A PUBLICATION OF THE DEPARTMENT OF METEOROLOGY
OF THE UNIVERSITY OF CHICAGO

*

Miscellaneous Reports No. 16 (1944)

THE NOCTURNAL MAXIMUM OCCURRENCE OF THUNDERSTORMS IN THE MIDWESTERN STATES

By

LYNN L. MEANS

University of Chicago

UNIVERSITY OF CHICAGO PRESS
CHICAGO · ILLINOIS

TABLE OF CONTENTS

ABSTRACT

A nocturnal maximum occurrence of thunderstorms is found in certain geographical regions.

Airway hourly observation data for Omaha show that the month of maximum occurrence of nocturnal thunderstorms falls later in the year than the month of maximum occurrence of daytime thunderstorms at Omaha. July and August are important months for the occurrence of nocturnal thunderstorms. The average duration of nocturnal thunderstorms at Omaha is greater than the average duration of daytime thunderstorms.

Factors bearing a causal relationship to the nocturnal maximum should (1) have a geographical distribution similar to that of the nocturnal thunderstorm maximum, (2) show a diurnal variation with a nocturnal maximum in the region of the nocturnal thunderstorm maximum, and (3) be sufficiently persistent to be consistent with the longer average duration of nocturnal thunderstorms.

In the study of individual occurrences at Omaha (airport elevation, 996 feet above M.S.L.) for 1941, using radiosonde data and hodograph analysis of pilot-balloon data, the principal factor contributing to instability for the formation of thunderstorms was found to be advection of warmer air in the lower layers of the atmosphere (2,000-8,000 feet above M.S.L.). Thunderstorms that occurred with advective warming in these lower layers occurred both day and night but were more frequently nocturnal.

This factor was therefore examined with regard to the above requirements with the following results:

1. Data for the 1,500-m level show a geographical area of maximum occurrence of apparent warm-air advection corresponding to the general region of nocturnal maximum occurrence of thunderstorms, while at the 3,000-m level the warm-air advective component is less.

2. The diurnal wind variations for summer months at Omaha as described by A. Wagner show stronger winds at night for the levels from 750 to 2,000 m. These winds have an average direction from regions which are warmer in the mean. Mean temperature gradients for these levels show small variations from day to night in the Middle West, with possibly slightly greater temperature gradients occurring at night.

3. Instances have been found where persistent advection of warmer

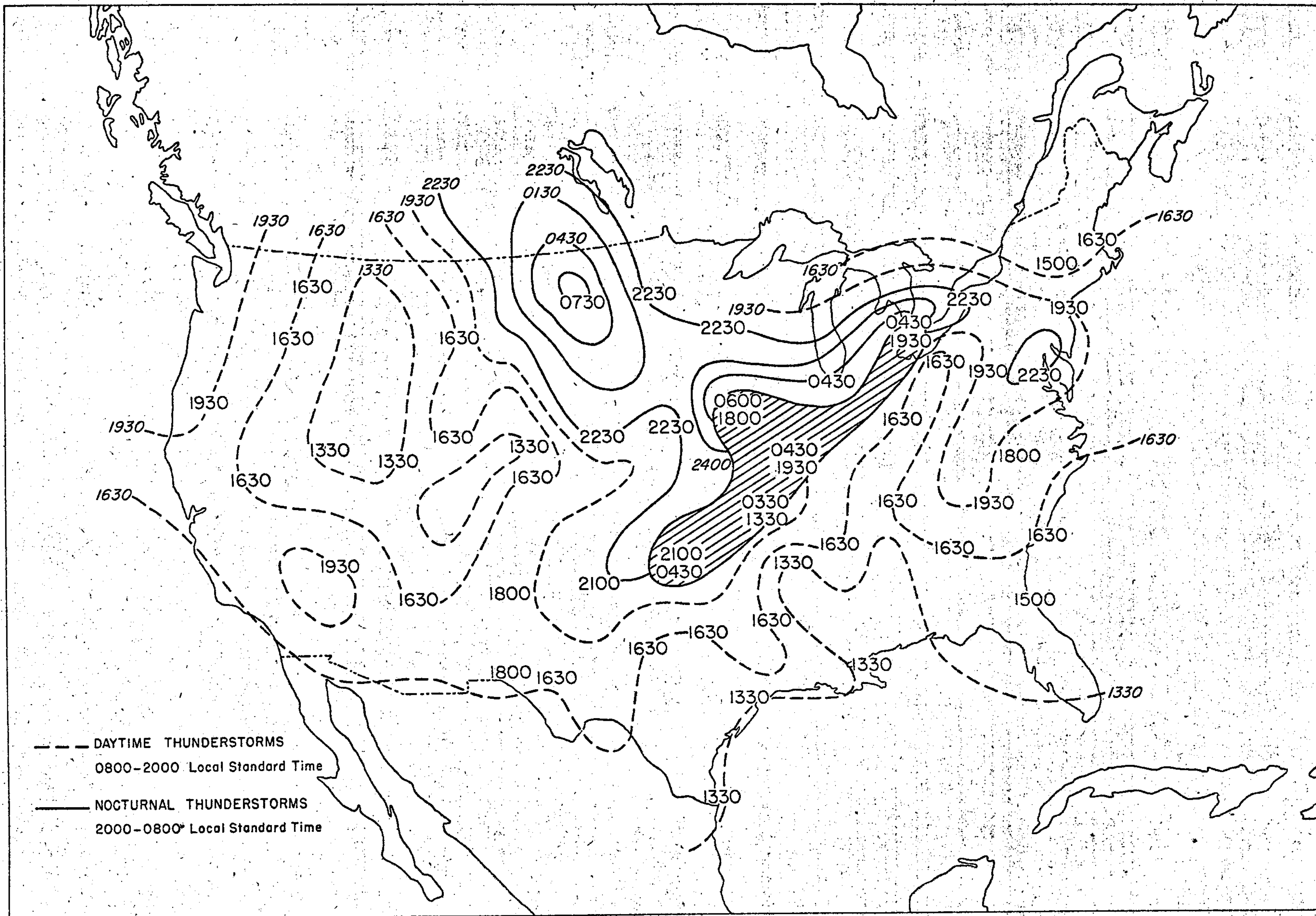

FIG. 1.—Time of maximum frequency of thunderstorm occurrence (local standard time). The hatching covers a region in which nearly equal maxima are found at two different periods (data from United States Weather Bureau, Airway Meteorological Atlas).

air in the lower layers at night has contributed to the maintenance of instability and thunderstorm activity at Omaha for more than 4 hours at a time.

These data show that the advection of warmer air in the lower layers of the atmosphere is a consistent factor as related to the nocturnal maximum occurrence of thunderstorms in the Middle West.

INTRODUCTION

The geographical distribution of the time of maximum occurrence of thunderstorms in July is given in Figure 1. The shaded area represents that portion of the United States where similar maxima are found both in the afternoon and at night. (In this report a nocturnal thunderstorm is defined as one that occurs between the hours of 8:00 P.M. and 8:00 A.M., local standard time; a daytime thunderstorm, as one that occurs between 8:00 A.M. and 8:00 P.M., local standard time). A nocturnal maximum is found at Omaha, North Platte, Kansas City, Bismarck, Minneapolis, Chicago, and also at Buffalo and Washington (data from Airway Meteorological Atlas for the United States). The most important general regions of occurrence of nocturnal thunderstorms are the Missouri River Valley and the valley of the upper Mississippi. Areas having a maximum occurrence of thunderstorms during the day are the Gulf states, most of the Atlantic states, the mountain and plateau regions of the West, and the Pacific Coast area.

The predominance of nocturnal thunderstorms in the center of the United States seems paradoxical, since many publications have for years ascribed the occurrence of summer precipitation to instability showers formed by the steepening of the lapse rate due to the surface heating of land areas by the intense insolation during spring and summer months. Daytime thunderstorms constitute one important factor in summer precipitation. However, for nocturnal thunderstorms a cause other than instability through surface heating must be found.

PURPOSE

The purpose of this paper is to present data and conclusions regarding nocturnal thunderstorms and factors bearing a causal relationship to the occurrence of a nocturnal maximum of thunderstorms at Omaha and in the surrounding regions.

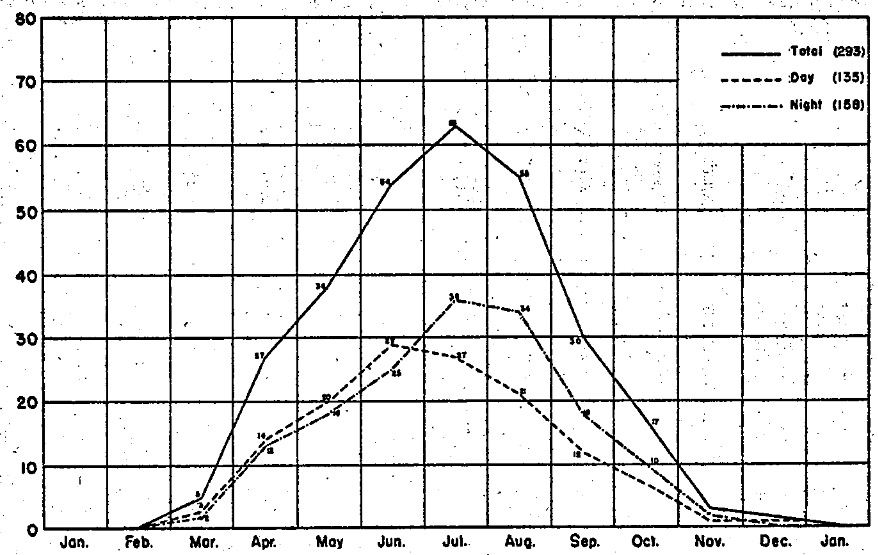

FIG. 2.—Annual distribution of number of thunderstorm occurrences at Omaha (1937-41 inclusive).

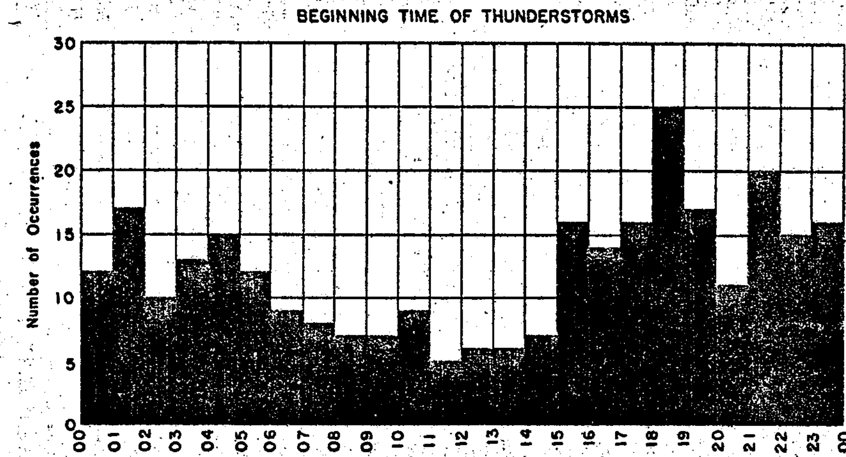

FIG. 3.—Diurnal distribution of number of thunderstorms beginning in a given hour at Omaha (1937-41 inclusive).

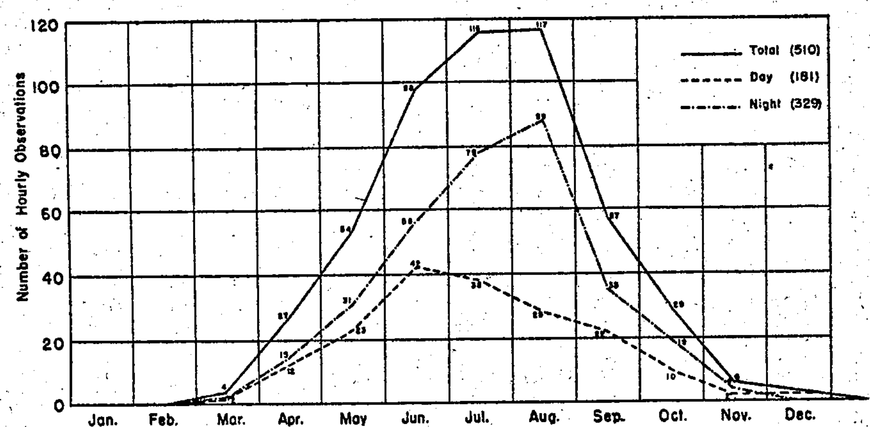

FIG. 4.—Annual distribution of number of hourly observations with thunderstorms at Omaha (1937-41 inclusive).

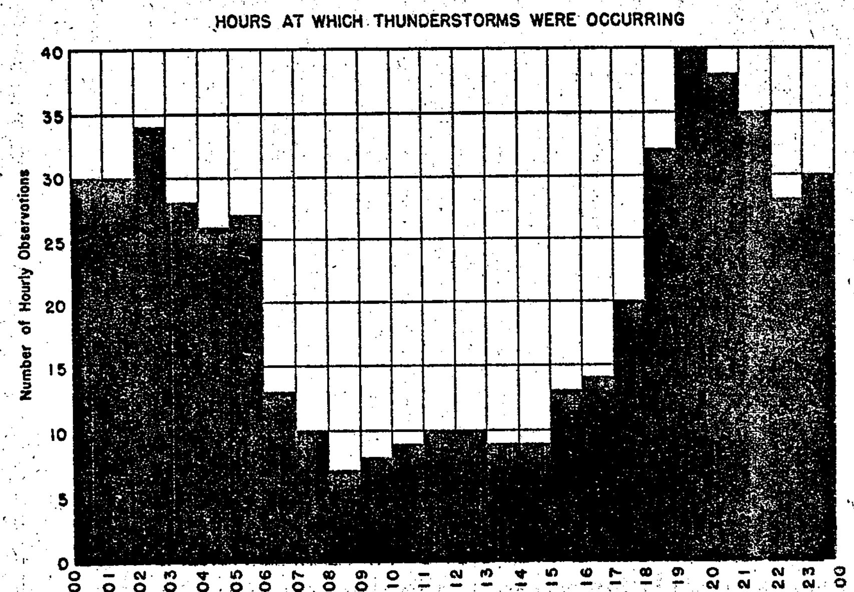

FIG. 5.—Diurnal distribution of number of hourly observations with thunderstorms at Omaha (1937-41 inclusive).

PART I

THE OCCURRENCE OF NOCTURNAL THUNDERSTORMS

1. Summary of Data

Omaha was chosen as a representative station for a detailed investigation because it is well located with reference to the area of general occurrence of nocturnal thunderstorms and because radiosonde and upper-air wind data are available for that station.

Data examined include 5 years (1937-41) of hourly airway observations, as recorded on Weather Bureau Form 1130, and one year (1941) of upper-air wind and radiosonde data.

The annual distribution of daytime and nocturnal thunderstorms is given in Figure 2. It is apparent that more nocturnal thunderstorms than daytime thunderstorms occur at Omaha and that the time of maximum occurrence of the nocturnal thunderstorms falls later in the year. July and August are the important months for the occurrence of nocturnal thunderstorms at Omaha. The June maximum for daytime thunderstorms may be explained, since in that month the air masses crossing Omaha have not attained a temperature as high as is reached in July and August, while the long days and relatively intense insolation contribute a large amount of surface heating with a marked steepening of lapse rates in the lower layers of the atmosphere by midafternoon.

In Figure 3 a diurnal distribution of the beginning time of thunderstorms is given. In these data the hours of maximum occurrence of the beginning time of thunderstorms are between 6:00 P.M. and 7:00 P.M. and between 9:00 P.M. and 10:00 P.M. (local standard time). If a smooth curve were drawn for the data, a nocturnal maximum would be shown.

An annual distribution of the number of hourly observations at which thunderstorms were occurring is given in Figure 4. Here the duration of the thunderstorms is taken into account. The month having the greatest number of hourly observations with nocturnal thunderstorms is August.

Figure 5 shows the distribution for the 24 hours of the day of the hourly observations at which thunderstorms were occurring. Comparison of Figures 3 and 5 demonstrates that nocturnal thunderstorms last longer than daytime thunderstorms. These data show the average duration of daytime thunderstorms at Omaha to be 95 minutes; the average duration for nocturnal thunderstorms is 118 minutes.

Any proposed solution of the cause of nocturnal thunderstorms must be consistent with the descriptive facts thus far presented. These facts in outline are as follows:

a) Since the nocturnal maximum of thunderstorms occurs in certain geographical regions, factors having a causal relationship should show a similar geographical distribution.

b) Since nocturnal thunderstorms occur more frequently than daytime thunderstorms at Omaha and in a large surrounding area, the causal factors should show a diurnal variation with a nocturnal maximum for that area.

c) Nocturnal thunderstorms last longer than daytime thunderstorms. Therefore, a dynamic factor must be discovered which provides sufficient energy for continuous overturning over a longer period of time than that required for the overturning of a layer of air heated from below, such as occurs in surface-heating type thunderstorms.

2. Current Theories

Several theories have been advanced to explain the nocturnal maximum occurrence of thunderstorms. Those referred to most commonly are perhaps (1) radiational cooling at the top of a cloud layer and (2) advection of cold air aloft. Another theory based on advection of warmer air in lower layers is considered in this report.

Ordinarily over land a stabilizing effect due to radiational cooling is expected, since the surface of the earth usually cools more rapidly than the air aloft. Radiational cooling of the free atmosphere does occur, but with clear conditions the order of magnitude of differences in temperature changes between two levels due to this cause is small, being about 1° C or 2° C for a 12-hour period.

Perhaps the most marked effect of radiation would be noted with a cloud layer aloft. Net radiational cooling at the top of the cloud would occur at night, since a cloud radiates approximately as a black body. The base of the cloud, on the other hand, would be absorbing quantities of heat that are radiated from the surface of the earth and from the water vapor in the atmosphere between the earth's surface and the base of the cloud. With a steep lapse rate above the cloud, vertical accelerations might possibly carry air parcels aloft, which, with the condensation of moisture, would release latent heat for continued upward acceleration, giving sufficient energy for the formation of a thunderstorm. However, some observers have pointed out that an inversion or stable layer would probably be formed just above the top of the cloud due to radiational cooling, or any stable layer or inversion that had been present would be intensified so that vertical motions from within the cloud to a region above the cloud would be suppressed.[1] Also, any temperature changes that might be expected with the addition or loss of heat by radiation to the base of the cloud or from the top of the cloud would be retarded to some degree by the release of latent heat with cooling and condensation and by the absorption of latent heat with heating and evaporation.

No evidence has been offered to show that such a radiation effect has a geographical distribution corresponding to the area in which nocturnal thunderstorms occur. Data on the mean distribution of clouds for the season June to August do not show such a geographical effect. Also, there is no immediately apparent reason that raditional cooling aloft would produce thunderstorms of longer duration than those due to surface heating.

The largest temperature changes per unit time at a fixed point in the free atmosphere are due to advection. Advection of colder air aloft therefore has been offered as an explanation of nighttime instability. Winds-aloft data might seem to support this theory. Frequently before the occurrence of a nocturnal thunderstorm the winds at the ground have a southerly direction, turning in a clockwise direction with height as viewed from above, so that at a level of 7,000 or 8,000 feet above M.S.L. the wind may have a slight component from the north. Near the surface of the earth southerly winds are usually warm and northerly winds are usually cold. However, a northerly wind aloft does not necessarily bring in colder air. As is pointed out in several texts dealing with dynamic meteorology,[2] the clockwise turning of wind with height usually implies advection of warmer air in the layer in which this occurs (in the Northern Hemisphere). No evidence has been offered to show that advection of colder air aloft has a nocturnal maximum nor that it would occur more frequently or to a greater degree in the area of frequent occurrence of nocturnal thunderstorms. It is probable that advective effects, either the advection of colder air aloft or the advection of warmer air in lower layers, could contribute continued instability over a period of time longer than that required for the overturning of a heated layer of air near the ground.

---

[1]This is an opinion privately communicated to the author by Major Harry Wexler, U.S.A.A.F.

[2]Cf. B. Haurwitz, Dynamic Meteorology (New York: McGraw-Hill Book Co., 1941), p. 150.

PART II

INSTABILITY DUE TO ADVECTION OF WARMER AIR

3. Summary of a Year's Occurrences at Omaha

In the examination of data for 1941 at Omaha, radiosonde data were plotted on pseudo-adiabatic charts for the periods before and after the occurrence of each thunderstorm. Hourly airway observation data including three-hourly cloud and pressure-change data for periods before, during, and after each thunderstorm were obtained from Weather Bureau Forms 1130-Aer. Hodographs of pilot-balloon observation data before and after thunderstorms were prepared. Of a total of 69 thunderstorms at Omaha in 1941, sufficient relevant data were found to classify 51 of the occurrences as to the principal factor contributing to instability for the formation of the thunderstorm. A summary of this classification follows:

	No. of Cases
ADVECTIVE WARMING IN LOWER LAYERS (Usually at elevations between 2,000 and 8,000 feet above M.S.L.)	28
ADVECTIVE COOLING ALOFT (Usually more than 8,000 feet above M.S.L.)	5
INSTABILITY AND CONVECTION FROM THE GROUND (Surface heating or turbulence, or both)	8
CONVECTIVE INSTABILITY (Release through frontal lifting directly with passage of cold front or instability from lifting of air up warm-front surface)	4
INSTABILITY OF SATURATED AIR WITH RESPECT TO THE PSEUDO-ADIABAT (Where long-continued precipitation was an important factor in supplying moisture for saturation of an air mass through which precipitation was falling)	6

The thunderstorms that occurred with advective warming in the lower layers were more frequently of the nocturnal type. The greater number of them occurred during the latter part of the thunderstorm season, that is, from July through October.

4. General Discussion of Instability Due to Advection of Warmer Air

Examination of radiosonde data for a number of cases with thunderstorms of this type shows a relatively stable layer of air from the ground to the gradient wind level or just above the gradient wind level. The stability of this layer may be due to radiational cooling or may result from the presence of a frontal inversion at the top of the layer.

In the case of a frontal inversion frequently no increase in moisture with height through the inversion is found, although from the surface map it is quite apparent that a frontal surface extends over the station. Above the stable layer next to the ground the air is conditionally unstable.

Frontal situations that are favorable for the formation of thunderstorms due to warm-air advection are: (a) a slow-moving cold front that has a NE-SW orientation, with the development of frontal waves southwest of the area in which thunderstorms occur; (b) a slow-moving cold front or occlusion in an elongated N-S trough to the west of the region in which the thunderstorms occur; and (c) the warm sector of a wave.

In (a) an appreciable influx of overrunning warm air to the north of the wave may contribute instability through the advection of warmer air just above the frontal surface, even though no overrunning type clouds (M1 or M2) are produced. If clouds were produced at the frontal boundary, the thunderstorms might be due primarily to convective instability. More frequently the air above the frontal inversion at Omaha is not saturated but has a relative humidity nearer to 50 per cent, with a convective condensation level near 10,000 feet above M.S.L., or higher.

Most thunderstorms occur with a frontal system within 250 miles of Omaha. Namias has shown that the occurrence of thunderstorms is related to the presence of moist tongues aloft as indicated by isentropic charts.[3] Moist tongues aloft are closely related to the presence of fronts and surface troughs of low pressure.

In a case where the thunderstorm is due to advection of warmer air in the lower layers the plot of the upper-air wind vectors on a polar diagram (hodograph) will show a pattern similar to that shown in Figure 6a prior to the thunderstorm. Hodographs of this type are found more frequently at the 2300 and 0500 pilot-balloon observations than at the 1100 and 1700

---

[3]Namias, Introduction to Air Mass and Isentropic Analysis (New York: American Meteorological Society, 1940).

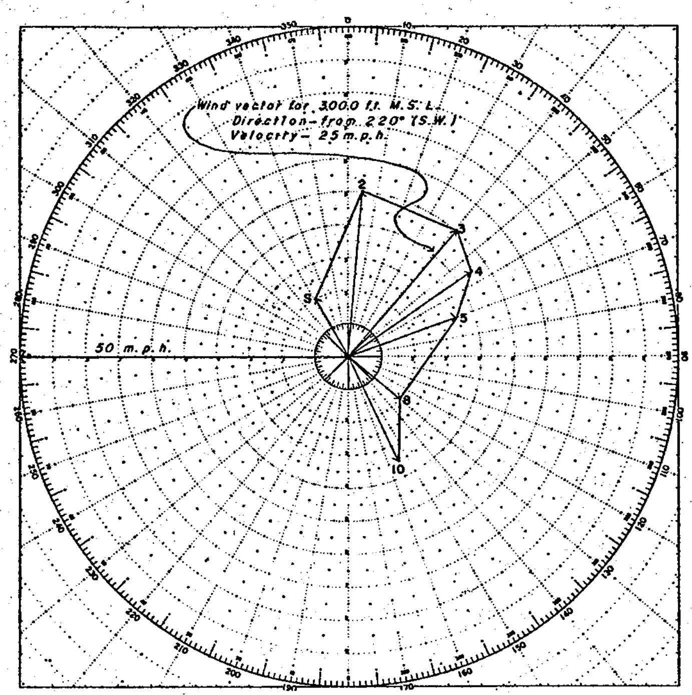

FIG. 6a.—Hodograph type for advection of warmer air in lower layers ahead of a surface trough.

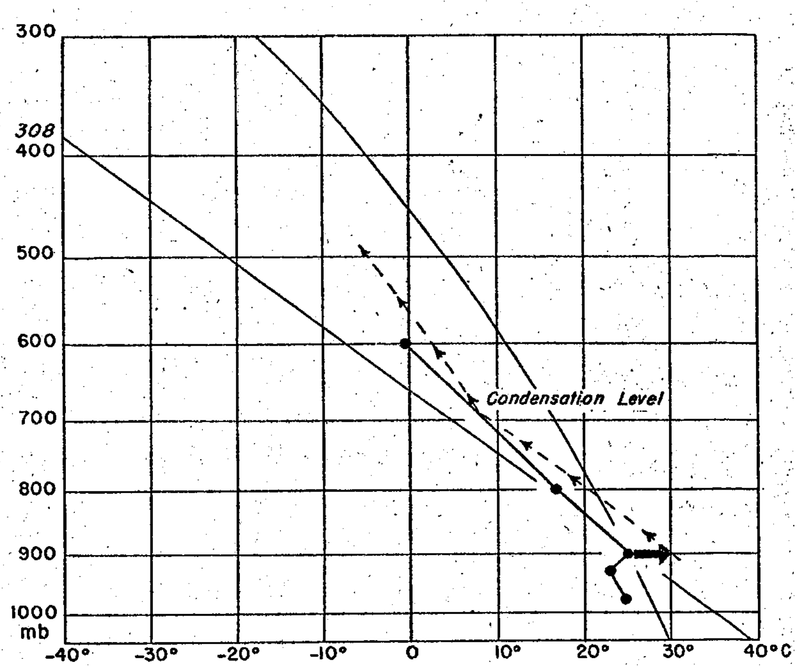

FIG. 6b.—Sounding illustrating release of latent instability by advective warming in the lower layers.

observations at Omaha. Advection of warm air as indicated by the hodograph is reflected in successive radiosonde soundings (cf. Fig. 6b) by an increase in temperature, especially between 2,000 and 8,000 feet, and by a steepened lapse rate, since advection of warm air usually occurs to a greater degree in the lower portion of the range from 2,000 to 8,000 feet. With continued advection of warmer air, instability increases and air parcels rise to their condensation level beyond which, with conditional instability, latent heat from condensation of water vapor contributes to the upward acceleration of the parcels. Continued convective lifting of air parcels will initiate the formation of a thunderstorm.

5. Synoptic Examples

1) July 9, 1941 (Omaha). (All times given are central standard time, all elevations given are in feet above M.S.L. except where specific mention is made that the elevation is relative to the surface.)

A thunderstorm began at Omaha at 0003 and lasted until 0435. The 0030 surface map (Fig. 7) shows a Pacific occlusion in an elongated slow-moving trough to the west of Omaha with a weak warm portion of the polar front in the vicinity of Omaha.

The isentropic chart for 0000 (Fig. 8) for the 314° potential

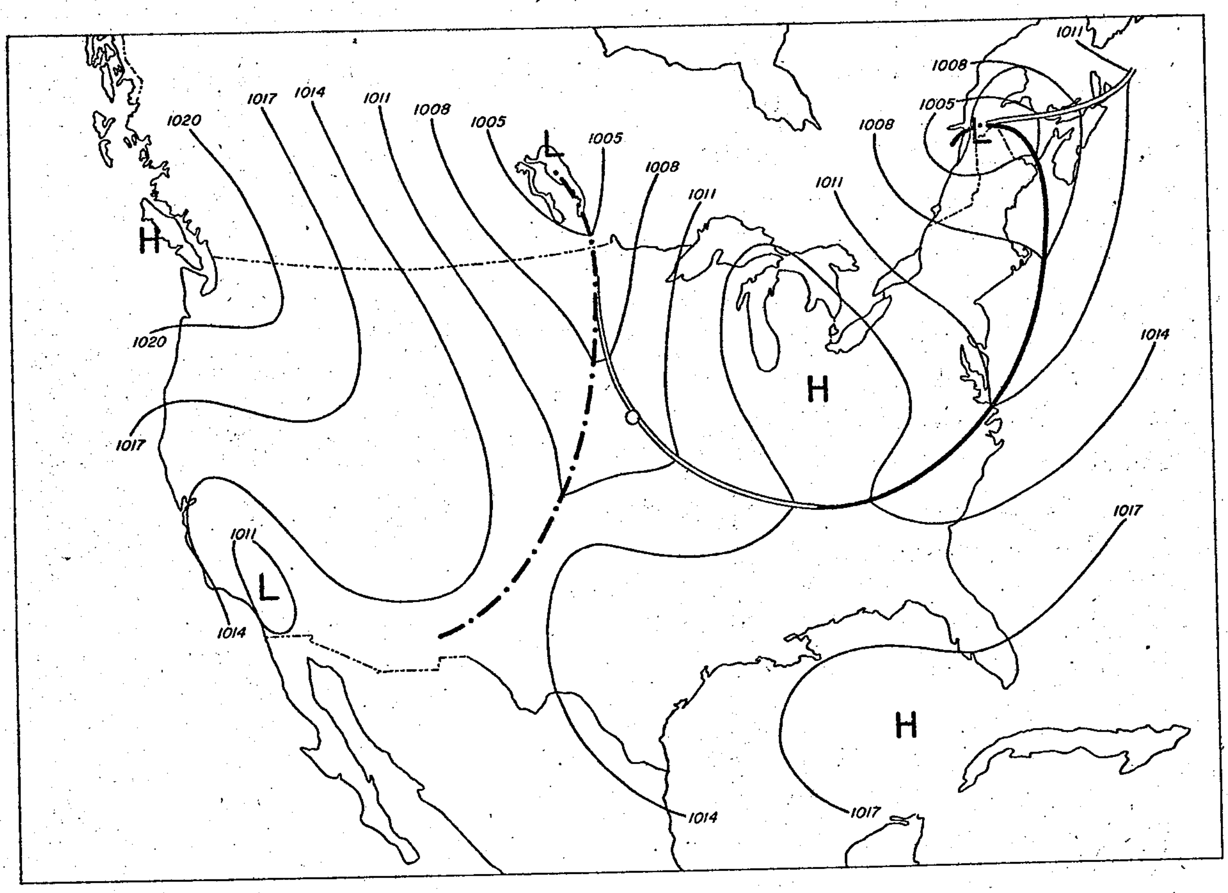

FIG. 7.—Surface map. July 9, 1941, 0030 C.S.T.

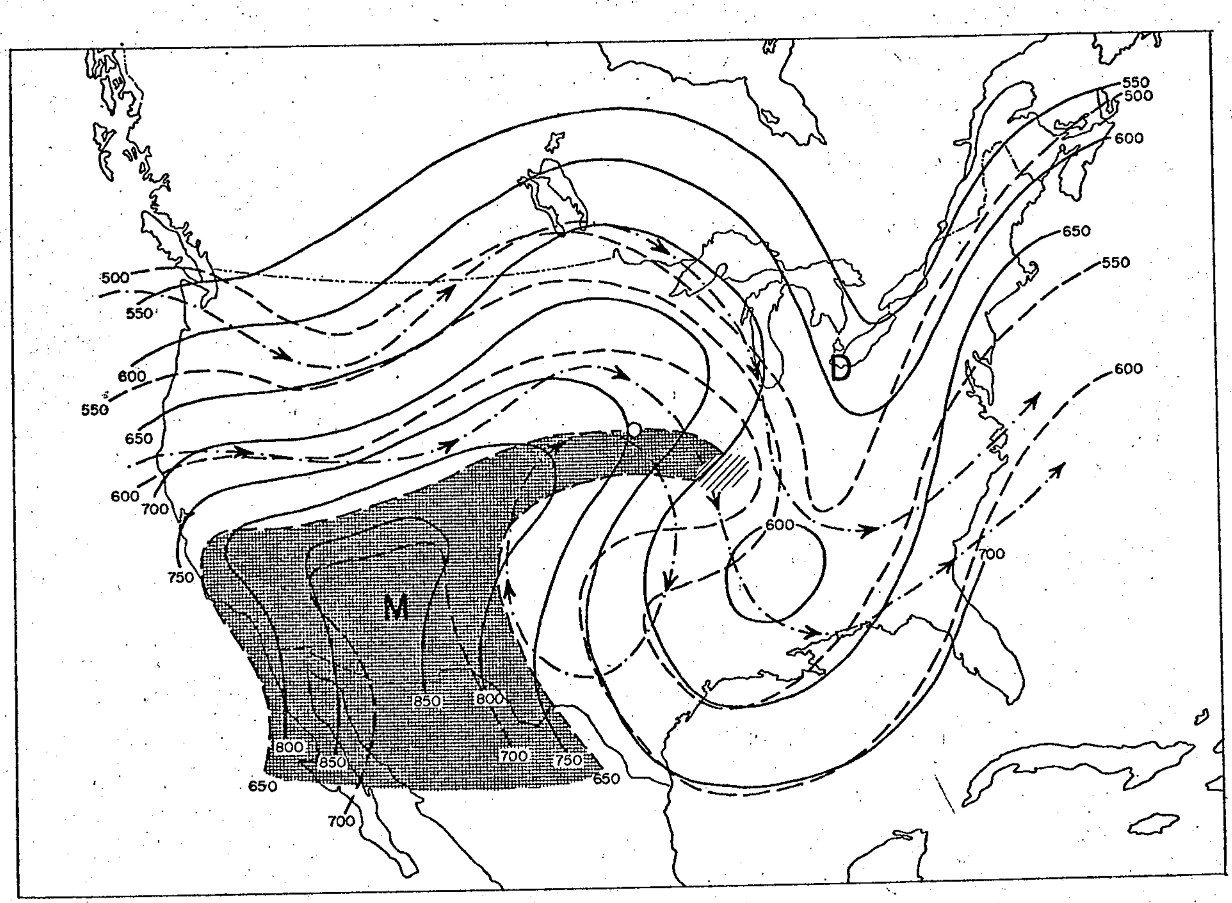

FIG. 8.—Isentropic chart, $\Theta = 314^{\circ}$A. July 9, 1941, 0000 C.S.T.

temperature surface shows an anticyclonic circulation to the SSW of Omaha. A moist tongue is located aloft immediately south of Omaha. At the tip of the moist tongue in the vicinity of St. Louis a saturation area is found. At Omaha the height of the isentropic surface is about 9,000 feet with a 100-mb difference between the actual pressure and the condensation pressure. The time of the isentropic data corresponds very closely with the time of beginning of the thunderstorm. From these data it is apparent that convective instability was not an important factor, at least not at this level, in the production of instability for the formation of the thunderstorm at Omaha.

Airway observations for the period before, during, and after the thunderstorm show the following:

7/8 2135 Sky—High scattered clouds.
Cloud types—M7 from the west and H8.
Sea-level pressure—1010.8 mb.

7/9 0030 Sky—High broken clouds.
Cloud type—L9 from the southwest.
Sea-level pressure—1010.5 mb.

7/9 0040 Rain showers began.

7/9 0335 Sky—Overcast; ceiling estimated at 7,000 feet above the surface.
Cloud type—L9 from the southwest.
Sea-level pressure—1009.1 mb.

7/9 0630 Cloud types—L2 and H6 from the northwest.
Sea-level pressure—1008.8 mb.

Surface winds are southeasterly before, during, and after the thunderstorm.

Hodographs prepared from pilot-balloon data (Fig. 9) show the following:

7/8 1000 Slight advection of warm air from 4,000 to 10,000 feet, the greatest amount occurring from 8,000 to 10,000 feet.

7/8 1600 Pilot-balloon data for Omaha missing. The sounding for Sioux City, Iowa, showed small advection of warm air from 2,000 to 7,000 feet and from 9,000 to 12,000 feet, the greatest amount occurring from 4,000 to 6,000 feet.

7/8 2200 Advection of warm air from 2,000 to 6,000 feet and from 7,000 to 10,000 feet, the greatest amount occurring from 2,000 to 3,000 feet.

7/9 0400 Strong advection of warm air from surface to 9,000 feet (the highest level reached by the sounding), the greatest amounts occurring from 3,000 to 4,000 feet and from 6,000 to 7,000 feet.

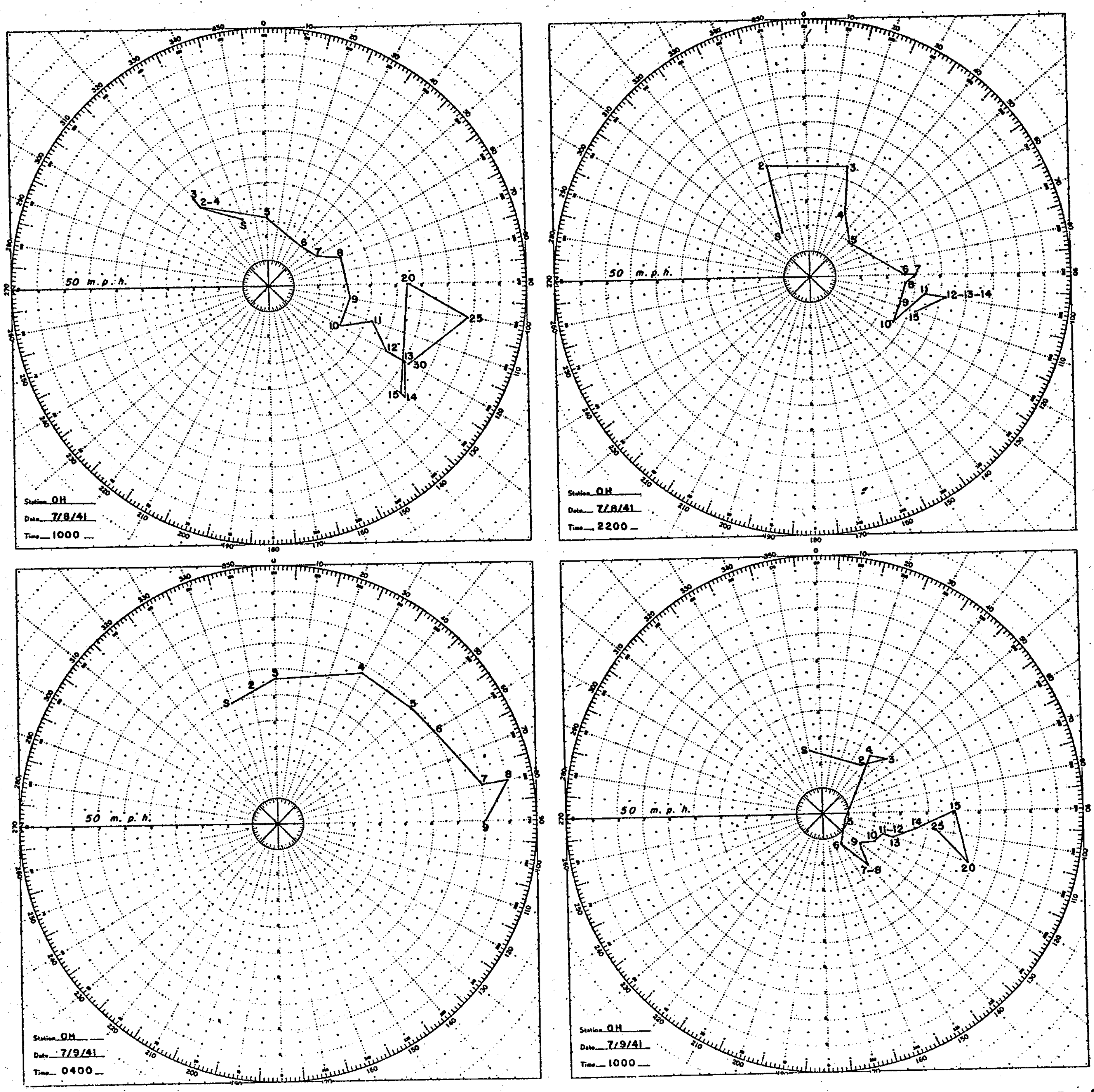

FIG. 9.—Hodographs of pilot-balloon soundings before and after thunderstorm. Omaha, July 8 and 9, 1941.

(Since this sounding was taken during a thunderstorm, it may be representative of a local circulation rather than a large scale circulation.)

7/9 1000 This sounding indicates relatively little advection of either warmer or colder air.

The pilot-balloon data reveal marked advection of warm air in the lower layers just prior to the thunderstorm. During the thunderstorm the 0500 sounding shows a very great advection of warm air below the base of the cumulonimbus. The 1000 sounding, taken after the thunderstorm ended, reveals that the advection of warmer air was cut off.

To illustrate the effect of the warm-air advection that is implied by the wind distributions shown in Figure 9, the magnitudes of the advective temperature changes averaged for each thousand-foot layer from 2,000 feet to 10,000 feet were computed for the 2200 sounding. These computations are tabulated in Table 1.

TABLE 1

ADVECTIVE TEMPERATURE CHANGES AVERAGED FOR EACH 1,000-FOOT LAYER, JULY 9, 1941, 2200 C.S.T.

Layer	Change in °C/12 Hr
2-3	26.4°C
3-4	4.4°C
4-5	3.8°C
5-6	9.6°C
6-7	0.0°C
7-8	2.4°C
8-9	5.2°C
9-10	5.8°C

In plotting actual temperature-pressure curves on pseudo-adiabatic charts, stability conditions may not be clearly indicated unless corrections to obtain virtual temperatures are made for the presence of any large quantities of moisture in the air. During the thunderstorm season significant differences may be found between the virtual temperature correction for air close to the ground and that aloft, the largest correction being that for air near the ground. The correction near the ground can easily be as much as 3° C, while that at 10,000 feet may be about 1° C. Since the adiabatic rate of change of temperature with pressure is very nearly the same for moist unsaturated air as for dry air, the slope of the curve of virtual temperature plotted against pressure may be compared with the slope of the ordinary dry adiabats on a pseudo-adiabatic chart for a determination of relative stability. The more accurate indication of the stability or instability of the air as shown by the plot of the virtual temperature rather than the dry-bulb temperature may be sufficient at times to cause a change in a thunderstorm forecast. For this reason, virtual temperatures have been used in evaluating the soundings considered here.

The following are summaries of the radiosonde data (see Fig. 10):

7/8 1200 Steep lapse rate in lowest layers, topped by an isothermal layer. Conditional instability aloft.

7/9 0000 Cooling has occurred close to the ground since the time of the previous sounding. Marked warming has occurred in the layer of air just above the inversion, the advective warming being indicated by the pilot-balloon data. Slight cooling occurred aloft. This cooling is not indicated by the pilot-balloon observation previous to this radiosonde sounding. The cooling may be related to vertical motions in near-by convection cells, since the thunderstorm had begun at the time of this sounding. Also, the cooling may have been related to adiabatic lifting of the higher layers resulting from convergence in the lower layers. An important point is that the condensation level for air parcels near the bottom of the layer having the steep lapse rate is below the region where a net cooling is shown between this sounding and the previous sounding. Therefore, warm-air advection was the more important factor in releasing the instability for thunderstorm formation, although the cooling aloft would contribute energy to the thunderstorm activity.

7/9 1200 Further warming is shown throughout the sounding. The warming in the lower layers could be accounted for as further advective warming. Aloft the warming was due to mixing with the potentially warmer air below during the thunderstorm and due to the release of latent heat with condensation of moisture during the thunderstorm. The surface temperature is about the same as that of the day before. The mean temperature for a layer from the surface to 10,000 feet is higher than that of the day before. There is no evidence that the passage of the occluded front occurred before the time of this sounding.

The pilot-balloon and radiosonde data show that the lapse rate was steepened above a shallow stable layer by advection of warmer air in the lower levels. The steepened lapse rate made possible the ascent of air parcels from relatively low elevations. These air parcels contained larger amounts of moisture than the air at higher elevations. The convective condensation level for the air parcels rising from the 4,000-foot or 5,000-foot levels was near 10,000 feet, while the sounding showed conditional instability above. The resulting cumulonimbus cloud was indicated as having a base above 9,750 feet. Later, with continued shower activity, the base of the cloud lowered. The long duration (about $4\frac{1}{2}$ hours) of thunderstorm activity and the continued advection of warm air as indicated by the hodograph are significant.

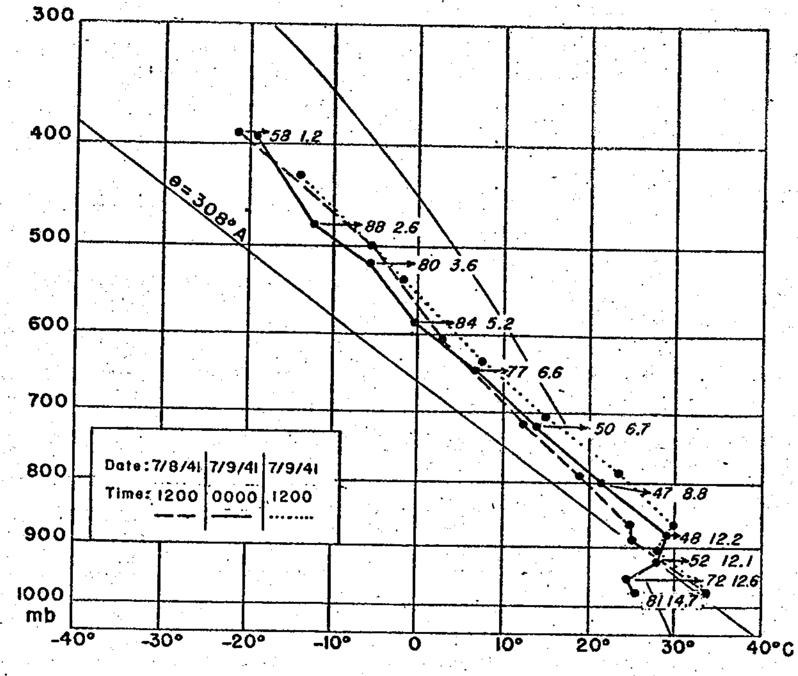

FIG. 10.—Pressure-temperature curves of radiosonde soundings before and after thunderstorm. Omaha, July 8 and 9, 1941.

ii) <u>May 16, 1941 (Omaha)</u>. (All times given are central standard time, all elevations given are in feet above M.S.L., except where specific mention is made that the elevation is relative to the surface.)

A thunderstorm at Omaha was recorded for the interval 0120-0220. At 0030 the surface map (Fig. 11) shows that a frontal wave had passed eastward south of Omaha with the wave crest just east of a position directly south of Omaha at map time.

The 0000 isentropic chart (Fig. 12) indicates a SW-NE oriented moist tongue to the southeast of Omaha. Condensation areas are found on the East Coast and on the West Coast but not in the Middle West for this particular isentropic surface. The streamflow in the vicinity of Omaha is approximately parallel to the isentropic contours. Following are the airway observations:

5/15 2135 Sky—Scattered clouds at 7,000 feet above the surface.
Cloud type—M3 from the south.
Sea-level pressure—1010.2 mb.

5/16 0030 Sky—Broken clouds at 7,000 feet above the surface.
Cloud type—M3 from the southwest.
Sea-level pressure—1013.2 mb.

5/16 0120 Sky—Overcast at 7,500 feet above the surface.
Sea-level pressure—1013.9 mb.

5/16 0335 Sky—High overcast.
Cloud type—M2, direction unknown.
Sea-level pressure—1014.2 mb.

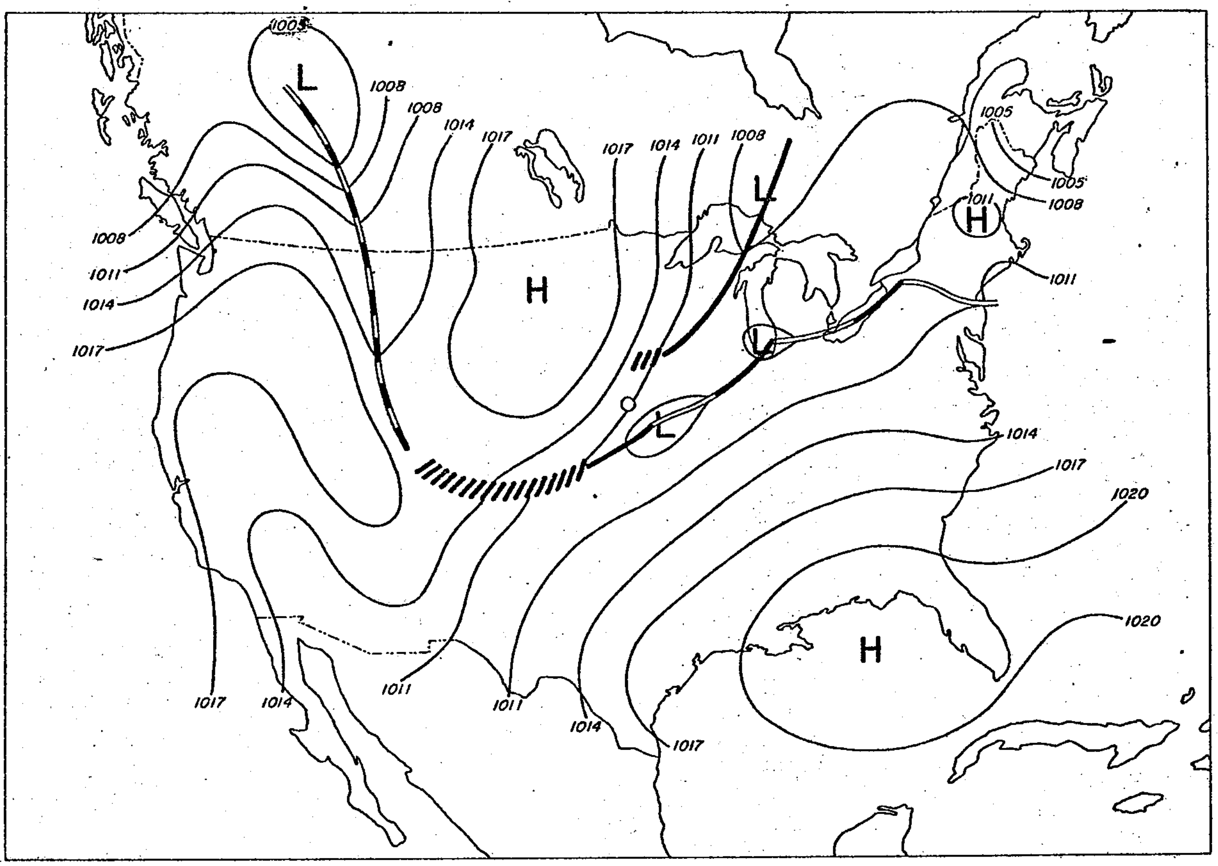

FIG. 11.—Surface map. May 16, 1941, 0030 C.S.T.

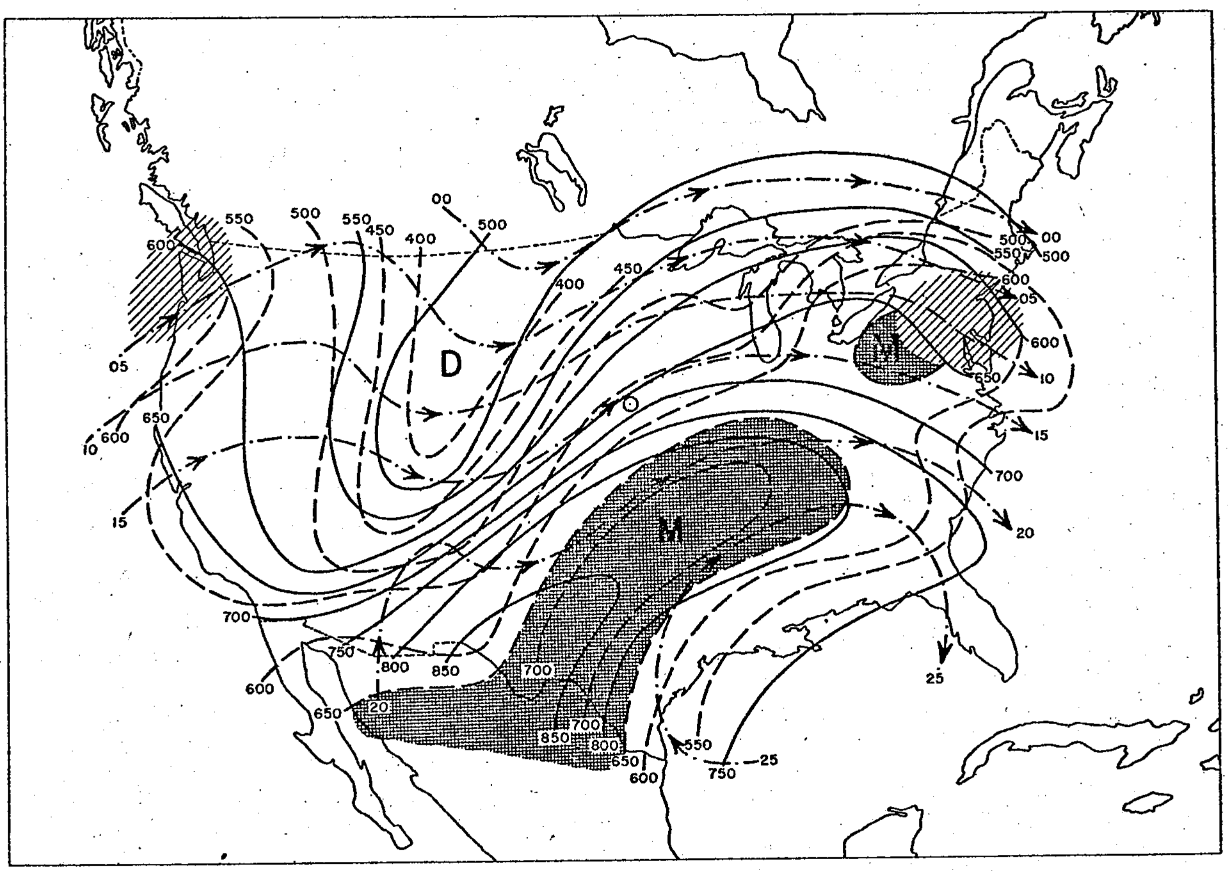

FIG. 12.—Isentropic chart, $\theta = 308^{\circ}A$. May 16, 1941, 0000 C.S.T.

Surface winds were N, varying to NNW and NNE at times with a velocity of 8-20 mph.

Advection as indicated by wind hodographs (Fig. 13) was as follows:

5/15 1000 Advection of warmer air from the surface to about 12,000 feet.
Also from 14,000 to 15,000 feet and from 20,000 to 25,000 feet.
Top of sounding—25,000 feet.
Greatest amount of warm-air advection 9,000 to 10,000 feet.

5/15 1600 Warm-air advection to the top of the sounding at 7,000 feet.
Greatest warm-air advection 2,000 to 4,000 feet.

5/15 2200 Warm air advection to 9,000 feet.
Top of sounding—10,000 feet.
Greatest warm-air advection—2,000 to 4,000 feet.

5/16 0400 Slight cold-air advection surface to 3,000 feet.
Slight warm-air advection 4,000 to 7,000 feet.
Top of sounding—7,000 feet.

Following are summaries of the radiosonde data (Fig. 14):

5/15 1200 Steep lapse rate to 900 m.
Cloud at 900 m at base of inversion (relative humidity 94 per cent).
Dry inversion to 1,300 m.
Conditional instability to 3,300 m.
Inversion to 3,700 m.
Conditional instability from 3,700 to 6,300 m.

5/16 0000 Steep lapse rate from surface to 700 m.
Inversion to 900 m.
Isothermal layer to 1,500 m.
Conditional instability from 1,500 to 3,200 m.
Cloud layer from 2,800 to 3,200 m (relative humidity 90 per cent to 100 per cent).
Isothermal layer from 3,200 to 3,600 m.
Conditional instability from 3,600 to 5,300 m.

5/16 1200 Much cooler in the lower levels.
Lapse rate about equal to the moist adiabatic.

The afternoon and evening soundings on the 15th show marked advection of warmer air in the lower layers. Altocumulus clouds were present at midnight as indicated by the airway observations and as indicated at a corresponding level by the radiosonde observation. The altocumulus clouds appeared at the top of a layer from 1,500 to 3,200 m having a steep lapse rate. The advection of a greater amount of warmer air at the bottom of the layer than at the top contributed to the steepness of the lapse rate, although a net cooling of the entire column occurred as the wave passed eastward. At the top of the cloud the isothermal layer retarded vertical

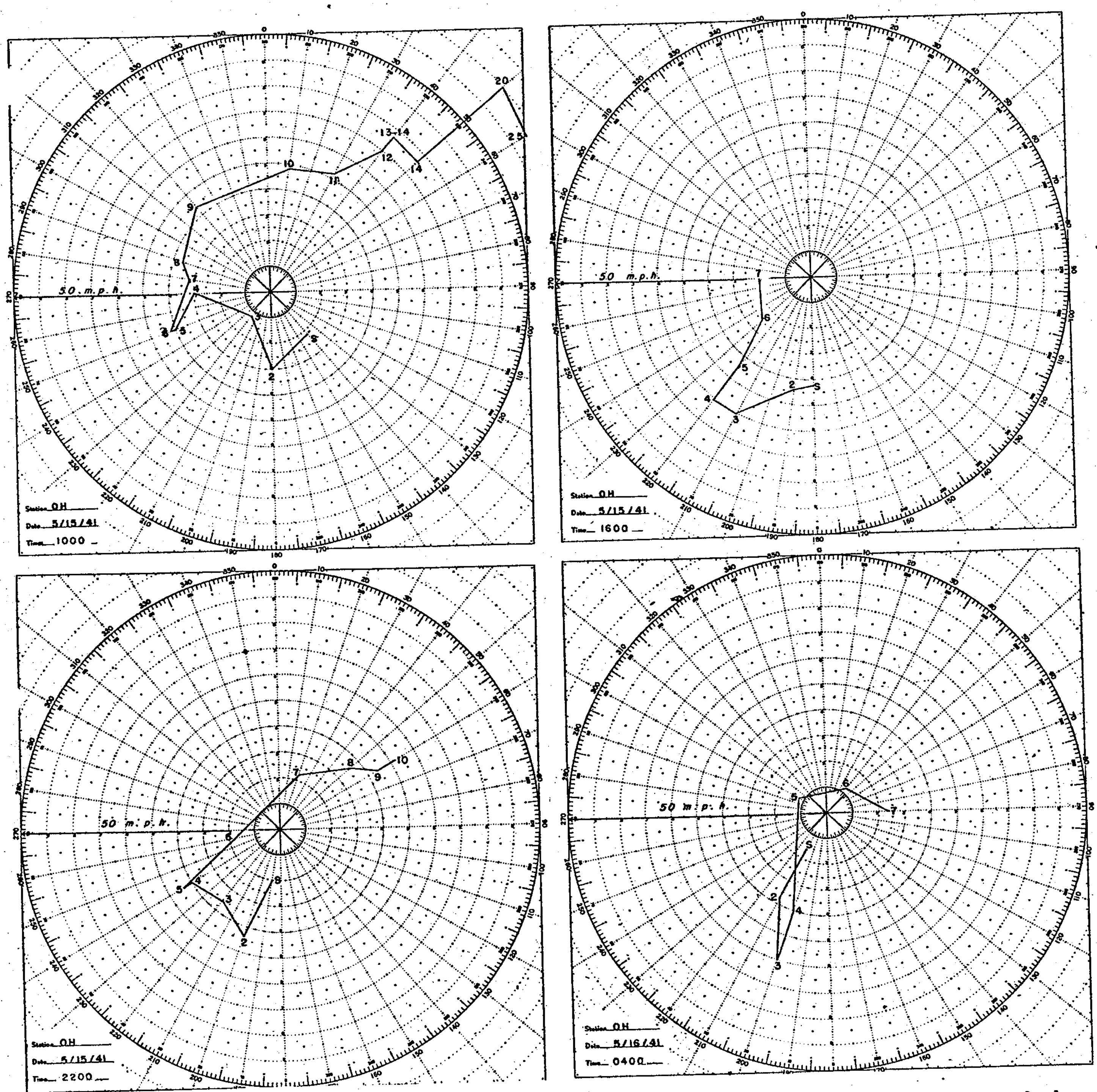

FIG. 13.—Hodographs of pilot-balloon observations before and after thunderstorm. Omaha, May 15 and 16, 1941.

motions until sufficient advection of warmer air occurred in lower layers to cause convective action to overcome the stabilizing effect of the isothermal layer.

iii) September 19, 1942 (Chicago). (All times given are central standard time, all elevations given are in feet above M.S.L. except where

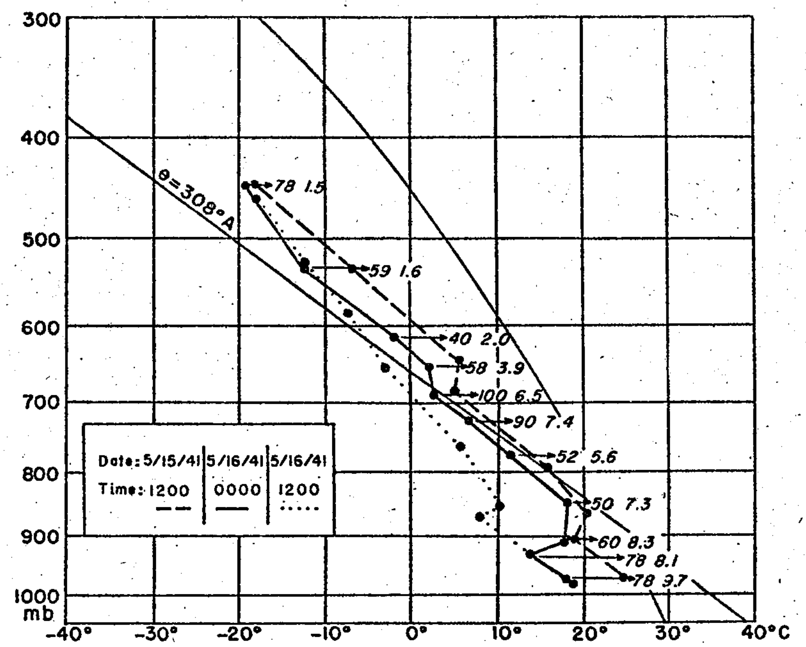

FIG. 14.—Pressure-temperature curves of radiosonde soundings before and after thunderstorm. Omaha, May 15 and 16, 1941.

it is specifically stated that the elevation given is relative to the ground.)

The thunderstorm began at Chicago at 0030 on the 19th and ended at 0130. The surface map for 0030 (Fig. 15) shows an elongated trough west of Chicago with a NNE-SSW oriented cold front within 150 miles of the station.

The isentropic chart for 0000 (Fig. 16) indicates that a moist tongue is located over Chicago with a condensation area a short distance to the west. The isolation of the area of maximum moisture as indicated by the 800-mb value for the condensation pressure suggests that moisture is being added to this region not by direct lateral mixing but by vertical transport from below.

The sequence of airway observations:

9/18 1930 Sky—Thin, high scattered clouds with lower scattered clouds at 3,000 feet above the surface.
Cloud types—L4, M6 from the SW, and H5.
Sea-level pressure—1009.1 mb.

9/18 2130 Sky—Thin, high overcast.
Cloud type—M1.

9/18 2230 The time of this airway observation corresponds very closely to that of the radiosonde observation.
Sky—Scattered clouds at 3,000 feet above the surface.

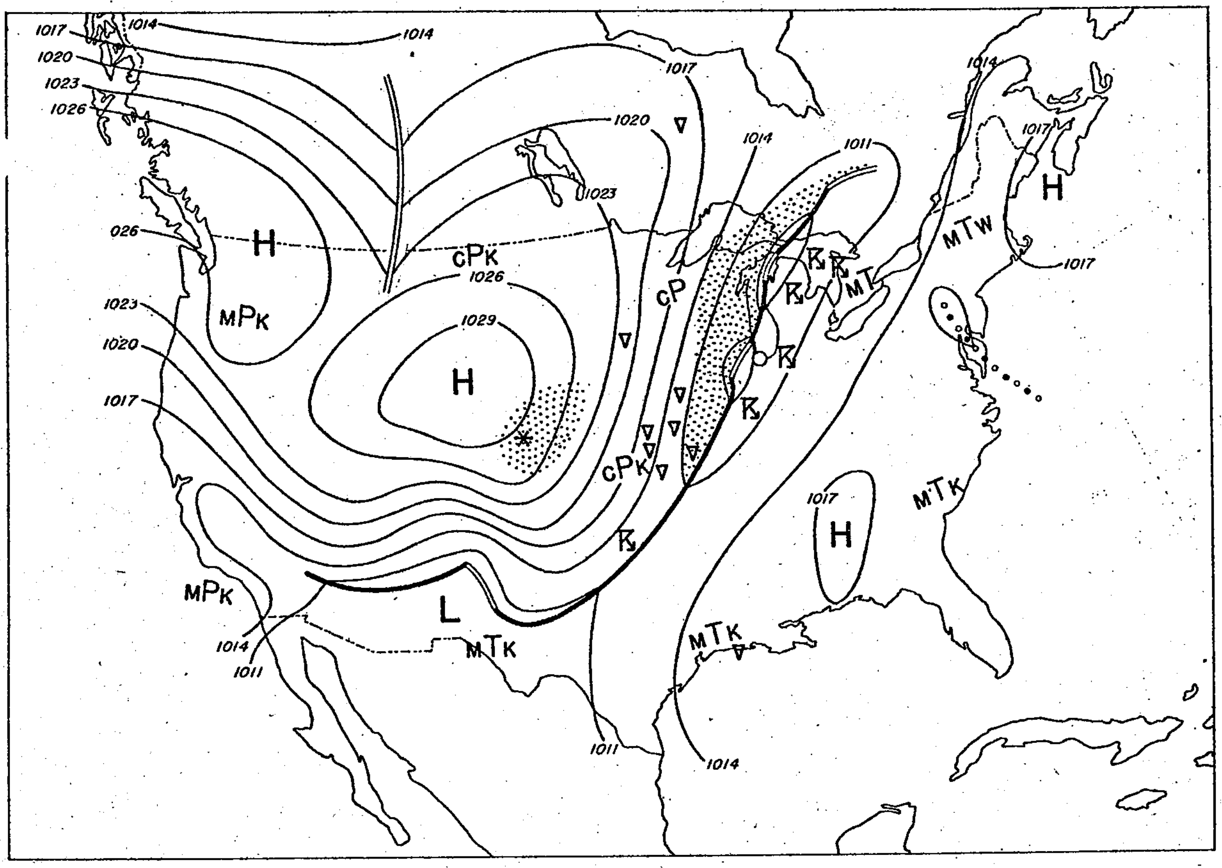

FIG. 15.—Surface map. September 19, 1942, 0030 C.S.T.

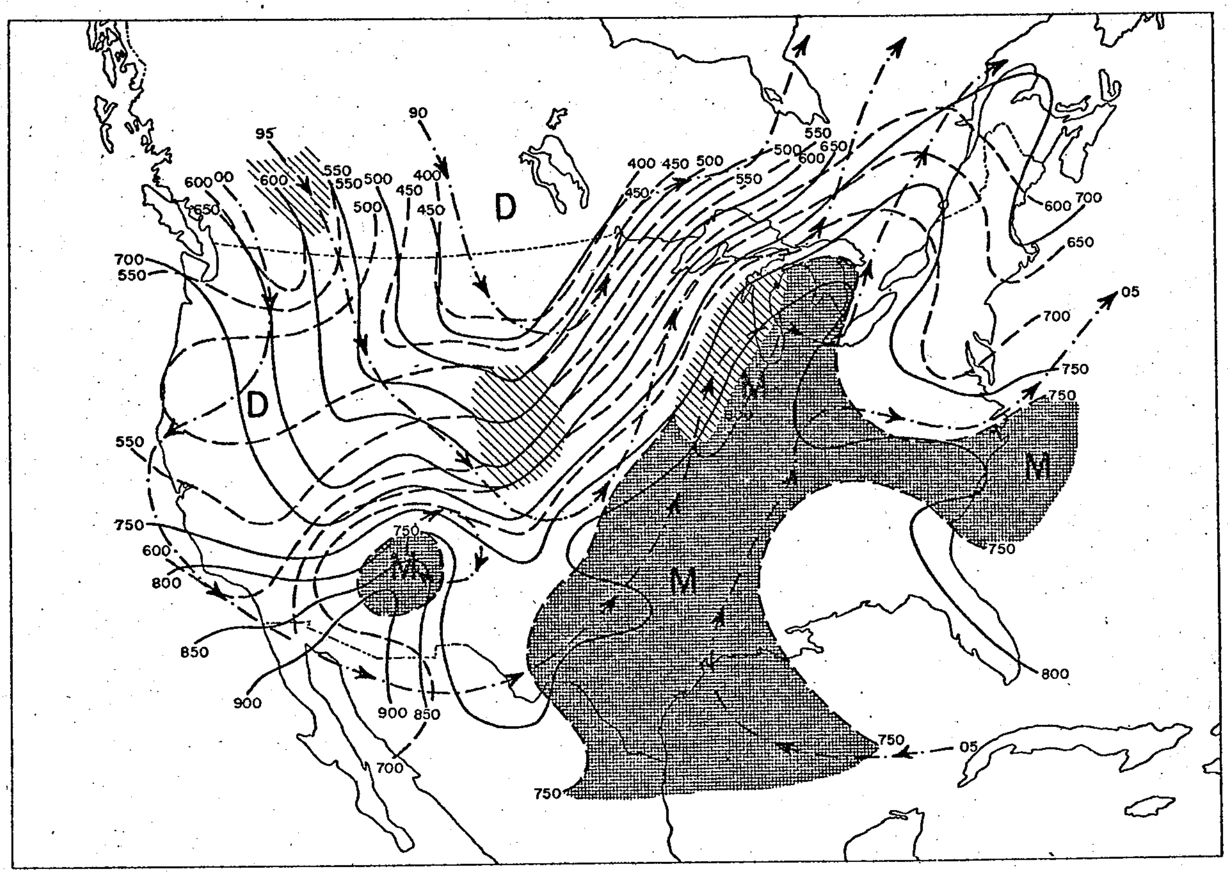

FIG. 16.—Isentropic chart, $\theta = 314^{o}A$. September 19, 1942, 0000 C.S.T.

(The radiosonde data suggest that the base of any low clouds present should have been higher than 3,000 feet above the surface.)

9/18 2330 Sky—Scattered clouds at 4,400 feet above the surface. (This height is more nearly consistent with the radiosonde data.)

9/19 0030 Sky—High overcast with lower broken clouds at an estimated height of 4,000 feet above the surface.
Cloud types—L3 and M7.
Sea-level pressure—1008.5 mb.
The thunderstorm with light rain showers began at this observation.
The surface wind continued from a southerly direction with velocities ranging from 8 to 15 mph.

9/19 0130 Sky—High overcast, lower scattered clouds at 4,000 feet above the surface.
The surface wind shifted briefly to NW with a velocity of 5 mph but had shifted back to a southerly direction by the time of the next observation.

Hodographs of the wind observations (Fig. 17) indicated advection as follows:

9/18 JO 1600 The pilot-balloon observation includes data to only 5,000 feet. There was no significant advection of warmer or colder air below that level.

9/18 JO 2100 Marked advection of warmer air between 2,000 and 3,000 feet and between 5,000 and 7,000 feet.

9/19 JO 0300 Slight advection of warmer air to 3,000 feet. Advection of colder air from 3,000 to 8,000 feet. This advection of colder air aloft may have contributed instability for the occurrence of rain beginning at 0530 and continuing until after the frontal passage which occurred at about 0930, when the wind shifted from WSW to NW and the pressure began a continuous rise.

A radiosonde observation (Fig. 18) was made at 2200:

9/18 JO 2200 This sounding shows a shallow ground inversion with the surface temperature lower than at the time of the previous sounding (9/18 1000). From a level about 100 m above the ground to about 1,800 m, M.S.L., a net warming had occurred. Relative humidities were more than 80 per cent to about 10,000 feet. Above the 872-mb level the lapse rate was very nearly equal to the dry adiabatic with a less steep lapse rate and relative humidities of 92 per cent reported for the layer between 2,100 and 3,000 m, M.S.L. Continued advective warming in the lower layers could easily result in the release of real conditional instability. The layer between 3,500 and 4,700 m, M.S.L., had a lapse rate that was approximately equal to the dry adiabatic. Particles rising from the base of this layer to the top would have reached

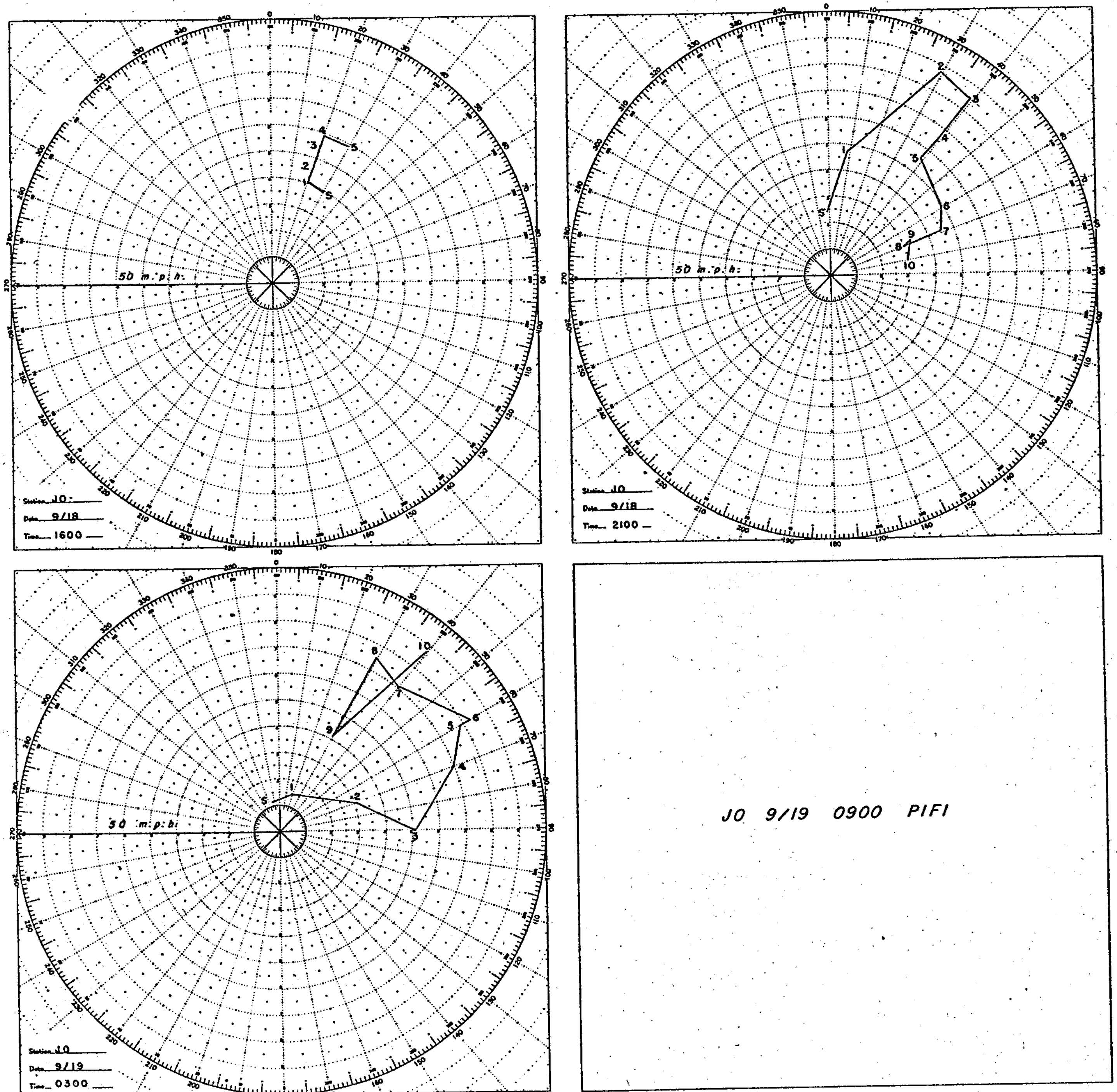

FIG. 17.—Hodographs of pilot-balloon observations before and after thunderstorm. Joliet, September 18 and 19, 1942.

their condensation level, but the more stable conditions above would have damped out further vertical motion. The lower layer between the 1,300 and 2,200 m, M.S.L., levels presents more potential possibilities for real instability. The 9/19 1000 sounding shows the effect of the frontal passage in the lowest layers.

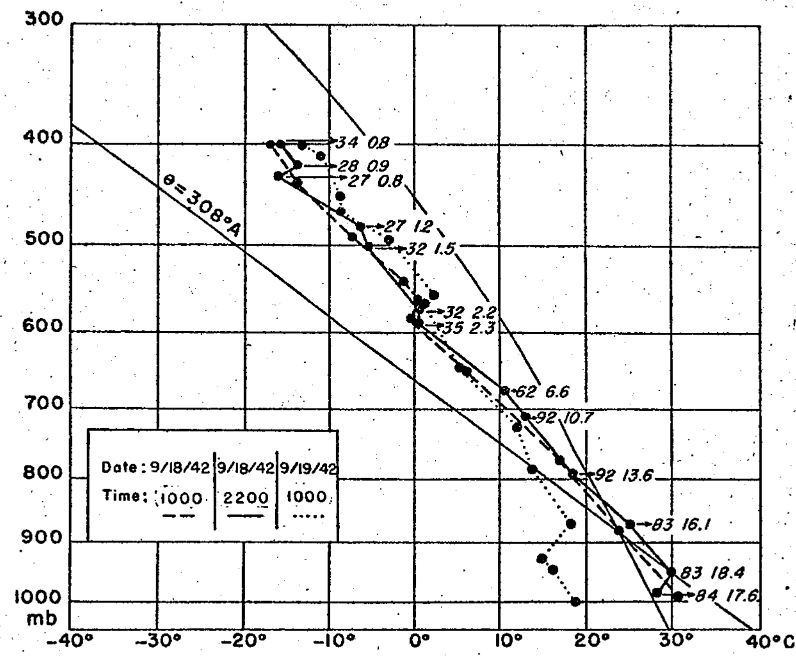

FIG. 18.—Pressure-temperature curves of radiosonde soundings before and after thunderstorm. Joliet, September 18 and 19, 1942.

The thunderstorm began several hours after the pilot-balloon observation had shown advection of warmer air in the lower layers and after the radiosonde observation, which showed conditional instability that could be released easily by continued advection of warmer air in the lower layers.

6. Supporting Evidence

If advection of warm air in the lower levels is a plausible explanation for the nocturnal maximum occurrence of thunderstorms in the Middle West, then such advection should (i) have a geographical distribution similar to that of the maximum occurrence of nocturnal thunderstorms, (ii) show a nocturnal maximum, and (iii) be sufficiently persistent to contribute to overturning for a period of time long enough to account for the greater average duration of nocturnal thunderstorms.

i) Geographical distribution of warm-air advection in the lower levels.—Data from the Monthly Weather Review showing the mean monthly isobars and isotherms for July and August, 1941, for the 5,000-foot and 10,000-foot levels are presented in Figures 19, 20, 21, and 22. In these figures the winds correspond to the 0500 sounding and the pressures and temperatures correspond to the 2300 sounding.

On the 5,000-foot charts the principal areas in which there is a well-marked temperature gradient, together with a sizable component of

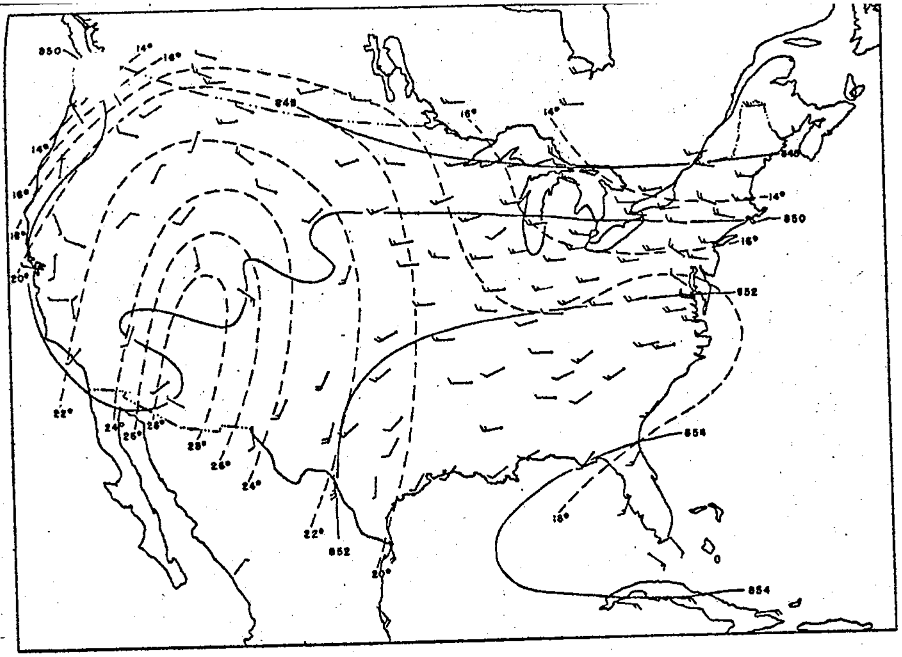

FIG. 19.—Mean isobars and isotherms at 5,000 feet, July, 1941 (data from Monthly Weather Review).

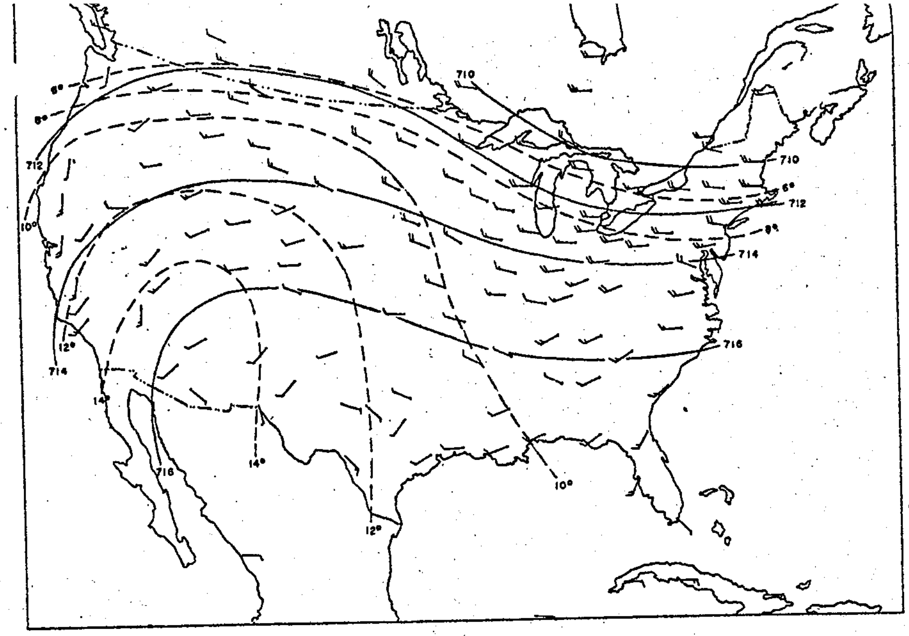

FIG. 20.—Mean isobars and isotherms at 10,000 feet, July, 1941 (data from Monthly Weather Review).

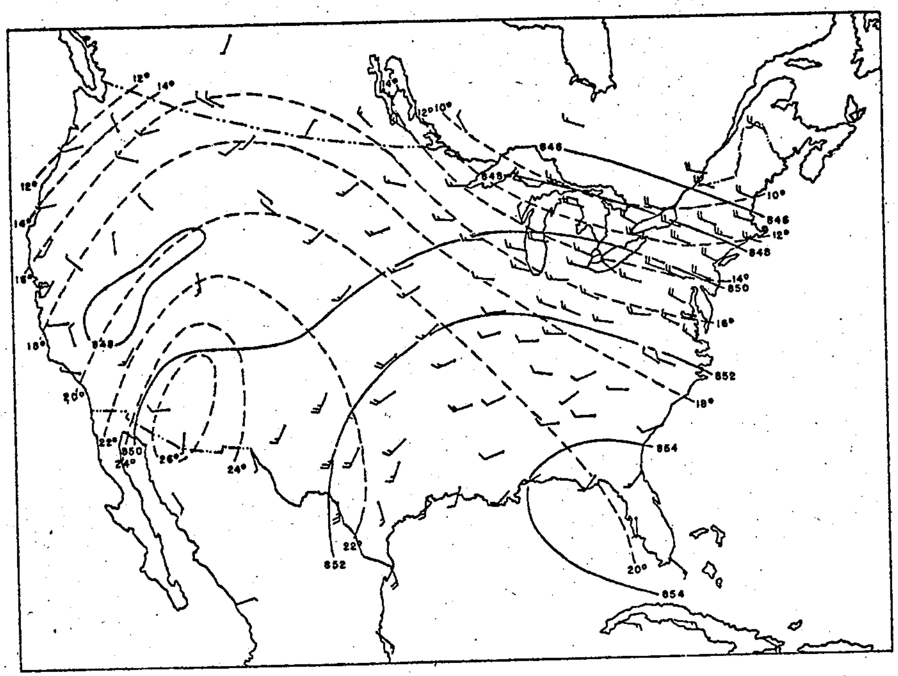

FIG. 21.—Mean isobars and isotherms at 5,000 feet, August, 1941 (data from Monthly Weather Review).

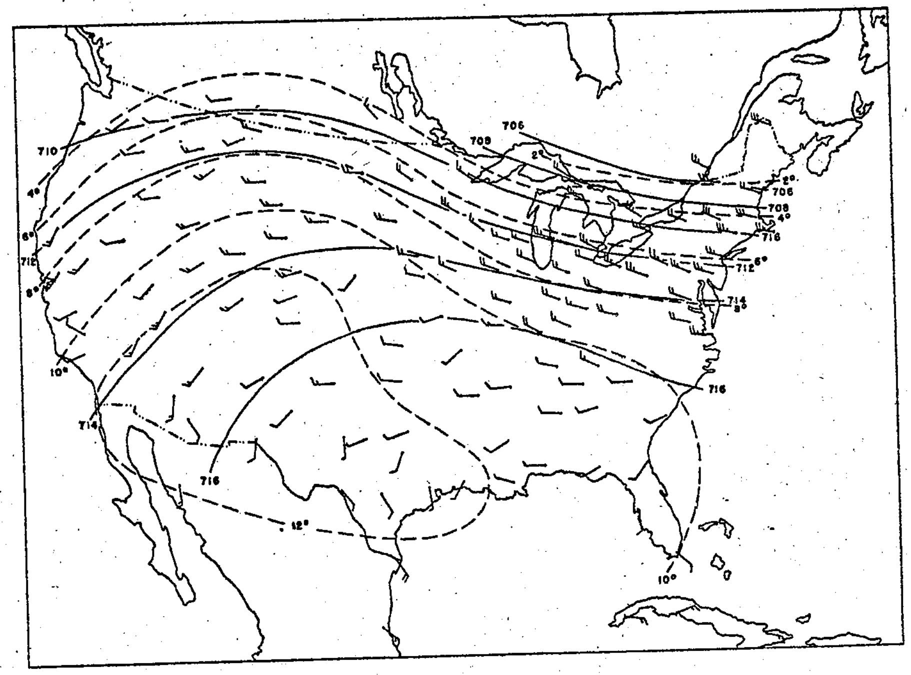

FIG. 22.—Mean isobars and isotherms at 10,000 feet, August, 1941 (data from Monthly Weather Review).

the winds or of gradient flow perpendicular to the isotherms, are immediately apparent, since they are inclosed by isotherms on two sides and by isobars on their remaining sides. In general, the smaller the area inclosed by isobars and isotherms drawn for given intervals, the greater will be the advective effect. On the 10,000-foot charts fewer inclosed areas of this type are found. A simple comparison of such areas on the 10,000-foot chart and the 5,000-foot chart is disturbed by the fact that a given geostrophic wind velocity at the 10,000-foot level corresponds to a larger spacing of the isobars than is found for the same geostrophic wind velocity at the 5,000-foot level. However, this factor is not sufficiently large to account for all the difference in amounts of apparent advection of warmer air indicated between the 5,000-foot level and the 10,000-foot level. At the higher level, where the wind velocities are relatively high, the wind direction is more nearly parallel to the isotherms; where the winds are nearly perpendicular to the isotherms, the wind velocities are low. Greater warm-air advection is shown by these data to occur, assuming approximately horizontal flow, nearer 5,000 feet than 10,000 feet. Summer mean monthly charts for other years show similar patterns.

Winter 5,000-foot mean charts show that the mean isotherms and mean isobars at that level are more nearly parallel during that season.

Similar temperature data are given in Figures 23 and 24. These data are shear vectors for the layer from the first standard level (500-, 1,000-, or 2,000-m level reported) to 3 km and from 3 to 5 km calculated from June, July, and August wind resultants (the figures comprise daytime and nighttime data taken from the <u>Airway Meteorological Atlas</u>). The shear vectors are parallel to the mean isotherms in their respective layers, cold air being on the left of the arrow for an observer facing the direction toward which the arrow points. The lengths of the arrows (when divided by the thickness of the layer) are approximately proportional to the mean horizontal temperature gradients. The largest temperature gradients are found mostly between longitudes 90° and 105° W. The region with the largest advective components of warm air, considering both the thermal gradients for the lower layer and the wind components normal to the isotherms, is again similar to the region of nocturnal maximum occurrence of thunderstorms. The advective components of warm air for the upper layer as indicated by thermal gradients and components of wind normal to the isotherms are less than the warm-air advective components for the

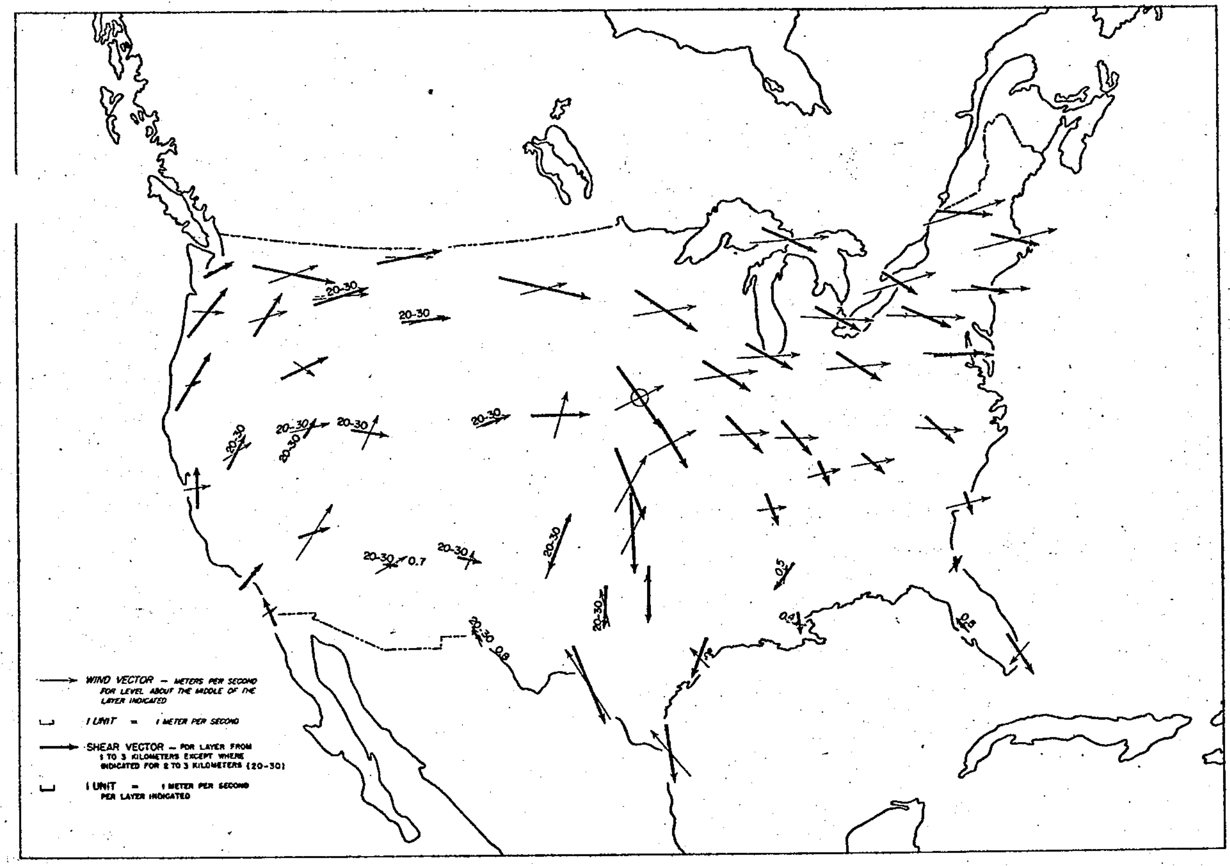

FIG. 23.—Mean wind vectors and wind-shear vectors for the layer of air below 3 km determined from June, July, and August, 1941, wind resultants (data from Airway Meteorological Atlas).

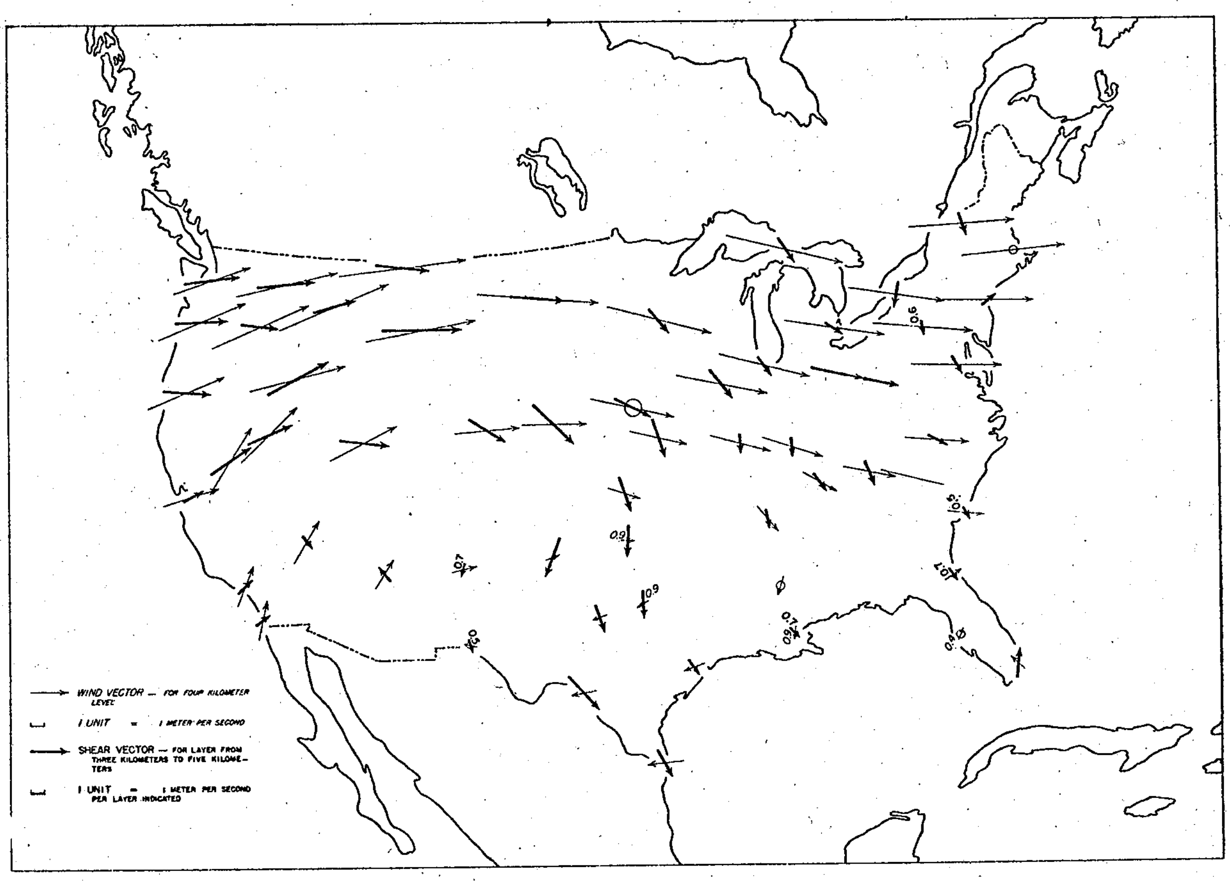

FIG. 24.—Mean wind vectors and wind-shear vectors for the layer of air from 3 to 5 km determined from June, July, and August, 1941, wind resultants (data from Airway Meteorological Atlas).

lower layer in the above-mentioned area.

A separate evaluation of the same data is presented in Figures 25 and 26. In constructing these charts, wind-resultant data for stations over the entire United States were plotted on polar coordinate diagrams and the net areas under the resulting curves evaluated to give values which are usually associated in individual instances with the amount of advection of warmer or colder air. Figure 25 gives isolines for such areas. The numerical value (in units of m^2/sec^2) recorded in each instance is equal to twice the area swept out by the wind vectors plotted as a hodograph, multiplied by the sine of the latitude. Areas under curves which turned in a clockwise direction are considered to be negative; areas under curves which turned in a counterclockwise direction are considered to be positive. In the midwestern states the net negative areas suggest a net advective warming in the mean condition.

Since mean data give an order of magnitude of the advective effect amounting to several degrees centrigrade per day (approximately 3.0° C per day at Omaha, averaged for the layer between 500 and 2,500 meters, M.S.L.), and since mean temperatures in the Middle West do not show continuous large rises during the summer months as evidenced by climatological data, the mean warmer-air advection must be counteracted to some degree by other dynamic or thermodynamic factors.

The following equation, after Petterssen,[4] is obtained for the variation of temperature with time at a fixed point in the atmosphere:

$$\frac{\partial T}{\partial t} = \frac{1}{c_p} \frac{dQ}{dt} - \underline{v}_h \cdot \nabla_h T - v_z (\gamma_d - \gamma),$$

where

$\frac{\partial T}{\partial t}$ = Time rate of temperature change at a fixed point in the atmosphere

c_p = Specific heat of air at constant pressure

$\frac{dQ}{dt}$ = Time rate of change in heat content of an individual unit mass of air

$\underline{v}_h$ = Horizontal wind velocity

---

[4]S. Petterssen, Weather Analysis and Forecasting (New York: McGraw-Hill Book Co., 1940), p. 30.

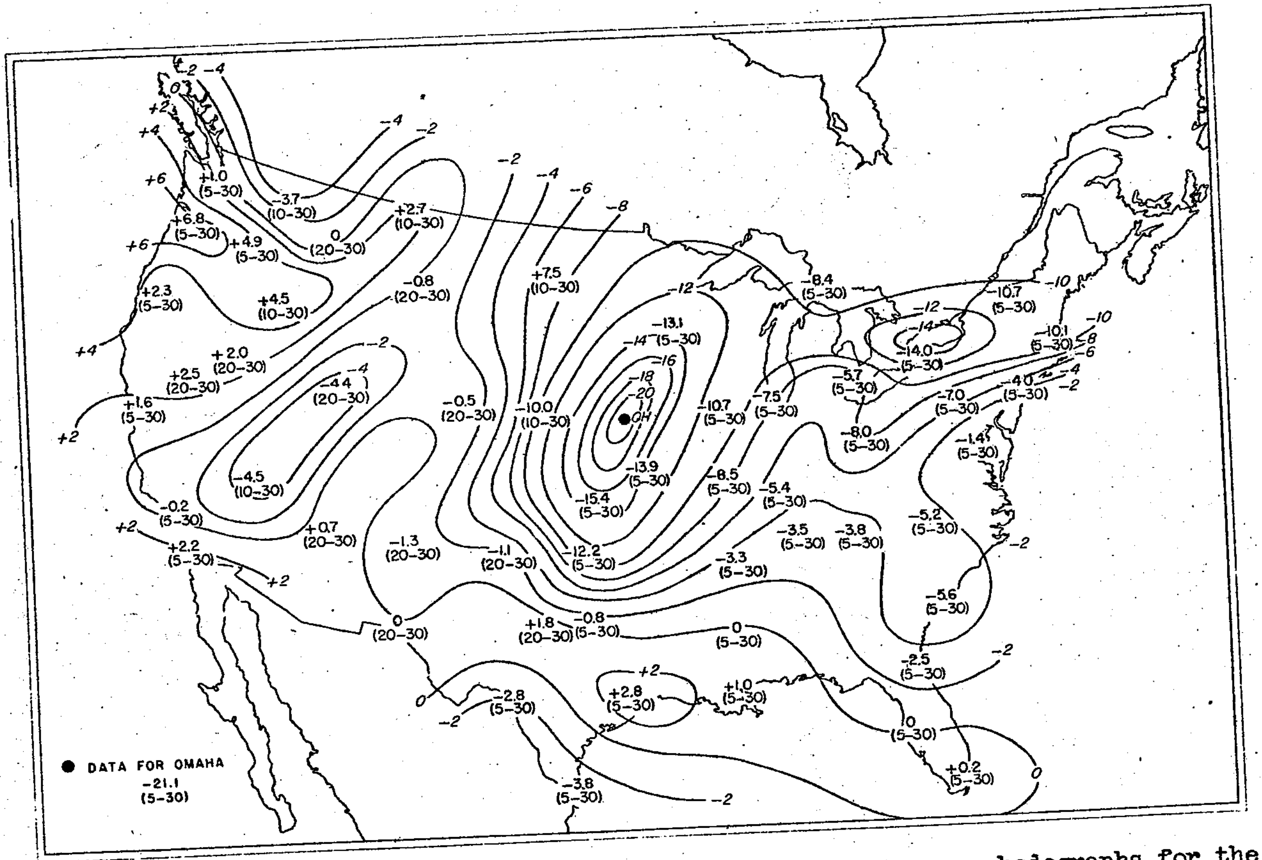

FIG. 25.—Isolines of areas swept out by mean wind vectors on hodographs for the layer of air below 3 km determined from June, July, and August, 1941, wind resultants (data from Airway Meteorological Atlas).

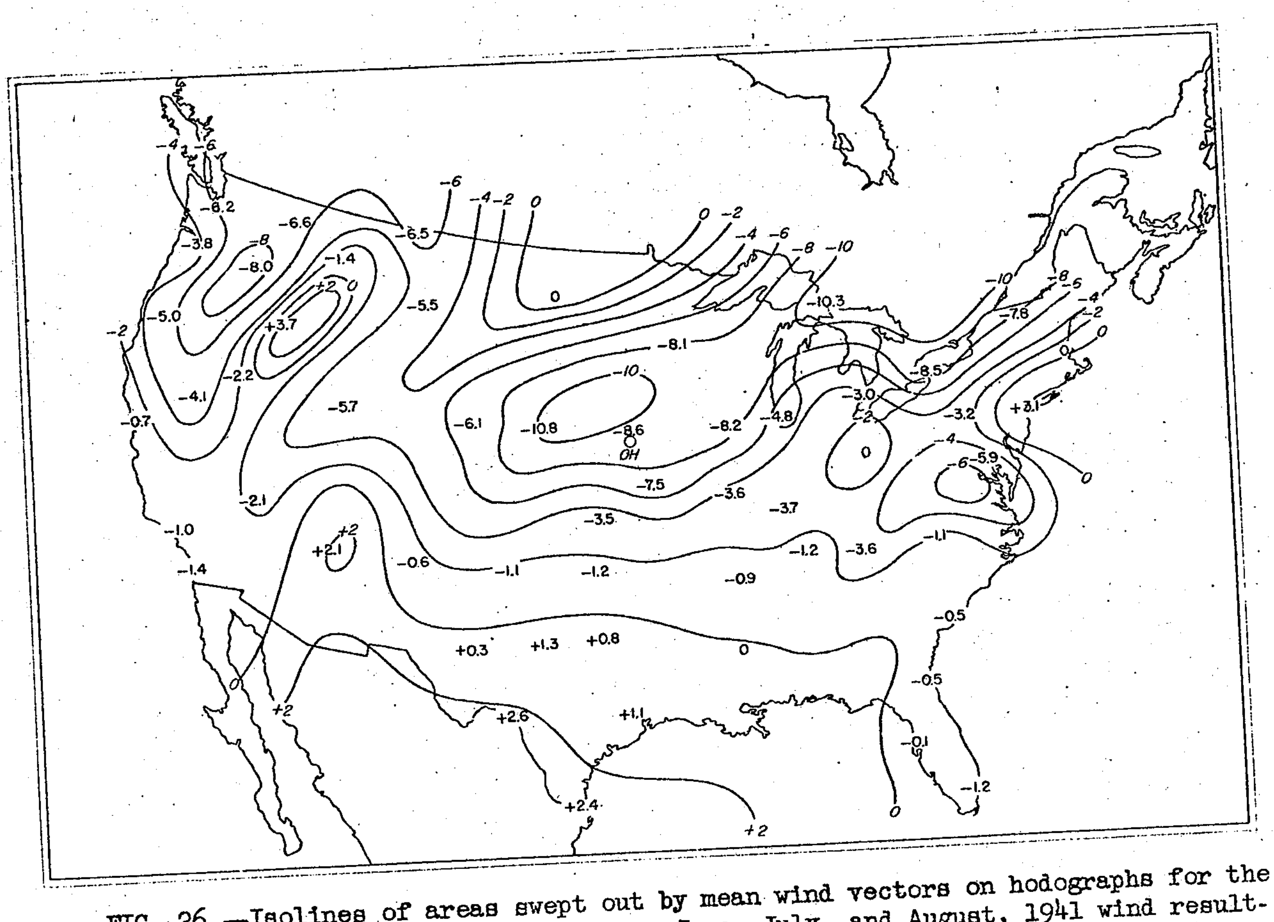

FIG. 26.—Isolines of areas swept out by mean wind vectors on hodographs for the layer of air from 3 to 5 km determined from June, July, and August, 1941 wind resultants (data from Airway Meteorological Atlas).

$\nabla_h T$ = Horizontal temperature gradient

v_z = Vertical wind speed

γ_d = Dry adiabatic lapse rate

γ = Actual lapse rate

In arriving at this equation, certain small terms involving the local pressure change and the horizontal pressure gradient have been neglected. The equation shows that at a fixed point the temperature variation is dependent upon (a) the absorption or loss of heat, (b) the horizontal advection, (c) the vertical wind speed, and (d) the stability of the air. Over a long period of time (e.g., from early summer to late autumn) the net change of temperature at a fixed point must be considered to be approximately zero. Then it is apparent from the equation that the principal dynamic factor tending to balance the net advection of warmer air must be vertical motions of the air.

Non-adiabatic thermodynamic factors may also be of importance in balancing a net advection of warm air, as indicated by the first term on the right-hand side of the above equation. Radiational cooling is not known to show a geographical distribution similar to that of the area of nocturnal occurrence of thunderstorms. However, in individual cases, evaporation of falling precipitation has been noted as being effective in lowering the temperatures in the lower portions of air masses even though the winds showed continued advection of warmer air. In the mean condition much summer nocturnal precipitation occurs in the area of nocturnal thunderstorms. Thus evaporation could be greater in this region and therefore a factor in the balancing of net advection of warmer air.

A comparison of the geographical region in which clockwise turning of the resultant winds sweeps out the greatest areas on the hodographs with the geographical region of maximum occurrence of nocturnal rainfall for the warmer part of the year (Fig. 27) is interesting. There appears to be a close correlation between the two. These regions also correspond closely with the above-mentioned region of maximum advection of warm air at 1,500 m in the monthly mean data.

ii) The nocturnal maximum of advection of warmer air.—A. Wagner describes the diurnal wind variation that occurs at Omaha and at several other stations in the United States from data for the months of May to

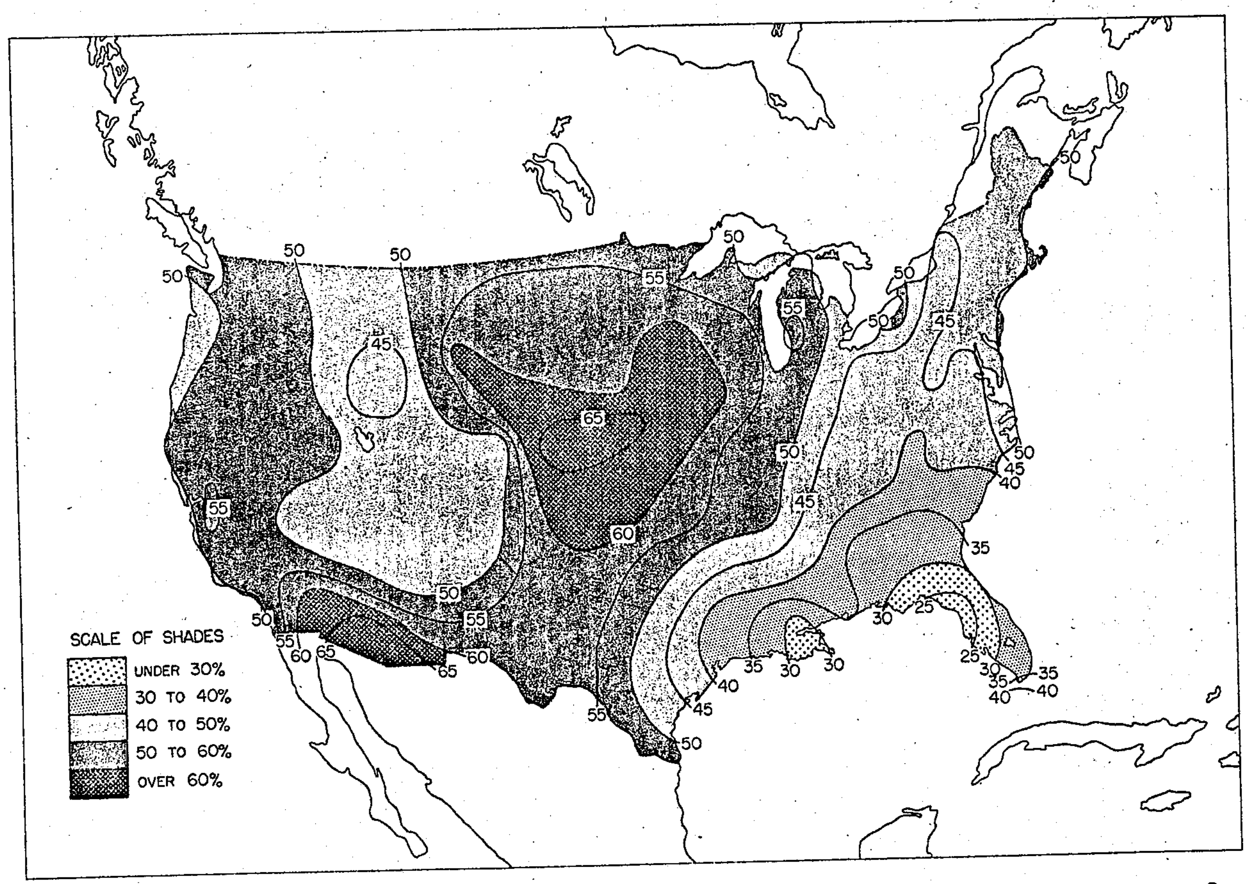

FIG. 27.—Percentage of total precipitation occurring at night (8:00 P.M. to 8:00 A.M., 75th Meridian Time), April to September inclusive. (After Kincer, Atlas of American Agriculture, Part II [Washington, D.C.: United States Department of Agriculture].)

August inclusive for 1934 and 1935.[5] He points out that the westerly component of the winds at Omaha from 750 m to 2,000 m is less during the period from 4:00 A.M. to 4:00 P.M. than during the period from 4:00 P.M. to 4:00 A.M. The maximum diurnal variation at Omaha occurs at an elevation of about 1 km above sea-level, amounting to about 4 m per second. A greater diurnal variation occurs over Oklahoma City. Wagner ascribes the diurnal wind variation at Omaha as being due to three factors: (a) the dry, relatively low region in the Southwest (Arizona and surrounding regions) having large diurnal temperature oscillations at the surface; (b) the temperature oscillations over the plateau and mountain regions of the West; and (c) the "breathing" of the continent itself with the relatively large continental temperature oscillations as compared with the surrounding oceans.

Comparison of Weather Bureau wind resultant data for summer months at Omaha for the 1,500-m and 3,000-m levels shows that the 1,500-m winds have a greater mean velocity at night than during the day, while the winds at 3,000 m show a much smaller variation, with the greater mean velocity occurring more frequently during the day. At the surface the winds are stronger during the day than at night. Downward transport of momentum during the day when steeper lapse rates occur in the air layers next to the ground may help to explain both the increase in the surface wind and the decrease in the 1,500-m wind during the day.

Assuming similar horizontal temperature gradients both night and day in the vicinity of Omaha and horizontal flow at the 1,500-m and 3,000-m levels, advection at the 1,500-m level would be greater at night than during the day, while at the 3,000-m level advection would be slightly less at night than during the day. Since the orientation of isotherms and isobars would give warm-air advection at both levels, greater steepening of the lapse rate would occur between 1,500 and 3,000 m at night, when greater similarity both night and day of horizontal temperature gradients is justified by Weather Bureau monthly radiosonde summary data for summer months for the region in which a nocturnal maximum of thunderstorms occurs. Data from the day and night radiosonde summaries plotted on constant-level charts show similar temperature gradients but with possibly slightly greater gradients occurring at night.

The areas swept out by clockwise turning of the wind vectors on

---

5A. Wagner, "Über die Tageswinde in der freien Atmosphäre," Beiträge zur Physik der freien Atmosphäre, XXV (1939), 146.

the hodographs have a maximum in the midwestern nocturnal maximum thunderstorm region, as in data previously discussed. Wind-resultant data have been computed to supplement the Weather Bureau wind resultants for levels up to 3,000 m for the four daily pilot-balloon observations for the month of August, 1938. Hodographs of the wind-resultant data are given in Figures 28a and 28b. It is now possible to check the diurnal variation of the areas swept out by clockwise turning of the wind vectors on hodographs. These data are given in Figure 29, A, B, C, and D. The 2300 and 0500 periods show greater degrees of this effect than the daytime periods, 1100 and 1700. The data suggest that greater warm-air advection for the layer from the surface to 3,000 m occurs at night in this region. Calculation from the mean hodographs for Omaha shown in Figure 28a shows that the advective temperature change for the 1100 soundings is 1.9° C per 12 hours, for the 1700 soundings it is 1.4° C per 12 hours, for the 2300 soundings it is 3.8° C per 12 hours, and for the 0500 soundings it is 3.5° C per 12 hours.[6] These pilot-balloon data at Omaha agree with the data of A. Wagner for the periods for which he had data.[7] He lacked data for the 2300 period.

iii) Duration of nocturnal thunderstorms.—Since a marked occurrence of the advective effect is noted at both the 2300 and the 0500 periods, the effect appears to be rather persistent during the night hours. Continued or persistent warm-air advection in lower layers could contribute instability for a long period of time and help to explain the longer average duration of nocturnal thunderstorms. An example of such persistent warm-air advection is found in data previously mentioned for July 9, 1941, at Omaha with marked advection of warmer air at both the July 8, 2300, and the July 9, 0500, pilot-balloon observations with a thunderstorm occurring from 0003 to 0435 on the 9th. Another example is that of thunderstorms at Omaha, the first beginning at August 20, 1941, 2327, and lasting until August 21, 0320; a second, from 0420 to 0453; and a third, from 0520 to 0803, with continued warm-air advection indicated by the August 20, 2300, and August 21, 0500, pilot-balloon data. The explanation of the nocturnal maximum occurrence of thunderstorms as due to warmer-air advection

---

[6] Cf., for the theory of these and other similar calculations, R. D. Fletcher, "Some Practical Relations Involving the Vertical Wind Shear," Bulletin of the American Meteorological Society, XXIII (1942), 361-65.

[7] Op. cit.

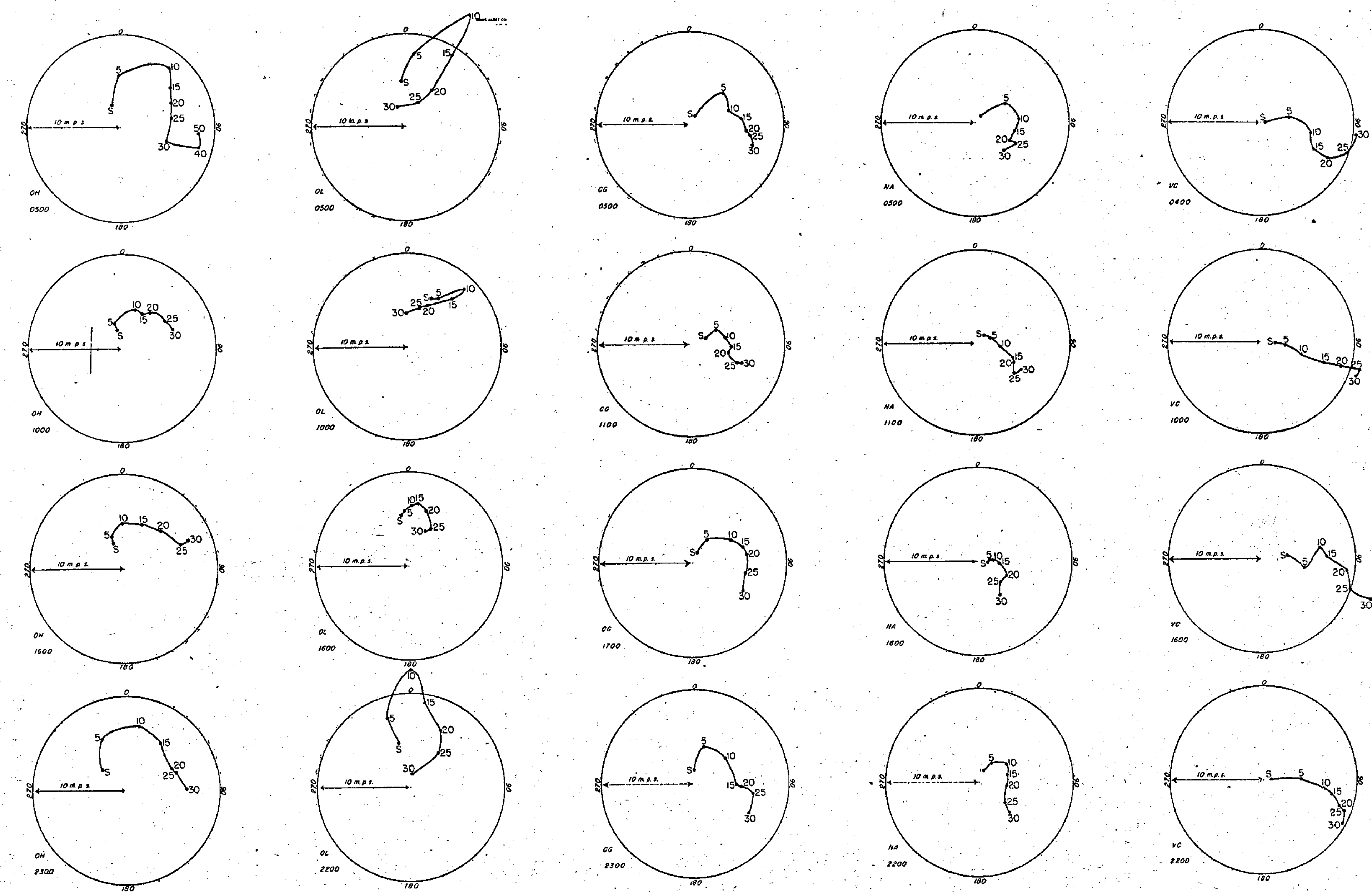

FIG. 28a.—Hodographs of wind-resultant data for Omaha, Oklahoma City, Chicago, Nashville, and Saulte Ste Marie.

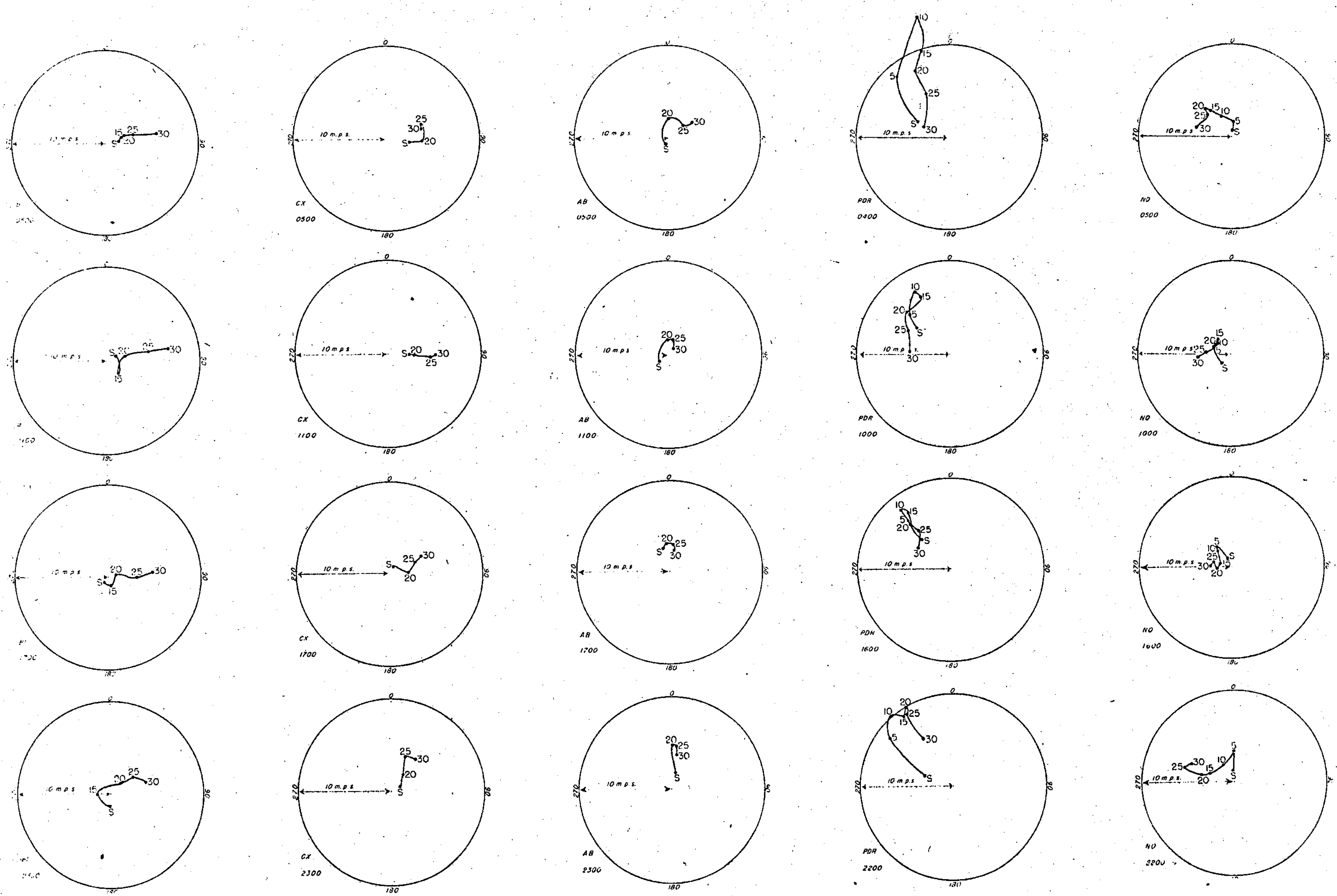

FIG. 28b.—Hodographs of wind resultant data for Billings, Cheyenne, Albuquerque, Del Rio, and New Orleans.

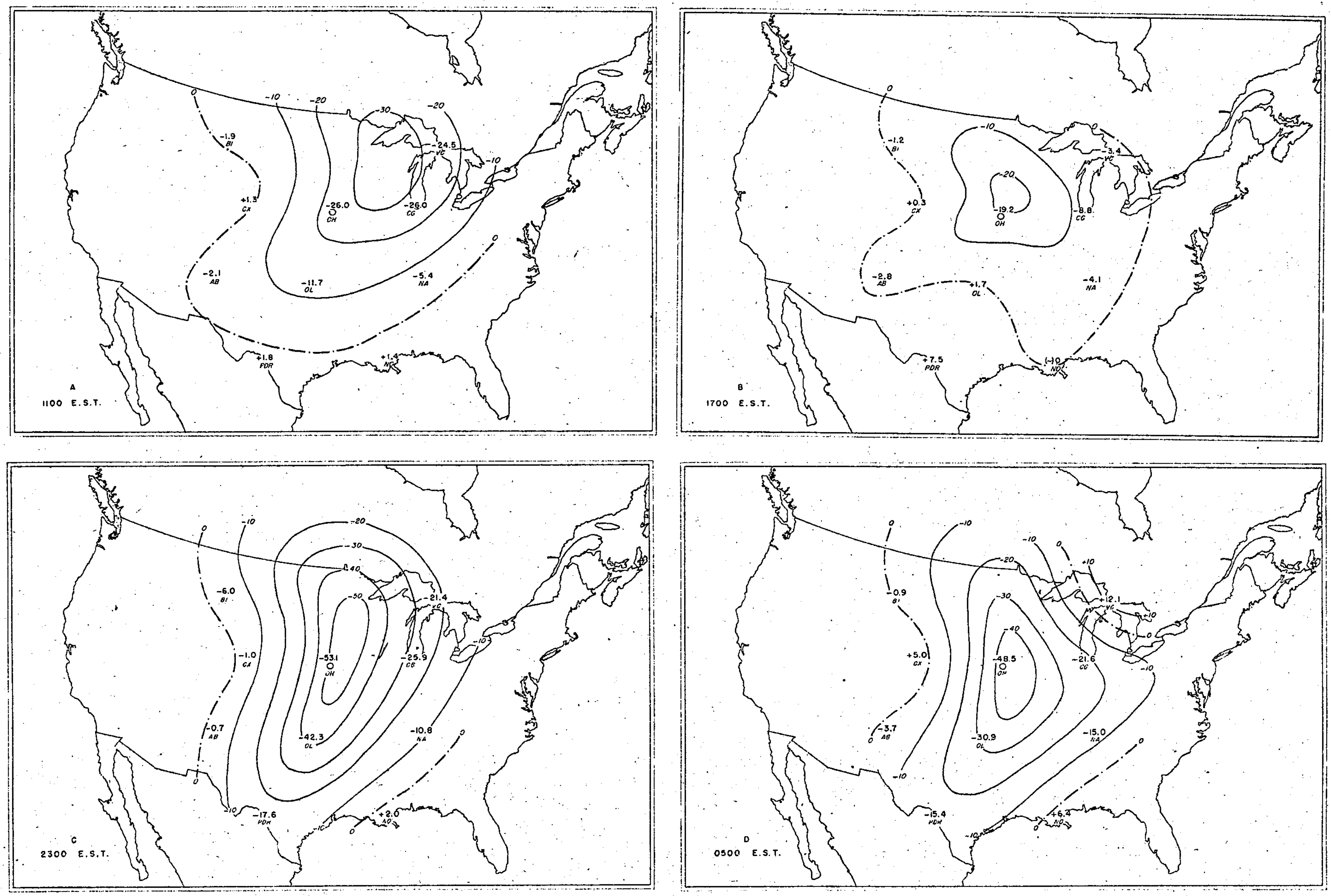

FIG. 29.—Isolines of areas swept out by mean wind vectors for the layer of air below 3 km, showing diurnal variation (based on hodographs of wind-resultant data shown in Fig 28a and 28b).

in the lower layers would seem consistent with the observation that nocturnal thunderstorms last longer than daytime thunderstorms, many of which may be due to surface heating of the layer next to the ground.

iv) Further evidence.—Pilots who fly in the Middle West have described nocturnal thunderstorms with many of them agreeing on the following observations:

a) Altocumulus clouds occur frequently during the 12-hour period preceding the occurrence of nocturnal thunderstorms.

b) The bases of nocturnal thunderstorms are usually high, existing many times above 10,000 feet.

c) Turbulence is frequently encountered below the base of the cumulonimbus.

For those nocturnal thunderstorms that occur due to advection of warmer air in the lower layers, these observations are quite consistent. (a) With such advection contributing to the formation of convection first in a layer between 2,000 to 8,000 feet above M.S.L., given continued convection and sufficient moisture, the production of cumuloform clouds at a high level is to be expected. If convection continues until any relatively stable layer at the top of the altocumulus clouds is overcome, convective action may contribute to the formation of a cumulonimbus cloud. (b) The average height of the convective condensation level for air parcels at the 1500-m level from radiosonde observations taken before the occurrence of nocturnal thunderstorms at Omaha in 1941 is near 10,000 feet above M.S.L. (c) Turbulence would be expected below the cloud down to the level at which the greatest warm-air advection is occurring.

CONCLUSION

The data presented in this report show that in the development of vertical instability the advection of warmer air in the lower layers of the atmosphere is a causal factor that is consistent with the nocturnal maximum occurrence of thunderstorms in the Middle West.

ACKNOWLEDGMENTS

The author wishes to acknowledge the helpful comments and criticisms by Professors Starr, Byers, and Rossby, of the University of Chicago, and also the generous co-operation of the Weather Bureau in providing data for this study.