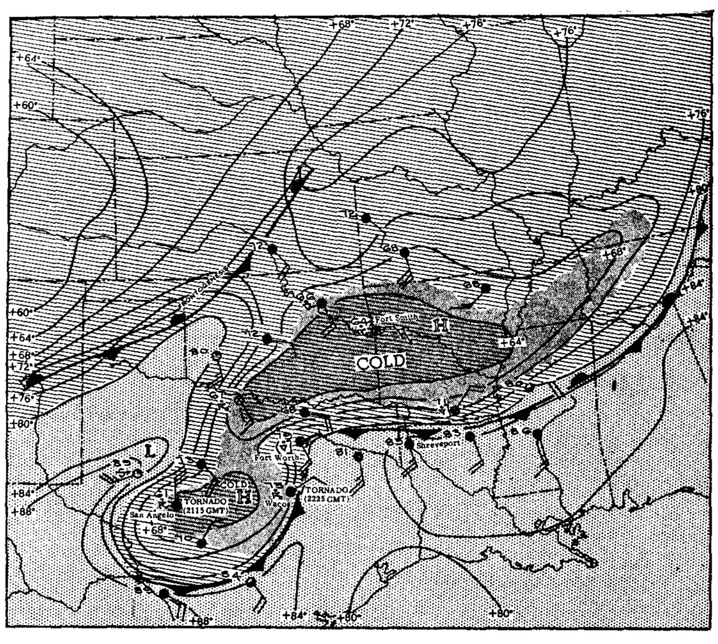

Figure 13.—Surface weather chart for 2130 GMT, May 11. Surface isotherms are for every 4° F. Note the occurrence of tornadoes at San Angelo and Waco, Tex.

THE CASE OF MAY 9–13, 1953

We turn now to the detailed example, which deals with a storm affecting the midwestern part of the United States from May 9–13, 1953, producing numerous tornadoes, widespread heavy rains, and rain-cooled air masses, some of which intensified existing fronts and some of which produced frontogenesis. A series of hourly maps (of which every third map is reproduced) was prepared to study the details of the movements of the fronts and the precipitation areas. Also prepared was a group of soundings at selected stations (figs. 2–6), showing the amount of cooling produced by the rain.

The first surface chart (fig. 7), synoptic with the cross-section of figure 8 and with the sounding of figure 2, shows a front, quasi-stationary in the south, oriented from northeast to southwest, and extending from Illinois to Texas. This front separates dry modified Pacific air to the northwest from an mT air mass to the southeast. An important feature of this situation was the structure of the mT air mass. It was much more unstable than usual, i. e., above the lower moist layer, there was aloft unusually dry air with a steep lapse rate. (The Showalter stability index [6] was −4.) This meant that the instability of the mT air could be released by a small amount of lifting. Some of this instability had already been released in the warm sector along the instability line through northern Louisiana and central Alabama (fig. 7). Frontogenesis was occurring along the instability line at this time due to convergence between the air flowing outward from the rain-cooled portion of the warm sector and the unmodified warm air to the south. Had the rain continued north of the zone of frontogenesis, an important front might have formed. The rain, however, died out during the night, ending the frontogenesis.

Behind the major cold front, on the other hand, the rain persisted with rapid cooling due to evaporation in the dry air. By 1230 GMT, May 11 (fig. 9) the rain-induced High was already distinctly formed in western Tennessee

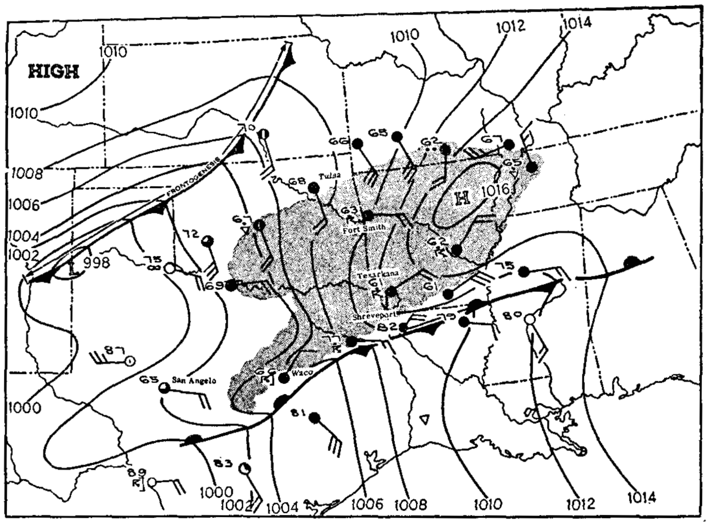

FIGURE 14.—Surface weather chart for 0030 GMT, May 12, 1953.

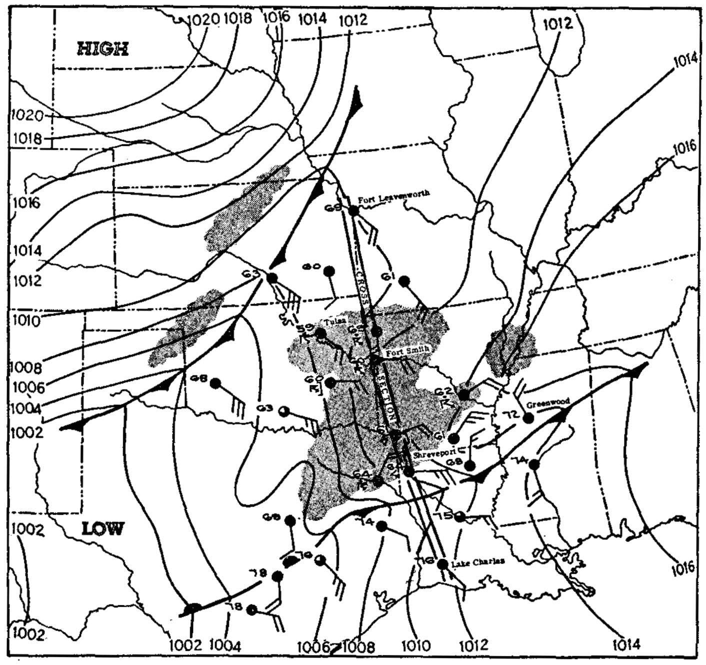

FIGURE 15.—Surface weather chart for 0330 GMT, May 12, 1953. Figure 16 is a corresponding cross-section.

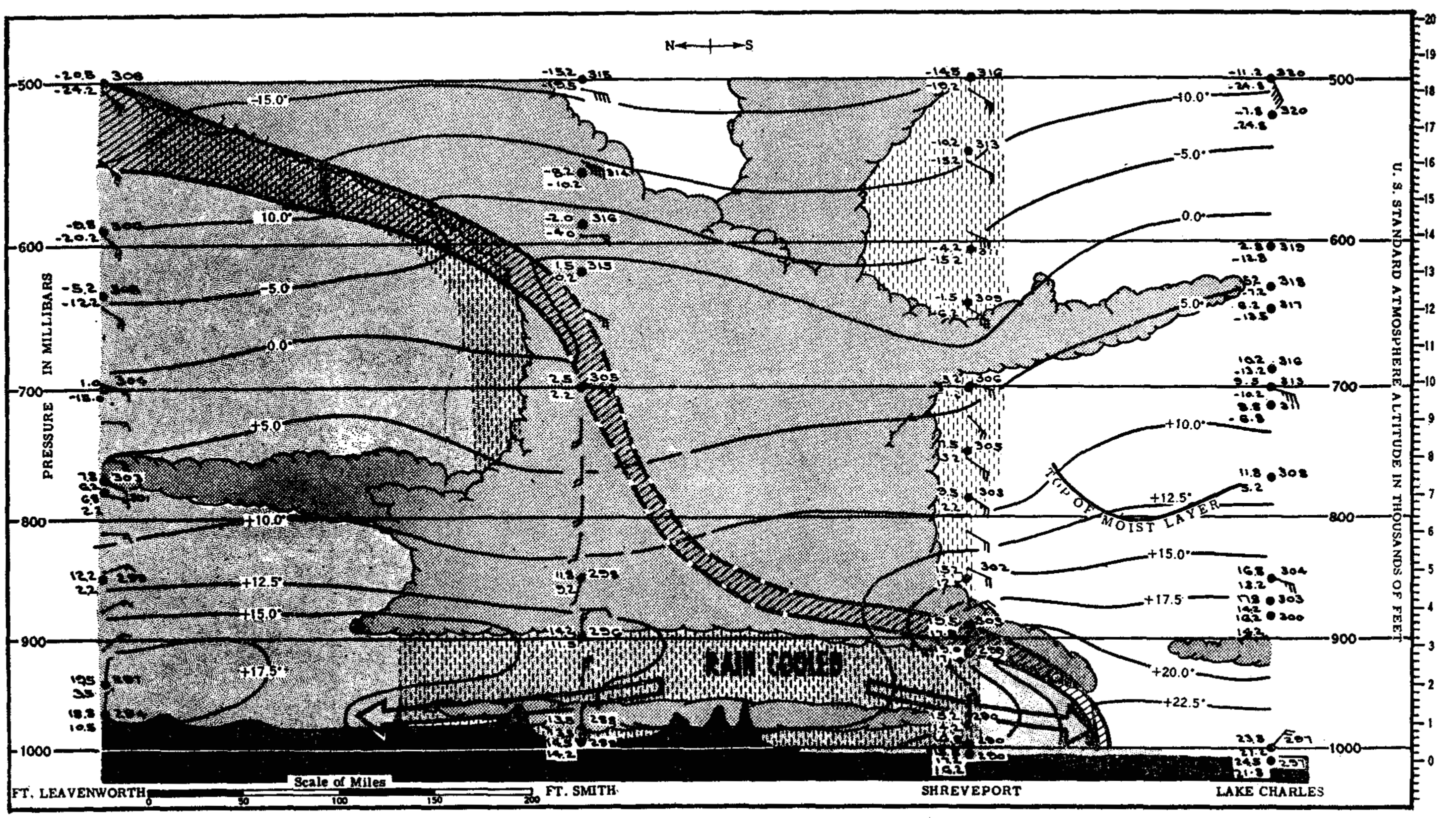

FIGURE 16.—Cross-section at 0300 GMT, May 12, 1953.

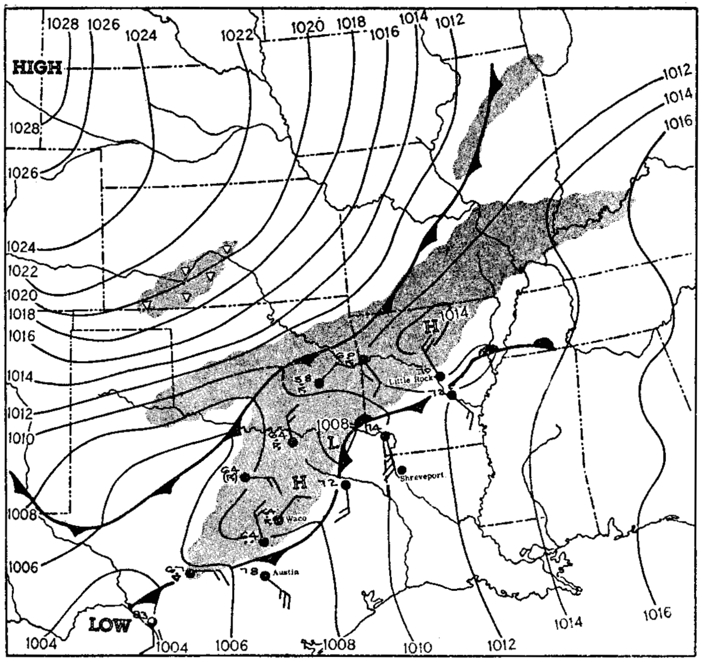

FIGURE 17.—Surface weather chart for 1530 GMT, May 12, 1953.

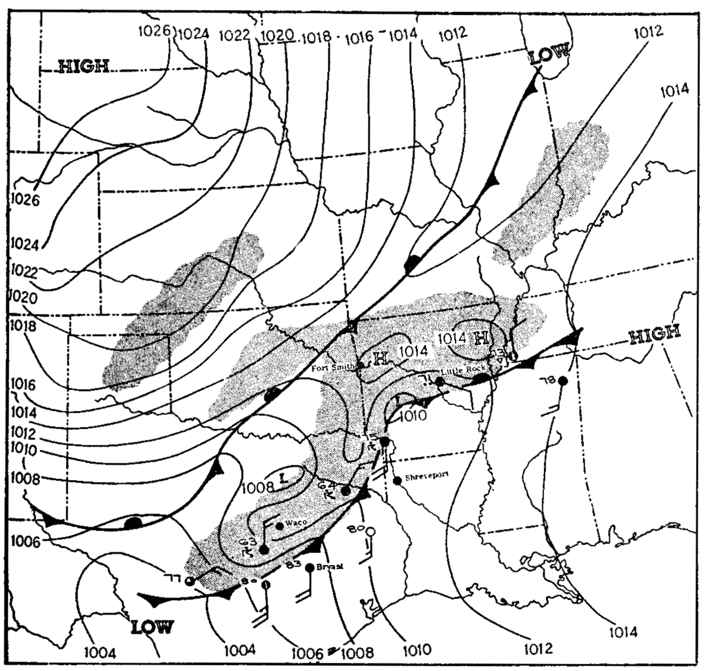

FIGURE 18.—Surface weather chart for 1830 GMT, May 12, 1953.

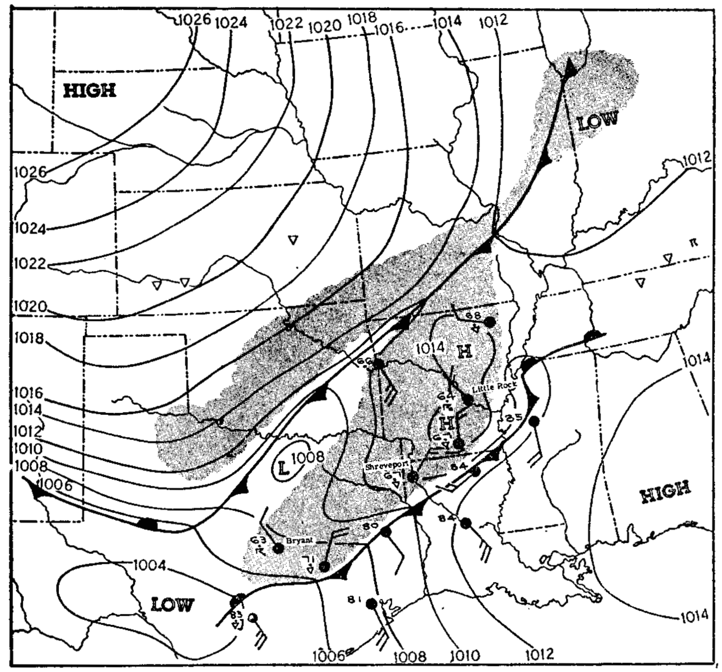

FIGURE 19.—Surface weather chart for 2130 GMT, May 12, 1953.

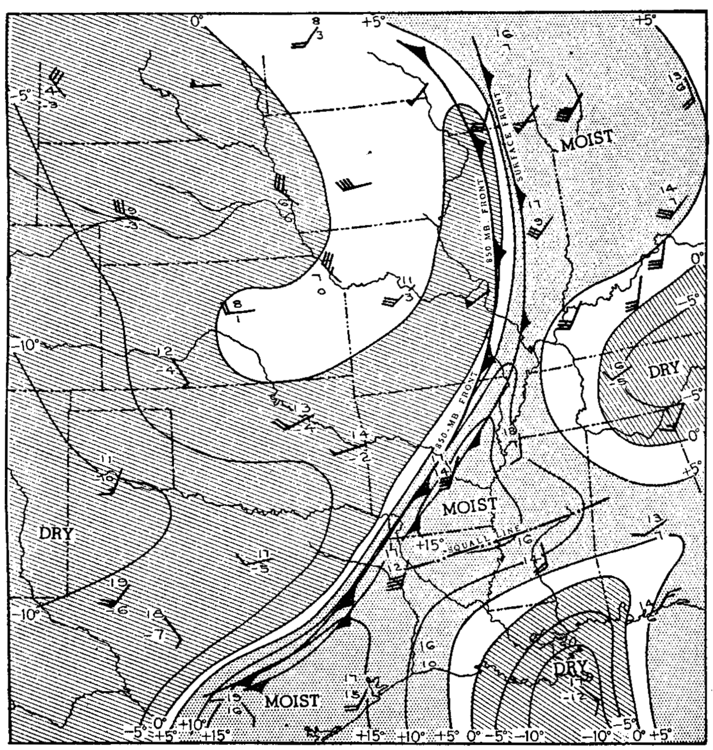

FIGURE 20.—850-mb. chart for 0300 GMT, May 11, 1953, showing moisture distribution. Isodrosotherms are for every 5° C. The data show the winds and temperatures as well as dew points.

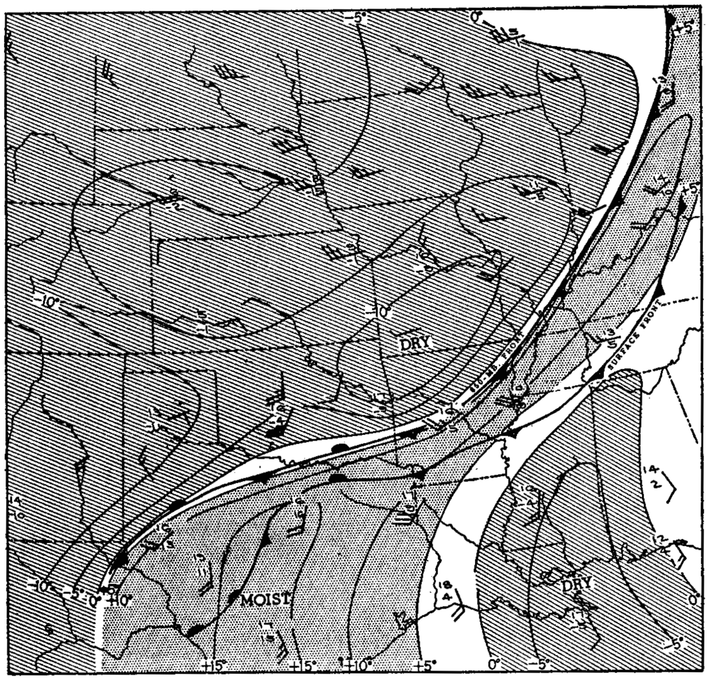

Figure 21.—850-mb. chart for 1500 gmt, May 11, 1953.

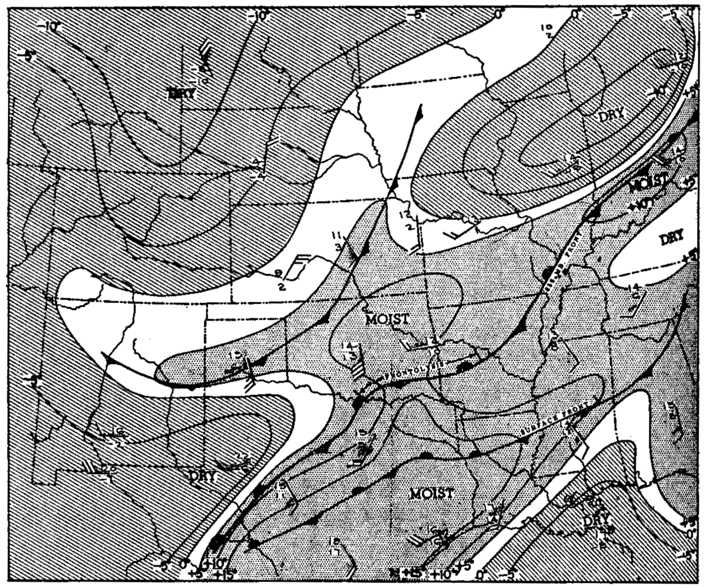

Figure 22.—850-mb. chart for 0300 gmt, May 12, 1953.

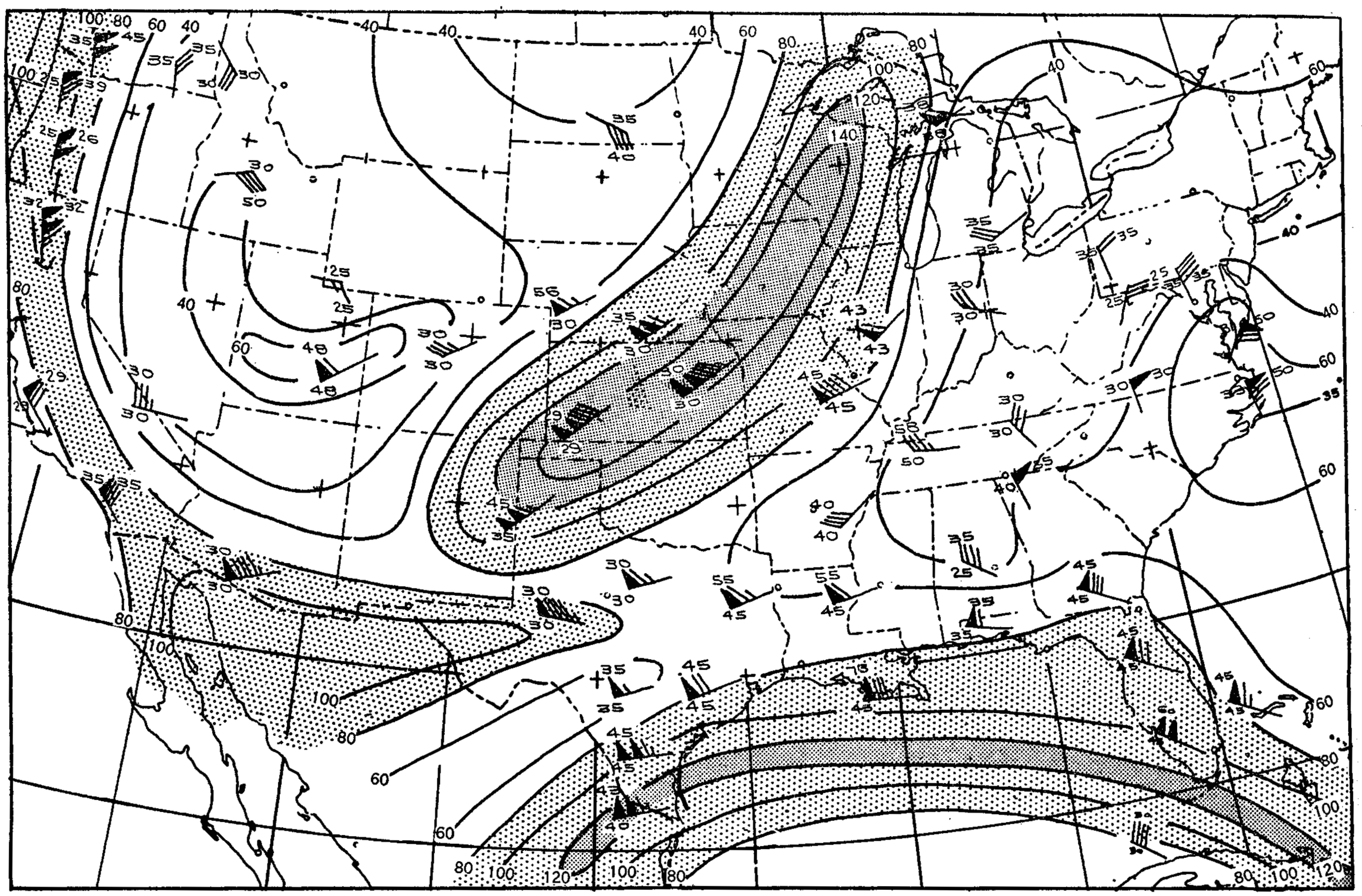

Figure 23.—Maximum-isotach chart for 0300 gmt, May 11, 1953. A full barb means 10 knots and a flag means 50 knots. The small number at the tip of the shaft indicates the level of the maximum wind in thousands of feet. The number at the tips of the flags and barbs is the highest level which the sounding reached. Isotachs are for every 20 knots.

and the surface winds were blowing out from it in all directions. During the next few hours the thunderstorms continued in this general area, merging with those in Texas (fig. 10). The evaporational cooling increased the temperature gradient across the front (fig. 10) and resulted in a flattening of the frontal slope, accomplished in this case by a southward motion of the surface front (fig. 11). On the ground the front was driven southward by a narrow band of northeasterly winds (see figs. 9 and 10), passing Greenwood, Miss., at 1730 GMT and continuing slowly to the south, despite the fact that during this entire period the general wind flow from about 3000 feet to the stratosphere was opposed to this motion (i. e., from the southwest).

A second southward surge of the rain-cooled air mass occurred in southwestern Arkansas (figs. 12–15). The soundings for Fort Smith, Ark. (fig. 2), best illustrate the cooling due to the rain which was instrumental in producing the change in frontal slope and this second southward surge of the front. The first sounding (0200 GMT, May 11), taken before the rain began at Fort Smith, shows the extreme dryness of the lower portion of the cold air mass north of the front. At this time the overrunning moist air was just reaching Fort Smith at about 10,000 feet. The sounding 19 hours later was taken when a thunderstorm was in progress at the station. Again, referring to the sea level charts near Fort Smith for this period, as the rain area increased in size and intensity in southwestern Arkansas, the sea level pressures rose as widespread cooling occurred (note the sea level isotherms in fig. 13 and the isotherms in low levels in fig. 16), surface winds became northeasterly, and the front which had been nearly stationary between Texarkana, and Shreveport, accelerated rapidly to the south (see figs. 11–13). It was during this period that devastating tornadoes struck Waco and San Angelo, Tex.

A third major surge of the cold front took place the following afternoon, again in southern Arkansas and northern Louisiana. By May 12, 1530 GMT (fig. 17) widespread severe thunderstorms again occurred north of the front. Again the rain, falling through the initially dry air mass, evaporated; the air mass was cooled, the frontal slope decreased, and the activated cold front surged to the south. As the cold air spread out, it lifted the overrunning warm air and set off additional thunderstorms. This is illustrated by the 1530—2130 GMT charts for May 12, 1953 (figs. 17–19).

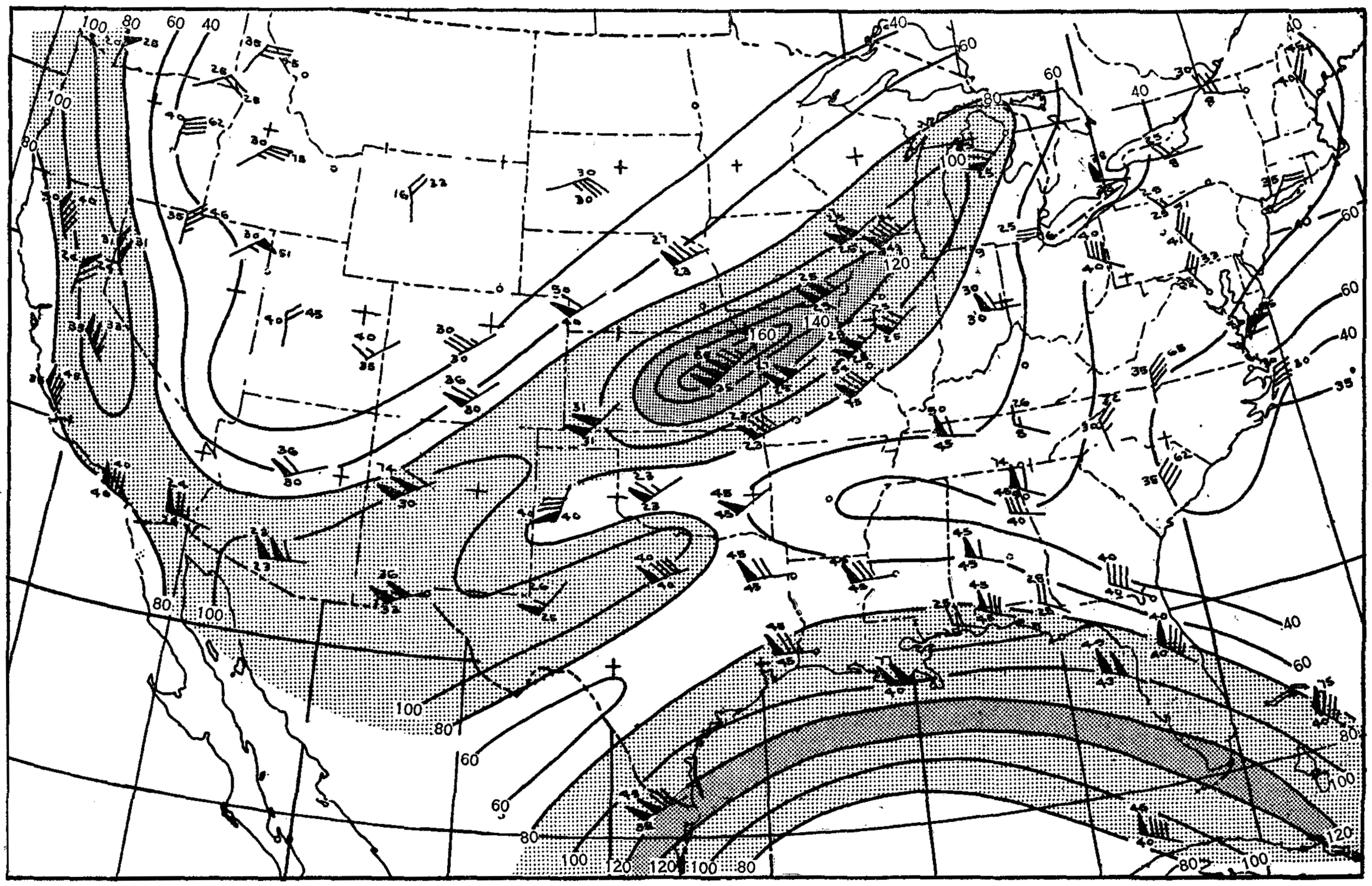

FIGURE 24.—Maximum-isotach chart for 1500 GMT, May 11, 1953.

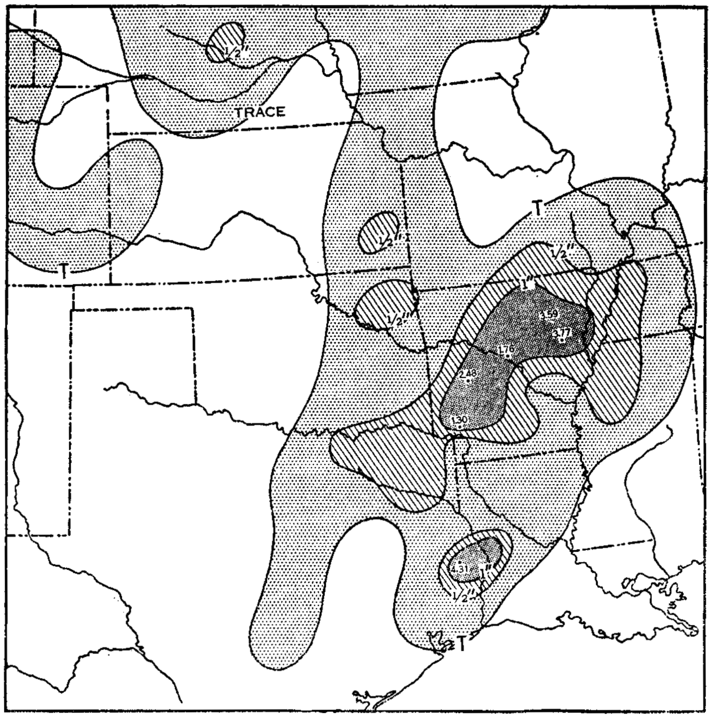

FIGURE 25.—24-hour precipitation chart ending at 1230 GMT, May 11, 1953. Some of the amounts in excess of one inch are shown.

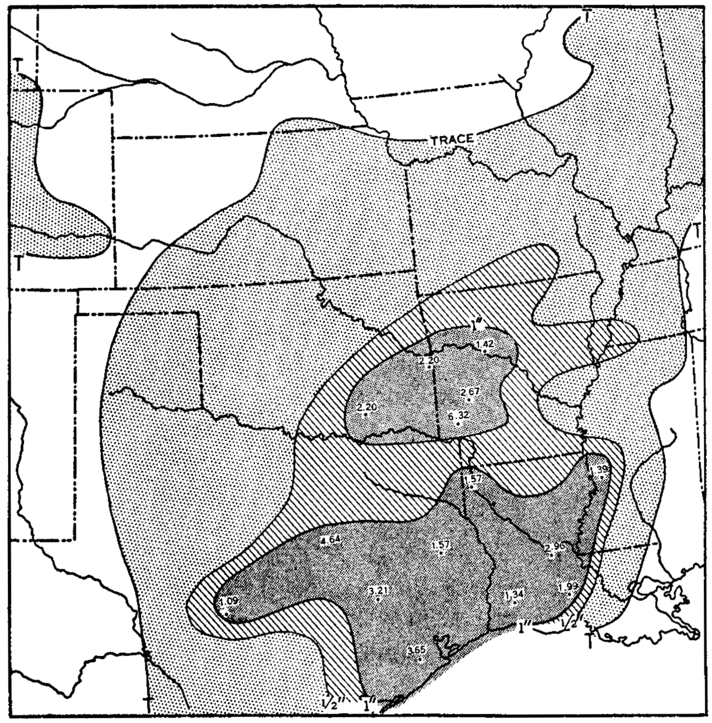

FIGURE 27.—24-hour precipitation chart ending at 1230 GMT, May 13, 1953.

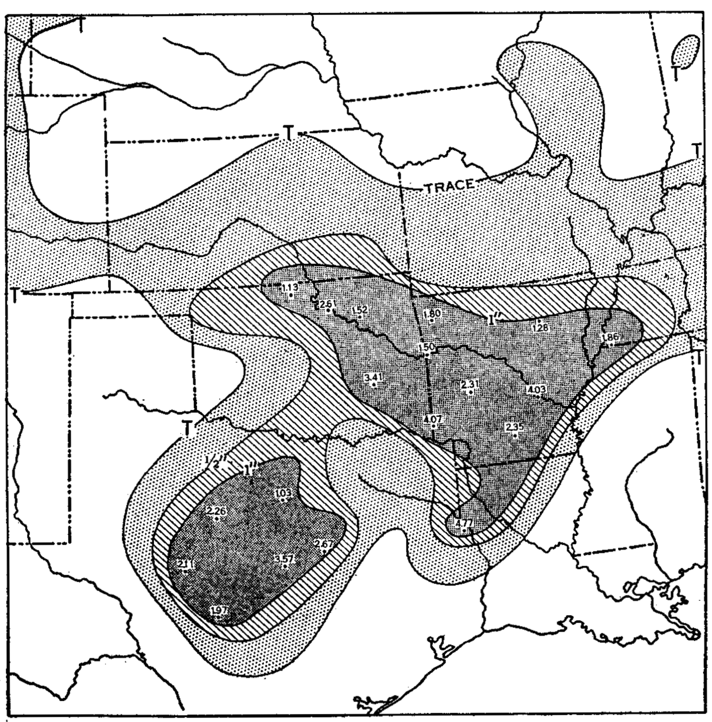

FIGURE 26.—24-hour precipitation chart ending at 1230 GMT, May 12, 1953.

An interesting feature of the spreading out of the cold air, a feature discernible in all three surges of the cold air, was its tendency to spread most rapidly towards the area of lowest sea level pressure. This can be seen on the series of maps from 1230 GMT on May 11 through 0030 GMT on May 12 (figs. 9, 10 and 12–14): the rain induced High moved from western Tennessee to the southwest following the center of 3-hourly pressure rises which was located, of course, where the accumulation and production of cold air due to influx were greatest.

CONCLUDING REMARKS

In concluding the discussion of this case, we shall summarize some of the salient aspects of the rainfall pattern since the whole physical process described here hinges on the rainfall.

850-mb. moisture charts for 0300 and 1500 GMT of the 11th, and 0300 GMT of the 12th (figs. 20–22) are included to show the effect of the rainfall on the slope of the front in the lower atmosphere. On these charts the frontal positions at 850 mb. and at the surface are entered. These charts show the transition from a steep front with a sharp moisture gradient to a flatter front with the moisture and temperature gradients obviously modified by the long-continued precipitation.

Two maximum-isotach charts, showing the location of the branches of the jet stream for 0300 and 1500 GMT, May 11 (figs. 23 and 24), are included here because of the intimate causal relationship which often exists between the jet and a narrow band of precipitation [7]. It is evident from the charts presented here that neither branch of the jet lies directly over the rainfall area in the south central United States. However, the chart for 1500 GMT shows that one maximum isotach was approaching the rain area at that time. In this case the relationship between the jet and the narrow band of precipitation is inconclusive, despite the suggestive elongated shape of the rainfall area.

Finally, 24-hour rainfall maps for May 11, 12, and 13 (ending at 1230 GMT) are included (figs. 25–27). Besides providing a graphical summary of the precipitation pattern for the period we have been discussing, these charts show that the area of maximum rainfall is coincident with the areas on the ground where maximum cooling and spreading out of the cold air took place. This close relationship between the rainfall pattern and the motion of the cold air lends credence to the original hypothesis: that warm-season precipitation falling into dry air behind a front can cause sudden accelerations of the front and endow it with attributes now generally associated only with instability lines.

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