CHAPTER 4

FORECASTING WEATHER ELEMENTS

Cloudiness, precipitation, and temperature are among the most important elements of any weather forecast. Cloudiness and precipitation may be subdivided into two main categories: occurrences associated with frontal activity and occurrences in air masses not associated with fronts.

There are many factors that influence the daily heating and cooling of the atmosphere. Some of these factors are cloudiness, humidity, nature of the undedying surface, surface winds, latitude, and the vertical temperature lapse rate. Cloudiness is quite obvious in its influence on heat gain at the earth's surface during daylight hours, and also heat loss due to radiational cooling at night.

This chapter discusses the forecasting of middle and upper cloudiness, precipitation, and local temperature. The forecasting of convective clouds and associated precipitation, along with fog and stratus is covered in chapter 5 of this manual.

CONDENSATION AND PRECIPITATION PRODUCING PROCESSES

LEARNING OBJECTIVES: Evaluate the processes necessary for condensation and precipitation to occur.

We will begin our discussion by identifying the conditions that must be present for condensation and precipitation to take place.

CONDENSATION PRODUCING PROCESSES

The temperature of a parcel of air must be lowered to its dewpoint for condensation to occur. Condensation depends upon two variables—the amount of cooling and the moisture content of the parcel. Two conditions must be met for condensation to occur; first, the air must be at or near saturation, and second, hydroscopic nuclei must be present. The first condition may be brought about by evaporation of additional moisture into the air, or by the cooling of the air to its dewpoint temperature. The first process (evaporation of moisture into the air) can occur-only if the vapor pressure of the air is less than the vapor pressure of the moisture source. The second condition (cooling) is the principal condensation producer.

Nonadiabatic cooling processes (radiation and conduction associated with advection) primarily result in fog, light drizzle, dew or frost.

The most effective cooling process in the atmosphere is adiabatic lifting of air. It is the only process capable of producing precipitation in appreciable amounts. It is also a principal producer of clouds, fog, and drizzle. The meteorological processes that result in vertical motion of air are discussed in the following texts. None of the cooling processes are capable of producing condensation by themselves; moisture in the form of water vapor must be present.

PRECIPITATION PRODUCING PROCESSES

Precipitation occurs when the products of condensation and/or sublimation coalesce to form hydrometers that are too heavy to be supported by the upward motion of the air. A large and continuously replenished supply of water droplets, ice crystals, or both is necessary if appreciable amounts of precipitation are to occur.

Adiabatic lifting of air is accomplished by orographic lifting, frontal lifting, or vertical stretching (or horizontal convergence). All of these mechanisms are the indirect results of horizontal motion of air.

Orographic Lifting

Orographic lifting is the most effective and intensive of all cooling processes. Horizontal motion is converted into vertical motion in proportion to the slope of the inclined surface. Comparatively flat terrain can have a slope of as much as 1 mile in 20 miles.

The greatest extremes in rainfall amount and intensity occur at mountain stations. For this reason, it is very important that the forecaster be aware of this potential situation.

Frontal Lifting

Frontal lifting is the term applied to the process represented on a front when the inclined surface represents the boundary between two air masses of different densities. In this case, however, the slope ranges from 1/20 to 1/100 or even less. The steeper the front, the more adverse and intense its effects, other factors being equal. These effects were discussed in detail in the *AG2 TRAMAN,* volume 1.

Vertical Stretching

Since it is primarily from properties of the horizontal wind field that vertical stretching is detectable, it is more properly called *convergence.* This term will be used hereafter.

The examples of convergence and divergence, explained in the foregoing, are definite and clear cut, associated as they are with the centers of closed flow patterns. Less easily detected types of convergence and divergence are associated with curved, wave-shaped, or straight flow patterns, where the air is moving in the same general direction. Variations in convergence and divergence are indicated in figures 4-1, 4-2, 4-3, and 4-4 by means of the following key:

cc marked convergence

c convergence

o little or no convergence or divergence

d divergence

dd marked divergence

Arrows have the following interpretations:

——→ wind direction and speed

——→ movement of center

——→ path of air current

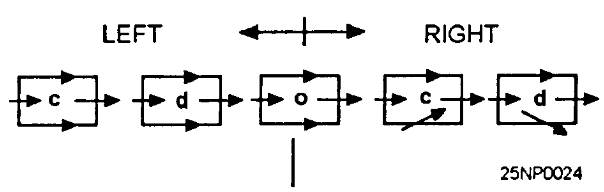

Figure 4-1.-Longitudinal and lateral convergence and divergence.

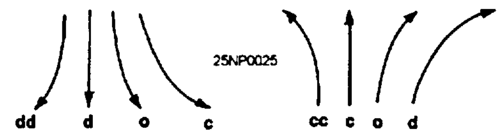

Figure 4-2.-Convergence and divergence in meridional flow.

The left side of figure 4-1 illustrates longitudinal convergence and divergence; the right side illustrates lateral convergence and divergence. Many more complicated situations can be analyzed by separation into these components.

It can be shown mathematically and verified synoptically that a fairly deep layer of air moving with a north-south component has associated convergence or divergence, depending on its path of movement. In figure 4-2 the arrows indicate paths of meridional flow in the Northern Hemisphere. In general, equatorward flow is divergent unless turning cyclonically, and poleward flow is convergent unless turning anticyclonically.

The four diagrams of figure 4-3 represent the approximate distribution of convergence and divergence in Northern Hemispheric cyclones and anticyclones. For moving centers, the greatest convergence or divergence occurs on or near the axis along which the system is moving. The diagrams of figure 4-3 show eastward movement, but they apply regardless of the direction of movement of the center.

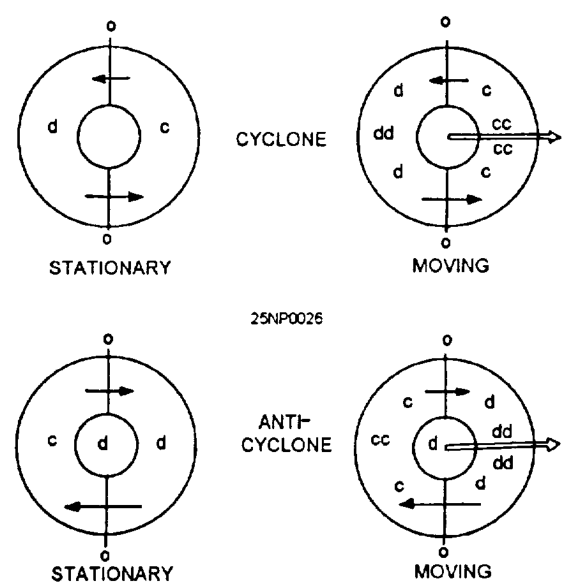

Figure 4-3.-Convergence and divergence in lows and highs.

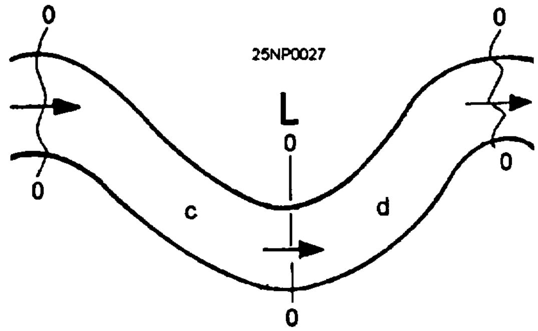

Figure 4-4.-Convergence and divergence in waves.

Convergence and divergence are not quite so easily identified in wave-shaped flow patterns because the wave speed of movement is often the factor that determines the distribution. The most common distribution for waves moving toward the east is illustrated in figure 4-4. There is relatively little divergence at the trough and ridge lines, with convergence to the west and divergence to the east of the trough lines.

This chapter devotes more time to a discussion of convergence because it is the most difficult characteristic to assess. Its extent ranges from the extremely local convergence of thunderstorm cells and tornadoes to the large-scale convergence of the broad and deep currents of poleward- and equatorward-moving air masses.

The amount, type, and intensity of the weather phenomena resulting from any of the lifting processes described in this chapter depend on the stability or convective stability of the air being lifted.

All of the lifting mechanisms (orographic, frontal, vertical stretching) can occur in any particular weather situation. Any combination, or all three, are possible, and even probable. For instance, an occluded cyclone of maritime origin moving onto a mountainous west coast of a continent could easily have associated with it warm frontal lifting, cold frontal lifting, orographic lifting, lateral convergence, and convergence in the southerly flow. All fronts have a degree of convergence associated with them.

WEATHER DISSIPATION PROCESSES

LEARNING OBJECTIVES: Identify processes leading to the dissipation of weather.

Each of the processes described in the preceding text has its counterpart among the condensation-preventing or weather-dissipating processes. Downslope flow on the lee side of orographic barriers results in adiabatic warming. If the air mass above and in advance of a frontal surface is moving with a relative component away from the front, downslope motion with adiabatic warming will occur. Divergence of air from an area must be compensated for by subsiding air above the layer, which is warmed adiabatically. These mechanisms have the common effect of increasing the temperature of the air, thus preventing condensation.

Likewise, these processes occur in combination with one another, and they may also occur in combination with the condensation-producing processes. This may lead to situations that require careful analysis. For instance, a current of air moving equatorward on a straight or anticyclonically curved path (divergence indicated) encounters an orographic barrier; if the slope of this orographic barrier is sufficiently steep or the air is sufficiently moist, precipitation will occur in spite of divergence and subsidence associated with the flow pattern. The dry, sometimes even cloudless, cold front that moves rapidly from west to east in winter is an example of upper level, downslope motion, which prevents the air being lifted by the front from reaching the condensation level.

The precipitation process itself opposes the mechanism that produces it, both by contributing the latent heat of vaporization and by exhausting the supply of water vapor.

FORECASTING FRONTAL CLOUDS AND WEATHER

LEARNING OBJECTIVES: Evaluate surface and upper level synoptic data in the analysis of frontal clouds and weather.

Cloud and weather regimes most difficult to forecast are those associated with cyclogenesis. It is well known that falling pressure, precipitation, and an expanding shield of middle clouds indicate that the cyclogenetic process is occurring and, by following these indications, successful forecasts can often be made for 6 to 48 hours in advance. Most of the winter precipitation of the lowlands in the middle latitudes is chiefly cyclonic or frontal in origin, though convection is involved when the displaced air mass is unstable.

Cyclones are important generators of precipitation in the Tropics as well as in midlatitudes.

Factors to be considered in arriving at an accurate forecast are listed below; these factors are not listed in any order of importance:

- The source region of the parent air mass.
- Nature of the underlying surface.
- The type and slope of the front(s).
- Wind and contour patterns aloft.
- Past speed and direction of movement of the low or front(s).
- Familiarization with the *normal* weather patterns.

As pointed out earlier, a thorough understanding of the physical processes by which precipitation develops and spreads is essential to an accurate forecast.

FRONTAL AND OROGRAPHIC CLOUDINESS AND PRECIPITATION

There are unique cloud and precipitation features and characteristics associated with the cold and warm fronts, as well as orographic barriers. The following text discusses these features and characteristics.

Cold Front

You will find it helpful to use constant pressure charts in conjunction with the surface synoptic situation in forecasting cold frontal cloudiness and precipitation. When the contours at the 700-hPa level are perpendicular to the surface cold front, the band of weather associated with the front is narrow. This situation occurs with a fast-moving front. If the front is slow moving, the weather and precipitation will extend as far to the rear of the front as the winds at the 700-hPa level are parallel to the front. In both of the above cases, the flow at 700 hPa also indicates the slope of the front. Since the front at the 700-hPa level lies near the trough line, it is apparent that when the flow at 700 hPa is perpendicular to the surface front, the 700-hPa trough is very nearly above the surface trough; hence, the slope of the front is very steep. When the 700-hPa flow is parallel to the surface front, the 700-hPa trough lies to the rear of the surface front and beyond the region in which the flow continues parallel to the front. Consequently, the frontal slope is more gradual, and lifting is continuing between the surface and the 700-hPa level at some distance to the rear of the surface front.

Another factor that contributes to the distribution of cloudiness and precipitation is the curvature of the flow aloft above the front. Cyclonic flow is associated with horizontal convergence, and anticyclonic flow is associated with horizontal divergence.

Very little weather is associated with a cold front if the mean isotherms are perpendicular to the front. When the mean isotherms are parallel to the front, weather will occur with the front. This principle is associated with the contrast of the two air masses; hence, with the effectiveness of lifting.

Satellite imagery provides a representative picture of the cloud structure of frontal systems. Active cold fronts appear as continuous, well-developed cloud bands composed of low, middle, and high clouds. This is caused by the upper wind flow, which is parallel, or nearly parallel, to the frontal zone (fig. 4-5).

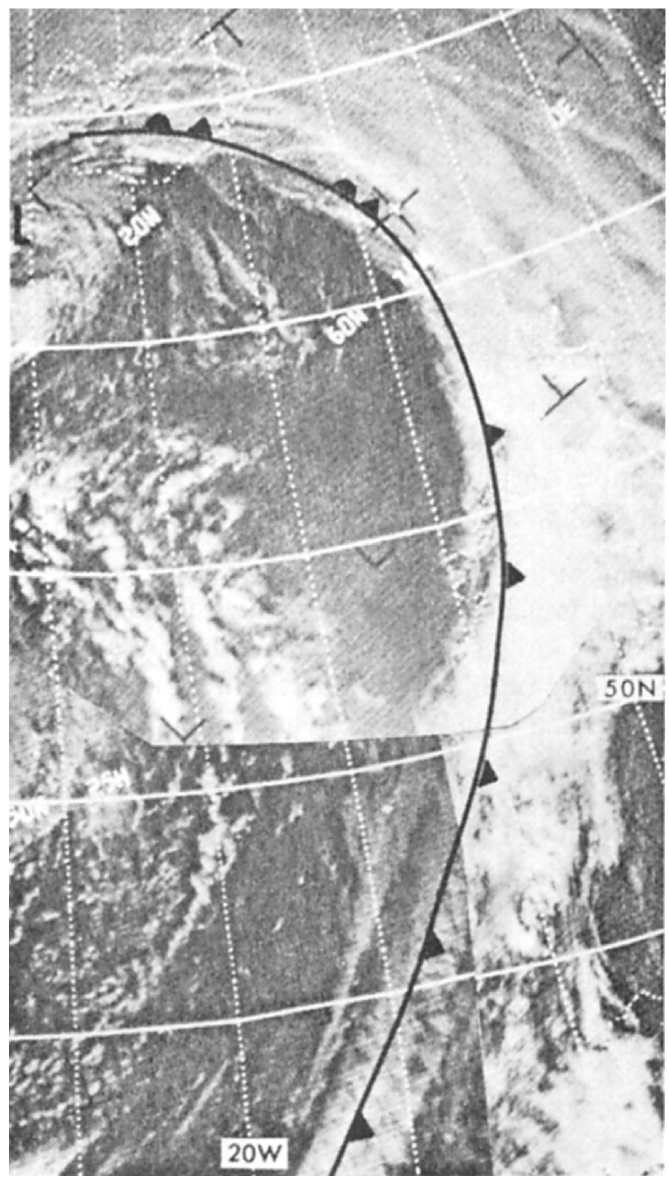

Figure 4-5.—An active cold front.

The perpendicular component of the upper winds associated with the inactive cold front causes the cloud bands to appear as narrow, fragmented, or discontinuous. The band of clouds is comprised mainly of low-level cumulus and stratiform clouds, but some cirriform may be present. Occasionally, inactive cold fronts over water will have the same appearance as active fronts over land, while overland they may have few or no clouds present. Figure 4-6 depicts the fragmented clouds associated with an inactive cold front in the lower portion, while a more active cold front cloud presentation is shown in the upper portion.

Warm Front

As with cold fronts, the use of constant pressure charts in conjunction with the surface synoptic situation is helpful in forecasting warm-frontal cloudiness and precipitation.

Cloudiness and precipitation occur where the 700-hPa flow across the warm front is from the warm air to the cold air, and is moving in a cyclonic path or in a straight line. This implies convergence associated with the cyclonic curvature. Warm fronts are accompanied by no weather and few clouds if the 700-hPa flow above them is anticyclonic. This is due to horizontal divergence associated with anticyclonic curvature.

The 700-hPa ridge line ahead of a warm front may be considered the forward limit of the prewarm frontal cloudiness. The sharper the ridge line, the more accurate the rule.

When the slope of the warm front is gentle near the surface position, and is steep several hundred miles to the north, the area of precipitation is situated in the region where the slope is steep. There may be no precipitation just ahead of the surface frontal position.

Warm fronts are difficult to locate on satellite imagery. An active warm front maybe associated with a well organized cloud band, but the frontal zone is difficult to locate. An active warm front maybe placed somewhere under the bulge of clouds that are associated with the peak of the warm sector of a frontal system. The clouds are a combination of stratiform and cumuliform beneath a cirriform covering. See figure 4-7.

You must remember that no one condition represents what could be called typical, as each front presents a different situation with respect to the air masses involved. Therefore, each front must be treated as a separate case, by using present indications, geographical location, stability of the air masses, moisture content, and intensity of the front to determine its precipitation characteristics.

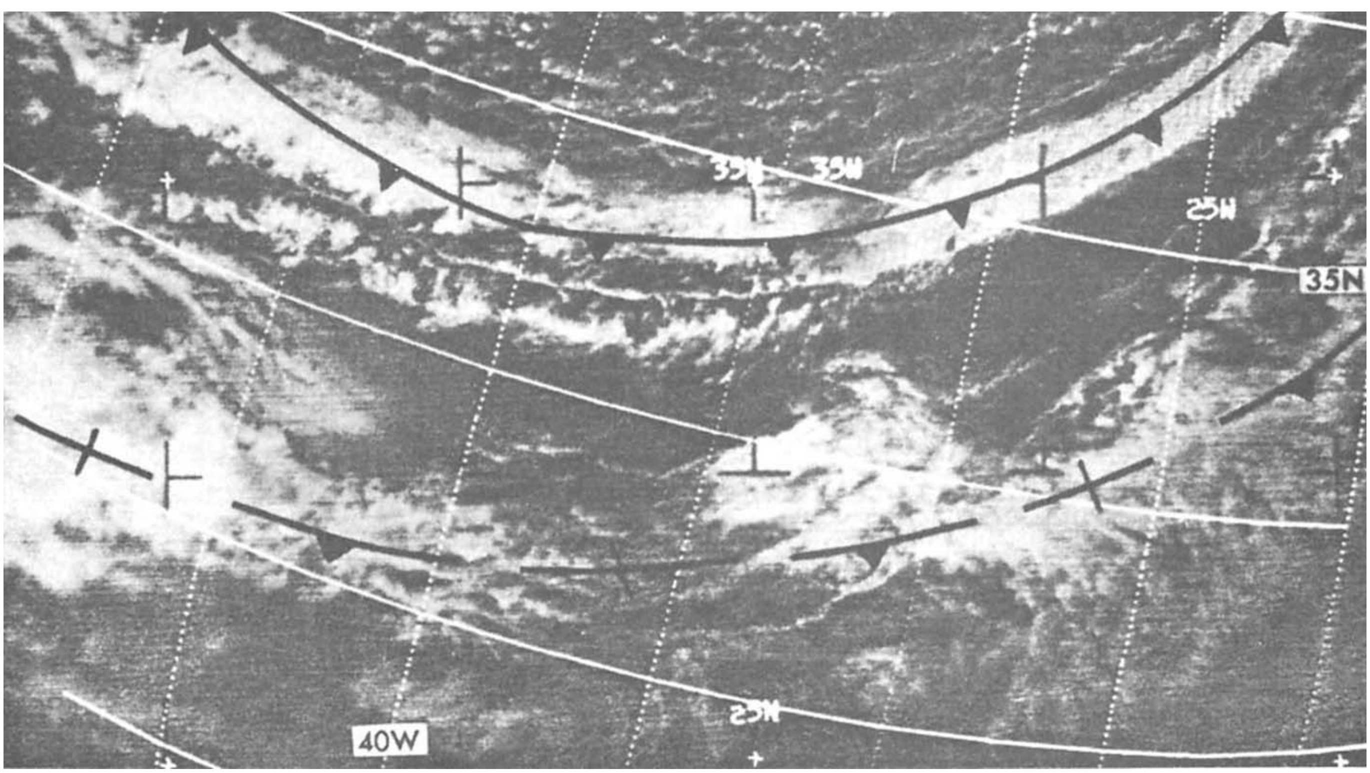

Figure 4-6.-Inactive and active cold front satellite imagery.

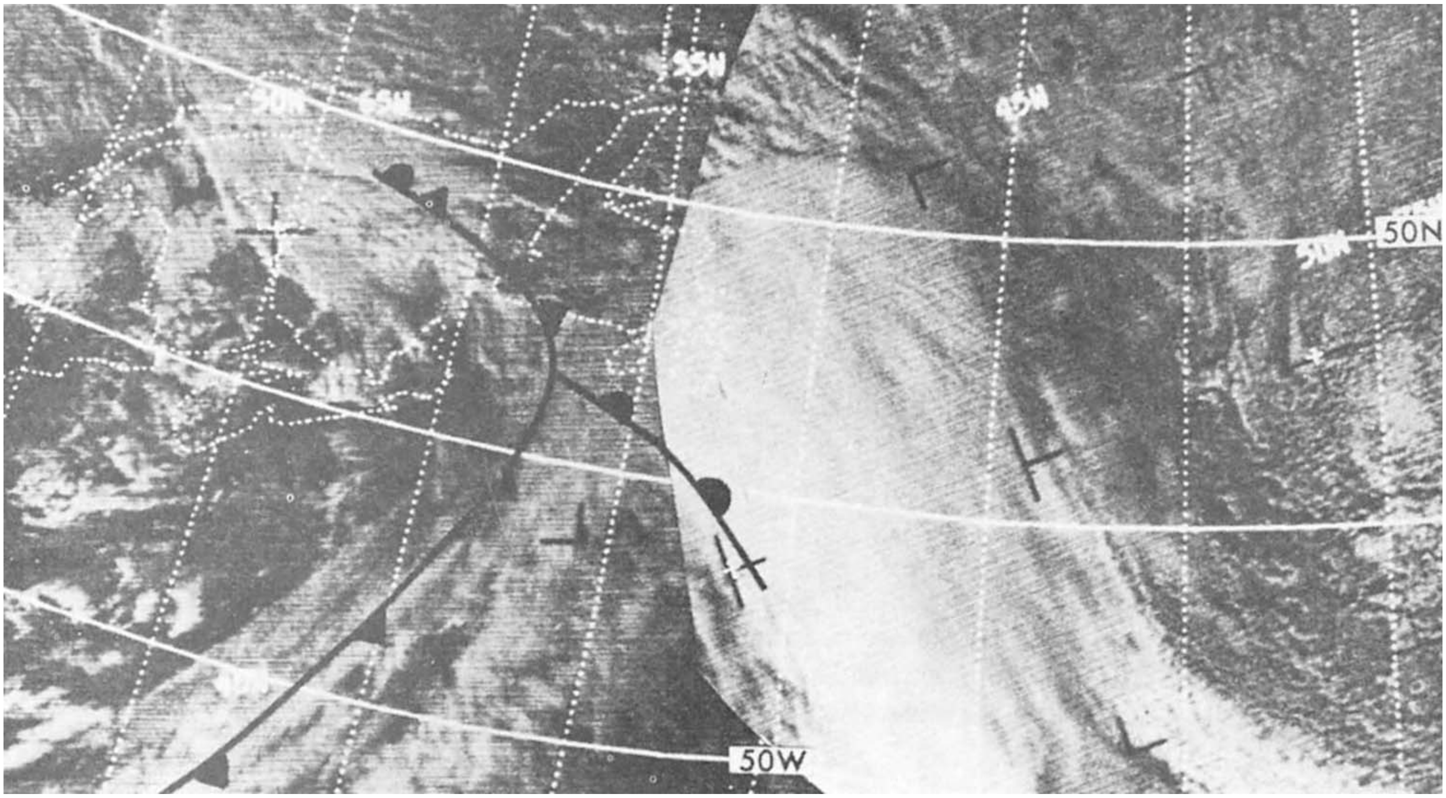

Figure 4-7.-Warm front satellite cloud imagery.

Orographic Barriers

In general, an orographic barrier increases the extent and duration of cloudiness and precipitation on the windward side, and decreases it on the leeward side.

AIR MASS CLOUDINESS AND PRECIPITATION

If an air mass is lifted over an orographic barrier, and the lifting is sufficient for the air to reach its lifting condensation level, cloudiness of the convective type occurs. If the air is convectively unstable and has sufficient moisture, showers or thunderstorms occur. The preceding situations occur on the windward side of the barrier.

Curvature (path of movement) of the flow aloft also affects the occurrence of cloudiness and precipitation. In a cool air mass, showers and cumulus and stratocumulus clouds are found in those portions of the air mass that are moving in a cyclonically curved path. In a warm air mass, cloudiness and precipitation will be abundant under a current turning cyclonically or moving in a straight line. Clear skies occur where a current of air is moving from the north in a straight line or in an anticyclonically curved path. Also, clear skies are observed in a current of air moving from the south if it is turning sharply anticyclonically. Elongated V-shaped troughs aloft have cloudiness and precipitation in the southerly current in advance of the troughs, with clearing at and behind the trough. These rules also apply in situations where this type of low is associated with frontal situations.

Cellular cloud patterns (open or closed), as shown by satellite imagery, will aid the forecaster in identifying regions of cold air advection, areas of cyclonic, anticyclonic, and divergent wind flow.

Open Cellular Cloud Patterns

Open cellular cloud patterns are most commonly found to the rear of cold fronts in cold, unstable air. These patterns are made up of many individual cumuliform cells. The cells are composed of cloudless, or less cloudy, centers surrounded by cloud walls with a predominant ring or U-shape. In the polar air mass, the open cellular patterns that form in the deep, cold air are predominately cumulus congestus and cumulonimbus. The open cells that form in the subtropical high are mainly stratocumulus, cumulus, or cumulus congestus clusters. For open cells to form in a polar high, there must be moderate to intense heating of the air mass from below.

Figure 4-8.-Open cells on satellite Imagery.

When this polar air mass moves out over the water, the moist layer is shallow and capped by a subsidence inversion near the coast. Further downstream the vertical extent of the moist layer and the height of the clouds increases due to air mass modifications by the underlying surface. In figure 4-8, the open cells behind a polar front over the North Atlantic indicate cold air advection and cyclonic curvature of the low-level wind flow. Vertical thickness of the cumulus at A is small, but increases eastward toward B.

Figure 4-9 shows a large area of the subtropical high west of Peru covered with open cells. These are not associated with low-level cyclonic flow or steady cold air advection.

Closed Cellular Cloud Patterns

Closed cellular cloud patterns are characterized by approximately polygonal cloud-covered areas bounded by clear or less cloudy walls. Atmospheric conditions necessary for the formation of closed cells are weak convective mixing in the lower levels, with a cap to this mixing aloft. The convective mixing is the result of surface heating of the air or radiational cooling of the cloud tops. This type of convection is not as intense as that associated with open cells. The cap to the instability associated with closed cellular cloud patterns is in the form of a subsidence inversion in both polar and subtropical situations. Closed cellular cloud patterns are made up of stratocumulus elements in both the polar and subtropical air masses. In addition to the stratocumulus elements, trade-wind cumulus may also be present with the subtropical highs. When associated with the subtropical highs, closed cellular clouds are located in the eastern sections of the high-pressure area. Closed cells are associated with limited low-level instability below the subsidence inversion. Extensive

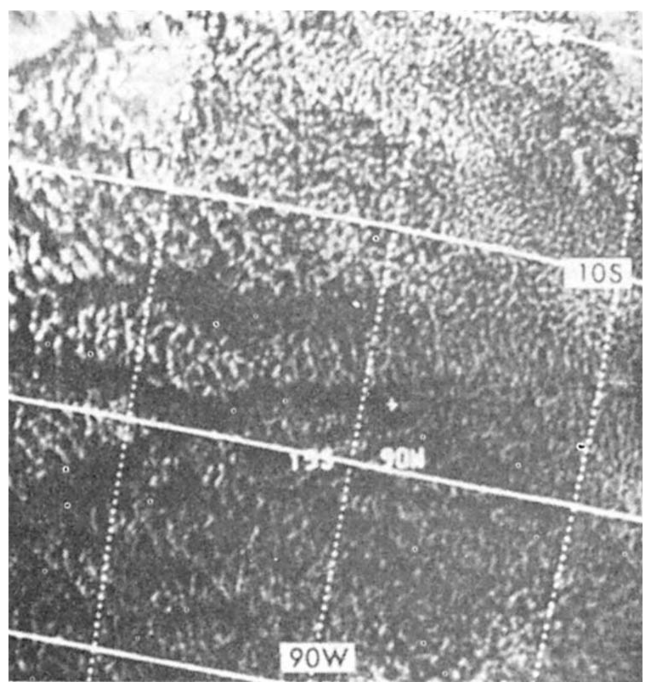

Figure 4-9.-Open cells in the subtropical high.

vertical convective activity is not likely. Figure 4-10 shows closed cells in the southeastern portion of a polar high near A.

Figure 4-11 shows closed cells in the eastern portion of a subtropical high in the South Pacific. West of A, the closed cells are composed of stratocumulus with some clear walls, and east of the walls, the cells are composed of thinner clouds.

The practical training publications, *Satellite Imagery Interpretation in Synoptic and Mesoscale Meteorology,* NAVEDTRA 40950, and *Tropical Clouds and Cloud Systems Observed in Satellite Imagery,* Volume 1, NAVEDTRA 40970, offer further information on the subject of satellite interpretation.

VERTICAL MOTION AND WEATHER

Upward vertical motion (convection) is associated with increasing cloudiness and precipitation, and downward vertical motion (subsidence) with improving weather.

Vertical motion analyses and prognostic charts are currently transmitted by the NWS and FNMOC. The values are computed and are on the charts. charts. Plus values represent upward vertical motion (convection), and minus values represent downward vertical motion (subsidence).

VORTICITY AND PRECIPITATION

Vorticity was discussed in chapter 1 of this manual, as well as the *AG2* TRAMAN, volume 1. We have seen that relative vorticity is due to the effects of both curvature and shear. Studies have led to the following rules:

- Cloudiness and precipitation should prevail in regions where the relative vorticity decreases downstream.
- Fair weather should prevail where relative vorticity increases downstream.

The fact that both shear and curvature must be considered when relative vorticity changes are investigated results in a large number of possible combinations on upper air charts. When both terms are in agreement, we can confidently predict precipitation or fair weather. When the two are in conflict, a closer examination is required.

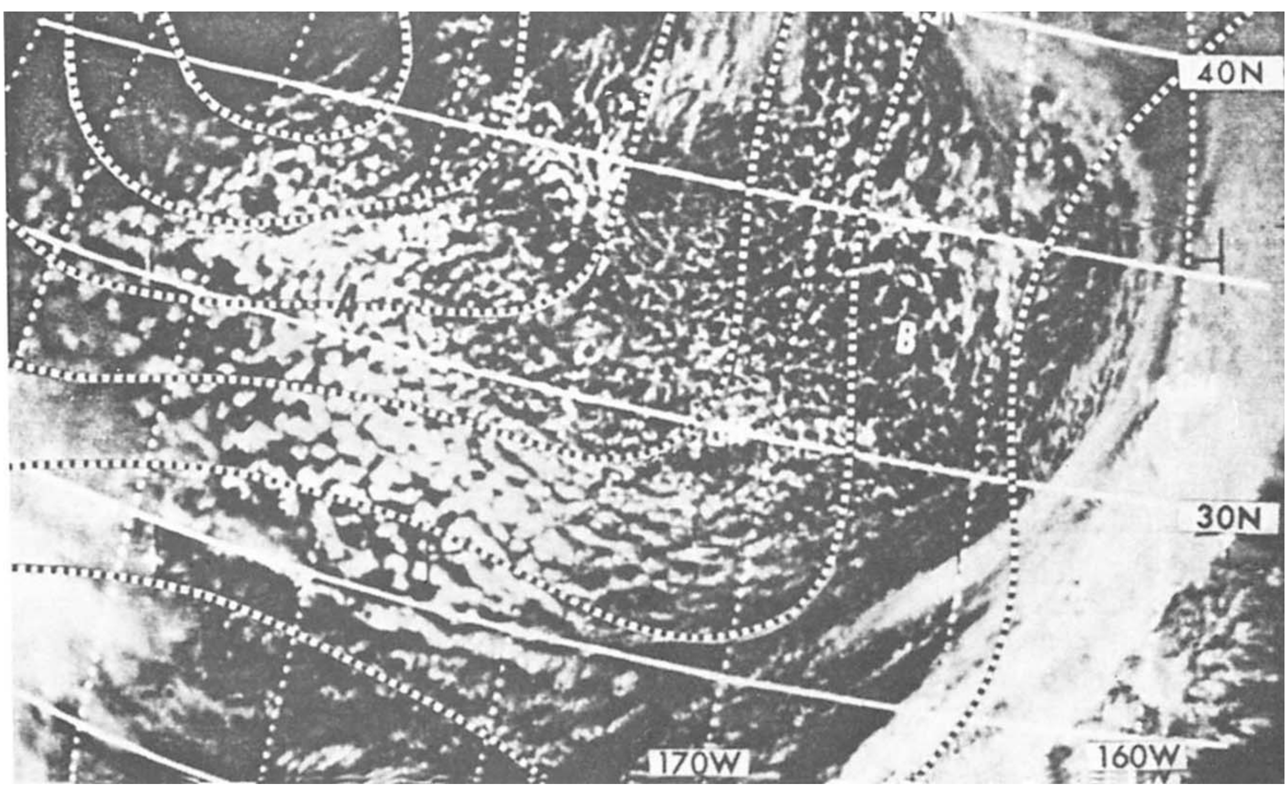

Figure 4-10.-Closed cells in polar high.

Figure 4-11.-Closed cells in a subtropical high.

MIDDLE CLOUDS IN RELATION TO THE JETSTREAM

Jets indicate as much individuality with respect to associated weather as do fronts. Because of the individuality of jetstreams, and also because of the individuality of each situation with respect to humidity distribution and lower level circulation patterns, statistically stated relationships become somewhat vague and are of little value in forecasting.

SHORT-RANGE EXTRAPOLATION

LEARNING OBJECTIVES: Compare short-range extrapolation techniques for the movement of frontal systems and associated weather.

The purpose of this section is to outline several methods that are particularly suited to preparing forecasts for periods of 6 hours or less. The techniques presented are based on extrapolation. Extrapolation is the estimating of the future value of some variable based on past values. Extrapolation is one of the most powerful short-range forecasting tools available to the forecaster.

NEPHANALYSIS

Nephanalysis may be defined as any form of analysis of the field of cloud cover and/or type. Cloud observations received in synoptic codes permit only a highly generalized description of the actual structure of the cloud systems.

Few forecasters make full use of the cloud reports plotted on their surface charts, and often, the first consideration in nephanalysis is to survey what cloud information is transmitted, and to make sure that everything pertinent is plotted. For very short-range forecast, the charts at 6-, 12-, and 24-hour intervals are apt to be insufficient for use of the extrapolation techniques explained in this chapter. Either nephanalysis or surface charts should then be plotted at the intermediate times from 3-hourly synoptic reports or even from hourly sequences. An integrated system of forecasting ceiling, visibility, cloud cover, and precipitation should be considered simultaneously, as these elements are physically dependent upon the same synoptic processes.

With present-day satellite capabilities, it is rare that a nephanalysis would be manually performed. Instead, the surface analysis and satellite imagery will be used together.

FRONTAL PRECIPITATION

For short-range forecasting, the question is often not whether there will be any precipitation, but when will it begin or end.

This problem is well suited to extrapolation methods. For short-range forecasting, the use of hourly nephanalyses often serve to “pickup” new precipitation areas forming upstream in sufficient time to alert a downstream area. Also, the thickening and lowering of middle cloud decks generally indicate where an outbreak of precipitation may soon occur.

Forecasting the Movement of Precipitation Areas by Isochrones

The areas of continuous, intermittent, and showery precipitation can be outlined on a large-scale 3-hourly or hourly synoptic chart in a manner similar to the customary shading of precipitation areas on ordinary synoptic surface weather maps. Different types of lines, shading, or symbols can distinguish the various types of precipitation. Isochrones of several hourly past positions of the lines of particular interest can then be added to the chart, and extrapolations for several hours made from them if reasonably regular past motions are in evidence. A separate isochrone chart (or acetate overlay) may be easier to use. Lines for the beginning of continuous precipitation are illustrated in figure 4-12. The isochrones for showery or intermittent precipitation usually give more uncertain and irregular patterns, which result in less satisfactory forecasts. When large-scale section surface weather maps are regularly drawn, it maybe sufficient and more convenient to make all precipitation area analyses and isochrones on these maps.

Forecasting the Movement of Precipitation by Using a Distance versus Time (x-t) Diagram

The idea of plotting observations taken at different times on a diagram that has horizontal or vertical distance in the atmosphere as one coordinate and time as the other has been used in various forms by forecasters for years. The time cross section that was discussed in the *AG2* TRAMAN, volume 1, unit 9, lesson 2. is a special case of this aid, where successive information at only one station is plotted.

LOWERING OF CEILING IN CONTINUOUS RAIN AREAS

One of the many obstacles the forecaster faces in preparing forecasts is the problem of determining "when" ceilings heights will lower in areas expecting rain. In the following paragraphs, we will discuss this dilemma.

Frontal Situations

The lowering of ceiling with continuous rain or snow in warm frontal and upper trough situations is a familiar problem to the forecaster in many regions. In very short-range forecasting, the question as to whether or not it will rain or snow, and when the rain or snow will begin, is not so often the critical question. Rather, the problem is more likely to be (assuming the rain/snow has started) how much will the ceiling lower in 1,2, and 3 hours, or will the ceiling go below a certain minimum in 3 hours. The visibility in these situations generally does not reach an operational minimum as soon as the ceiling. It has been shown that without sufficient convergence, advection, or turbulence, evaporation of rain into a layer does not lead to saturation, and causes no more than haze or light fog.

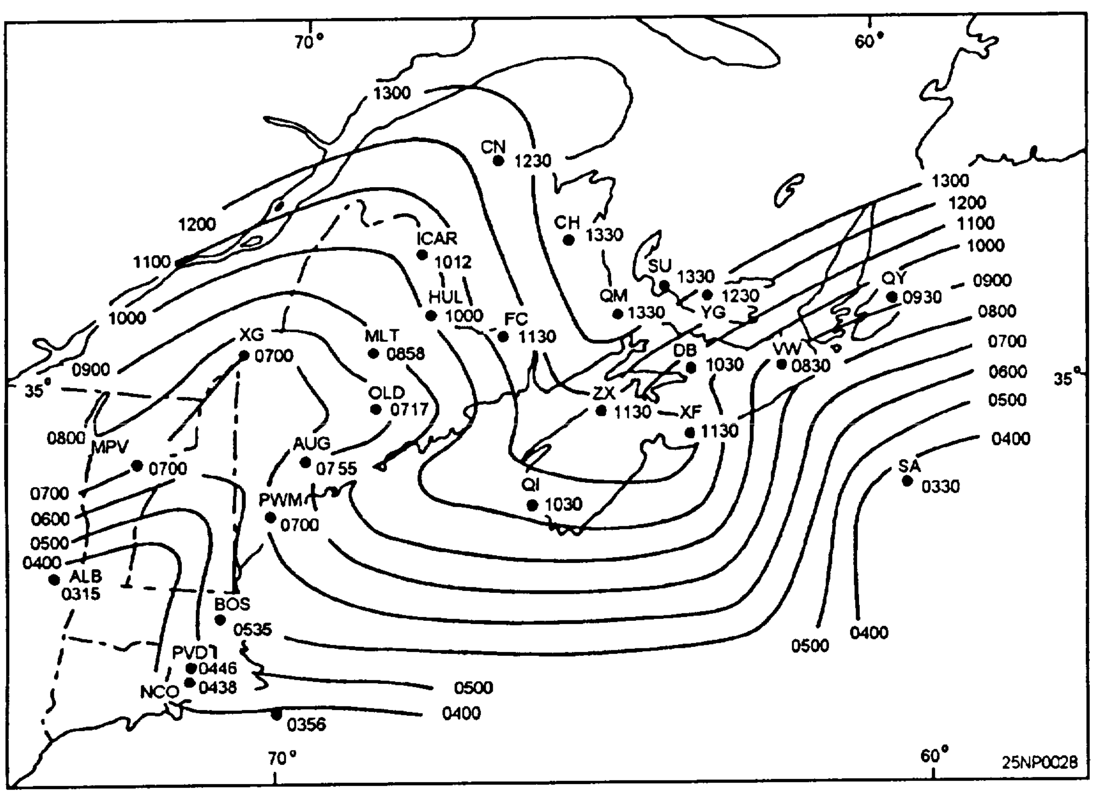

Figure 4-12.-Isochrones of beginning of precipitation, an early winter situation.

It is important to recognize the difference between the behavior of the actual cloud base height and the variation of the ceiling height, as defined in airway reports. The ceiling usually drops rapidly, especially during the first few hours after the rain or snow begins. However, if the rain or snow is continuous, the true base of the cloud layer descends gradually or steadily. The reason for this is that below the precipitation frontal cloud layer there are usually shallow layers in which the relative humidity is relatively high and which soon become saturated by the rain. The cloud base itself has small, random fluctuations in height superimposed on the general trend.

Time of Lowering of Ceiling

Forecasting the time when a given ceiling height will be reached during rain is a separate problem, Nomograms, tables, air trajectories, and the time air will become saturated can all be resolved into an objective technique tempered with empirical knowledge and subjective considerations. This forecast can be developed for your individual station.

Extrapolation of Ceiling Trend by Means of the x-t Diagram

The x-t diagram, as mentioned previously in this chapter, can be used to extrapolate the trend of the ceiling height in rain. The hourly observations should be plotted for stations near a line parallel to the probable movement of the general rain sea, originating at your terminal and directed toward the oncoming rain area. Ceiling-time curves for given ceiling heights may be drawn and extrapolated. There may be systematic geographical differences in the ceiling between stations due to local (topographic) influences. Such differences sometimes can be anticipated from climatological studies, or experience. In addition, there may be a diurnal ceiling fluctuation, which will become evident in the curve. Rapid and erratic up-and-down fluctuations also must be dealt with. In this case, a smoothing of the curves may be necessary before extrapolation can be made. A slightly less accurate forecast may result from this process.

In view of the previous discussion of the precipitation ceiling problem, it is not expected that mere extrapolation can be wholly satisfactory at a station when the ceiling lowers rapidly during the first hours of rain, as new cloud layers form beneath the front. However, by following the ceiling trend at surrounding stations, patterns of abrupt ceiling changes may be noted. These changes at nearby stations where rain started earlier may give a clue to a likely sequence at your terminal.

THE TREND CHART AS AN EXTRAPOLATION AID

The trend chart can be a valuable forecasting tool when it is used as a chronological portrayal of a group of related factors. It has the added advantage of helping the forecaster to become “current” when coming on duty. At a glance, the relieving duty forecaster is able to get the picture of what has been occurring. Also, the forecaster is able to see the progressive effect of the synoptic situation on the weather when the trend chart is used in conjunction with the current surface chart.

The format of a trend chart should be a function of what is desired; consequently, it may vary in form from situation to situation. It should, however, contain those elements that are predictive in nature.

The trend chart is a method for graphically portaying those factors that the forecasters generally attempt to store in their memory. Included in this trend chart is a list of key predictor stations. The forecaster uses the hourly and special reports from these stations as aids in making short forecasts for his/her station. Usually, the sequences from these predictor stations are scanned and committed to memory. The method is as follows:

1. Determine the direction from which the weather will be arriving; i.e., upstream.
2. Select a predictor station(s) upstream and watch for the onset of the critical factor; for example, rain.
3. Note the effect of this factor on ceiling and visibility at predictor station(s).
4. Extrapolate the approach of the factor to determine its onset at your station.
5. Consider the effect of the factor at predictor station(s) in forecasting its effect at your station.

The chief weakness of this procedure is its subjectivity. The forecaster is required to mentally evaluate all of the information available, both for their station and the predictor station(s).

A question posed, “How many trend charts do I need”? The answer depends on the synoptic situation. There are times when keeping a graphic record is unnecessary; and other times, the trend for the local station may suffice. The trend chart format, figure 4-13, is but one suggested way of portraying the weather record. Experimentation and improvisation are encouraged to find the best form for any particular location or problem.

TIME-LINER AS AN EXTRAPOLATION AID

In the preceding sections of this chapter, several methods have been described for “keeping track of the weather” on a short-term basis. Explanations of time-distance charts, isochrone aids, trend charts, etc., have been presented. It is usually not necessary to use all, or even most of these aids simultaneously. The aid described in this section is designed for use in combination with one or several of the methods previously described. Time-liners are especially useful for isochrone analysis and follow-on extrapolation.

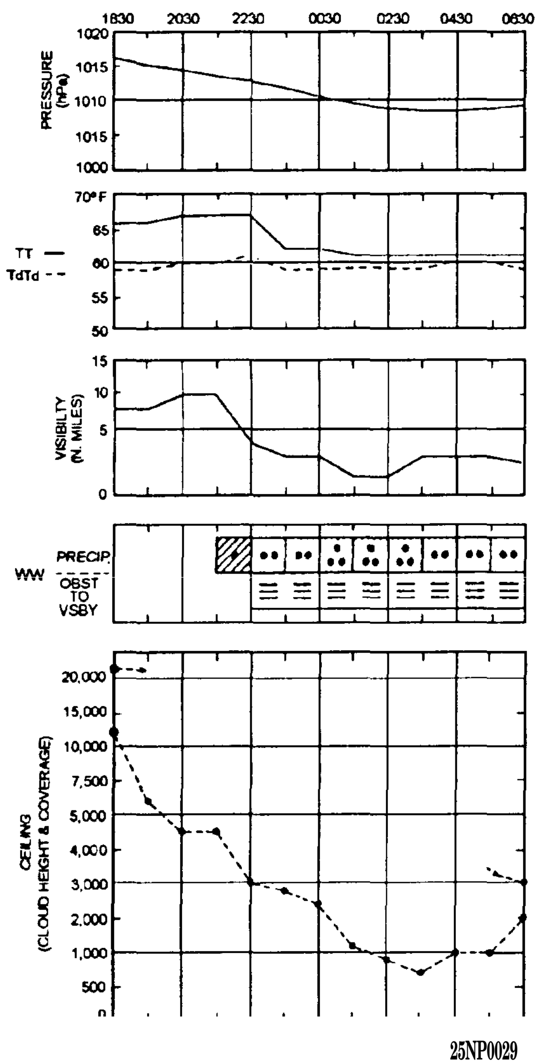

Figure 4-13.-Trend chart suggested format.

Inasmuch as a majority of incorrect short-range forecasts result from poor timing of weather already upstream, an aid, such as described below, may improve this timing.

Construction of the Time-Liner

The time-liner is simply a local area map that is covered with transparent plastic and constructed as follows:

1. Using a large-scale map of the local area, construct a series of concentric circles centered on your station, and equally spaced from 10 to 20 miles apart. This distance from the center to the outer circle depends on your location, but in most cases, 100 to 150 miles is sufficient.

2. Make small numbered or lettered station circles for stations located at varying distances and direction from your terminal. Stations likely to experience your future weather should be selected. In addition to the station circle indicators, significant topographical features, such as rivers or mountains, maybe indicated on the base diagram. (Aeronautical charts include these features.)

3. Cover and bind the map with transparent plastic.

Plotting and Analysis of the Time-Liner

By inspection of the latest surface chart, and other information, you can determine a quadrant, semicircle, or section of the diagram and the panmeters to be plotted. This should be comprised of stations in the direction from which the weather is approaching your station. Then, plot the hourly weather SPECIALS for those stations of interest. Make sure to plot the time of each special observation.

Overlay the circular diagram with another piece of transparent plastic, and construct isochrones of the parameter being forecast; for example, the time of arrival of the leading or trailing edge of a cloud or precipitation shield. The spacing between isochrones can then be extrapolated to construct “forecast isochrones” for predicting the time of arrival of occurrence of the parameter at your terminal. Refer to figure 4-14 for an example.

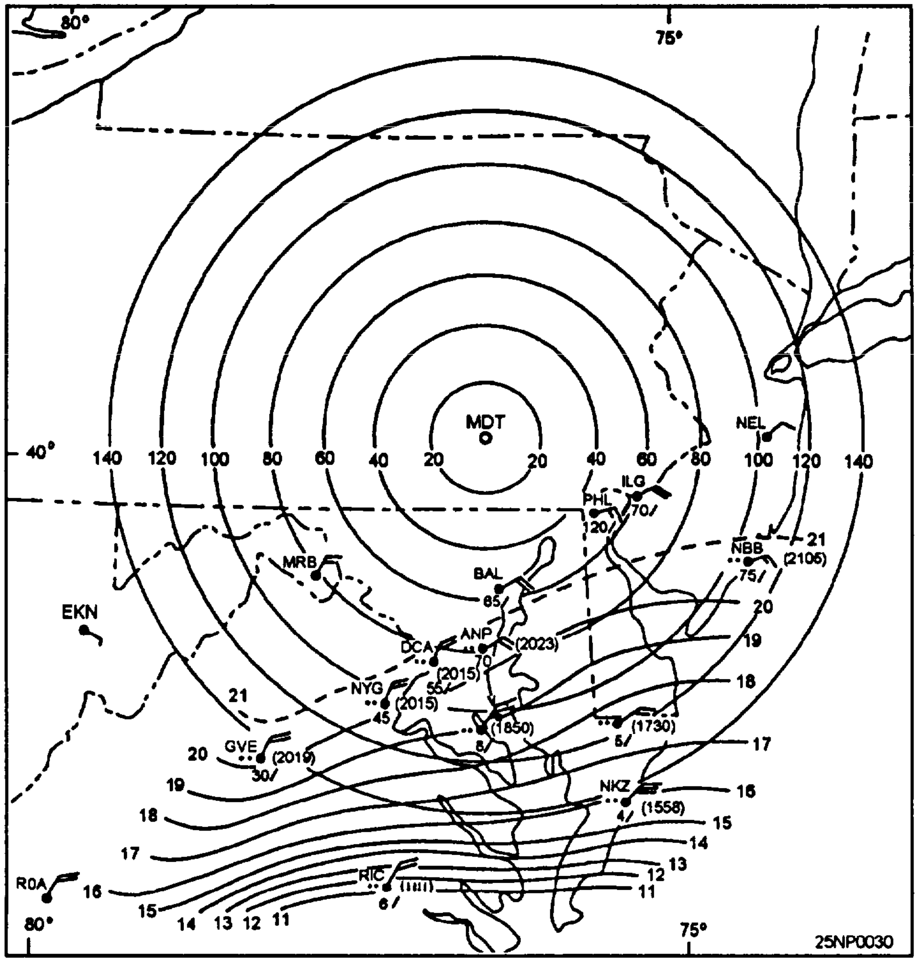

Figure 4-14.-Large-scale sample time-liner (Isochrones show advance of precipitation field).

Use of Doppler Radar in Cloud and Precipitation Forecasting

Doppler radar is very useful in determining weather phenomena approaching your station and estimating the probability of precipitation at your station, Refer to chapter 12 of this manual and the *Federal Meteorological Handbook No. 11,* Part B, for information on analysis of weather conditions and Doppler radar theory.

CLOUD ANALYSIS AND FORECASTING

LEARNING OBJECTIVES: Recognize upper air data and its value in forecasting. Recognize moisture features aloft and their significance to the forecaster.

Forecasters are frequently called upon to make forecasts of clouds over areas where synoptic observations are not readily available, or over areas where clouds above the lowest layer are obscured by a lower cloud deck. This section is designed to acquaint the forecaster with the principles of detection and analysis of clouds from rawinsonde data. A complete discussion of this problem is beyond the scope of this training manual. Further information on this subject may be found in the practical training publication, *Use of the Skew T Log P Diagram in Analysis and Forecasting,* NAVAIR 50-1P-5.

IMPORTANCE OF RAWINSONDE OBSERVATIONS (RAOB) IN CLOUD ANALYSIS

Cloud observations regularly available to forecasters in surface synoptic reports leave much to be desired as a basis for cloud forecasting.

Rawinsondes, which penetrate cloud systems, reflect, to some extent (primarily in the humidity trace), the vertical distribution of clouds. If the humidity element were perfect, there would usually be no difficulties in locating cloud layers penetrated by the instrument. Because of the shortcomings in the instrument, however, the relationship between indicated humidity and cloud presence is far from definite, and art empirical interpretation is necessary. Nevertheless, rawinsonde reports give valuable evidence that, when compared with other data, aids greatly in determining a coherent picture of stratiform and frontal cloud distributions. Their value in judging air mass cumulus and cumulonimbus distribution is negligible.

INFERRING CLOUDS FROM RAOB

Theoretically, we should be able to infer from the humidity data of RAOBs the layers where the rawinsonde penetrates cloud layers. In practice, the determination that can be made from temperature and dewpoint curves are often less exact and less reliable than desired. Nevertheless, RAOBs give clues about cloud distribution and potential areas of cloud formation. These clues generally cannot be obtained from any other source.

DEWPOINT AND FROST POINT IN CLOUDS

The temperature minus the dewpoint depression yields the dewpoint, which is defined as the temperature to which the air must be cooled at a constant vapor pressure for saturation to occur. The FROST POINT (that is, the temperature to which the air has to be cooled or heated adiabatically to reach saturation with respect to ice) is higher than the dewpoint except at 0°C, where the two coincide. In the graph shown in figure 4-15, the difference between dewpoint and frost point is plotted as a function of the dewpoint itself.

In a cloud with the temperature above freezing, the true dewpoint will coincide closely with the true air temperature, indicating that the air between the cloud droplets is practically saturated. Minor discrepancies may occur when the cloud is not in a state of equilibrium (when the cloud is dissolving or forming rapidly, or when precipitation is falling through the cloud with raindrops of slightly different temperature than the air); but these discrepancies are very small. In the subfreezing portion of a cloud, the true temperature is between the true dewpoint and the true frost point, depending on the ratio between the quantities of frozen

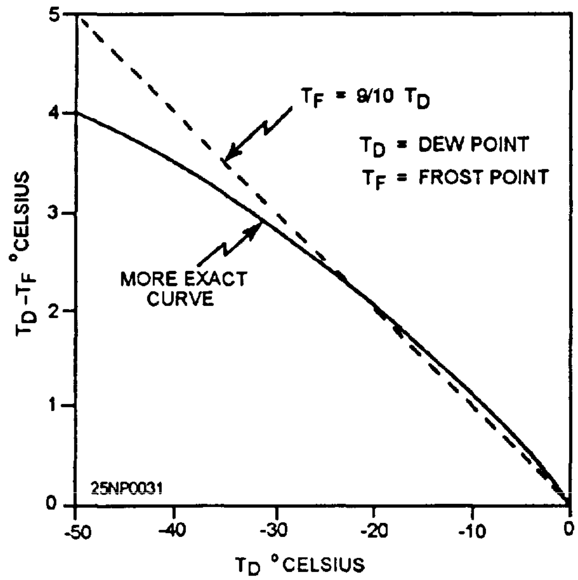

Figure 4-15.-Difference between frost point and dewpoint as a function of the dewpoint.

and liquid cloud particles. If the cloud consists entirely of supercooled water droplets, the true temperature and the true dewpoint will, more or less, coincide. If the cloud consists entirely of ice, the temperature should coincide with the frost point. Therefore, we cannot look for the coincidence of dewpoint and temperatures as a criterion for clouds at subfreezing temperatures. At temperatures below -12°C, the temperature is more likely to coincide with the frost point than the dewpoint.

The graph shown in figure 4-15 indicates that the difference between the dewpoint and frost point increases roughly 1°C for every 10°C that the dewpoint is below freezing. For example, when the dewpoint is –10°C, the frost point equals –9°C; when the dewpoint is –20°C, the frost point is –18°C; and when the dewpoint is –30°C, the frost point is –27°C. Thus, for a cirrus cloud that is in equilibrium (saturated with respect to ice) at a (frost point) temperature of –40°C, the correct dewpoint would be –44°C, (to the nearest whole degree).

We can state, in general, that air in a cloud at temperatures below about –12°C is saturated with respect to ice, and that as the temperature of the cloud decreases (with height), the true frost point/dewpoint difference increases. Any attempt to determine the height of cloud layers from humidity data of a RAOB is, there fore, subject to error. It is possible to overcome some of these errors by a subjective interpretation of the Rawinsonde Observations (RAOBs), as discussed in the following sections.

INTERPRETATION OF RAOB LAYERS WITH RESPECT TO CLOUD LAYERS

The following diagrams (figs, 4-16, 4-17, and 4-18) illustrate the behavior of a rawinsonde during cloud penetration. These diagrams are correlated with aircraft observations or the heights of cloud bases and tops from aircraft flying in the vicinity of an ascending rawinsonde. The difference in time and distance between the aircraft and sounding observations was usually less than 2 hours and 30 miles, respectively. Some of the aircraft reported only the cloud observed above 15,000 feet; others reported all clouds. In figure 4-16 through 4-18, the aircraft cloud observations are entered in the lower left corner of each diagram under the heading cloud; the surface weather report is entered under the aircraft cloud report. Where the low cloud was not reported by the aircraft, the height of the cloud base may be obtained from the surface reports, Aircraft height reports are expressed in thousands of feet, pressure-altitude. The temperature, frost point, and dewpoint curves are indicated by T, T_f, and T_D respectively.

In figure 4-16, a marked warm front is approaching from the south. Moderate continuous rain fell 2 hours later. At 1830 UTC, an aircraft reported solid clouds from 1,000 to 44,000 feet (tropopause). The 1500 UTC sounding shows an increasing dewpoint depression with height and no discontinuity at the reported cloud top of 15,000 feet. A definite dry layer is indicated between

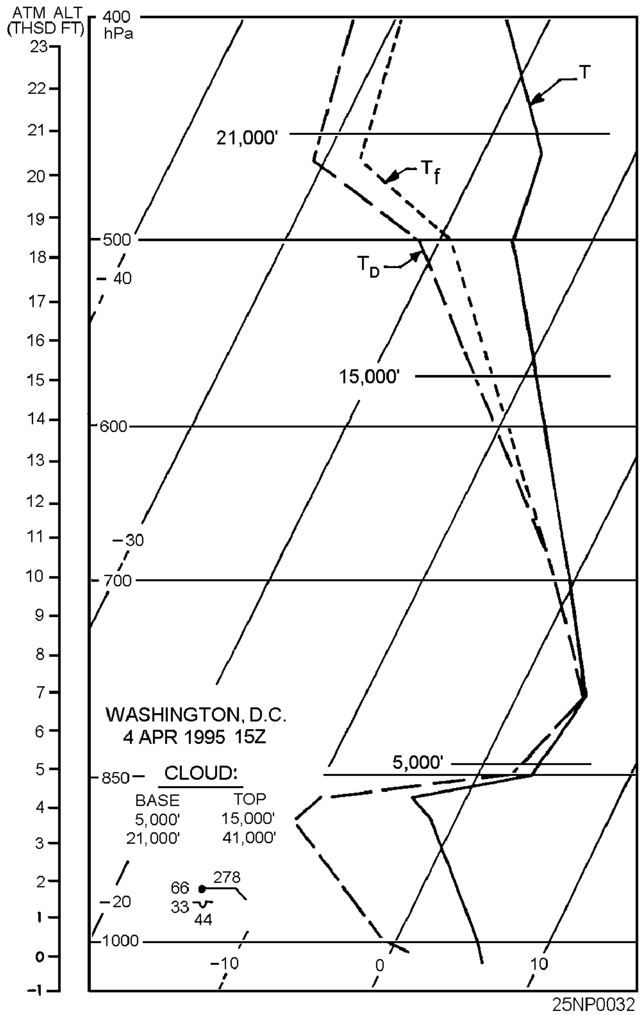

Figure 4-16.-Example of inferring clouds from a RAOB with an active warm front approaching from the south.

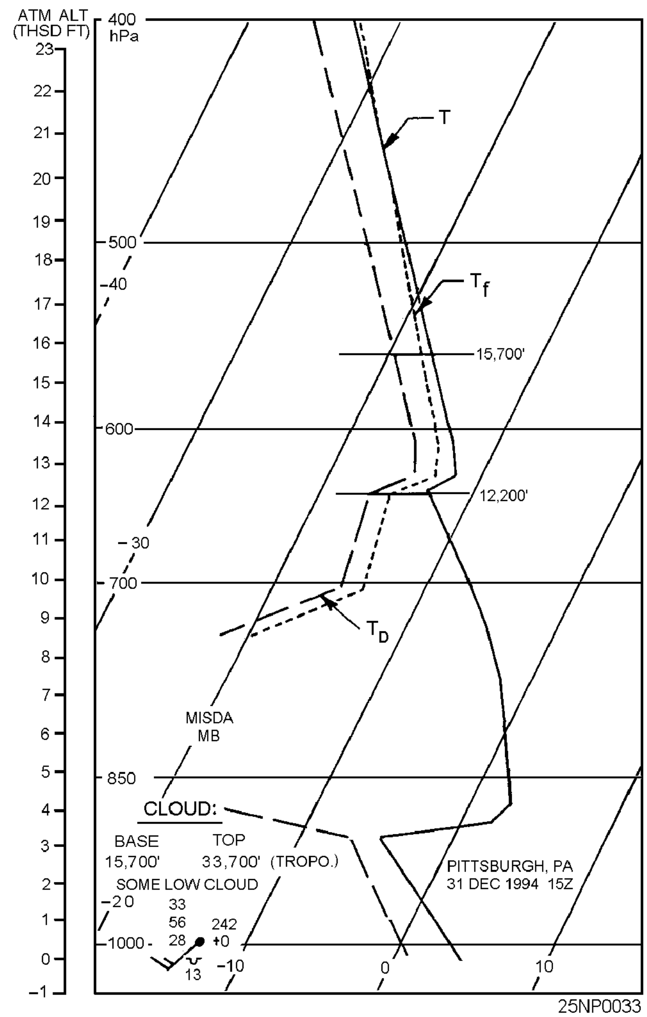

Figure 4-17.-Example of inferring clouds from a RAOB with a middle layer and no precipitation reaching the surface

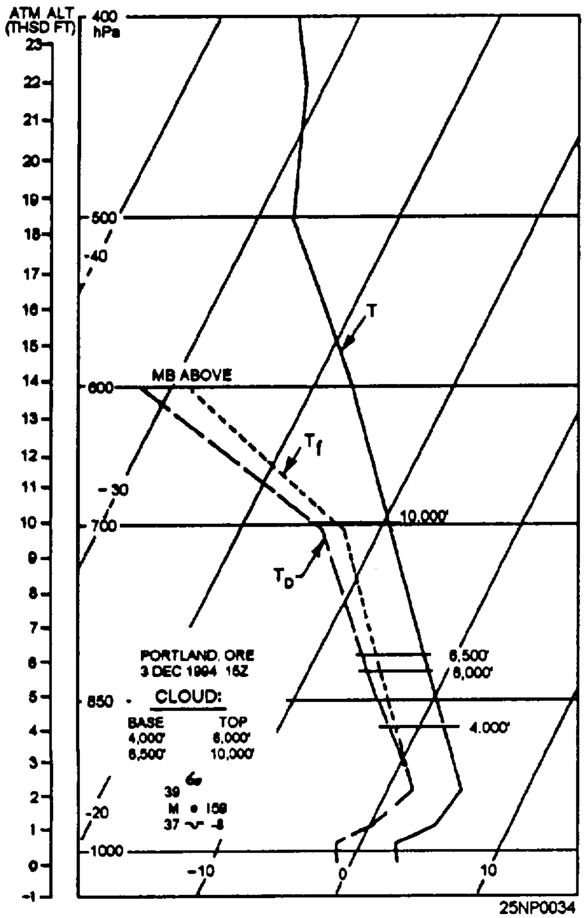

Figure 4-18.-Example of Inferring clouds from a RAOB showing layer clouds with their Intermediate clear layers not showing in the humidity trace.

18,300 and 20,000 feet. The second reported cloud layer is indicated by a decrease in dewpoint depression, but the humidity element is obviously slow in responding. The dewpoint depression at the base of the cloud at 21,000 feet is 14°C and at 400-hPa; after about a 3-minute climb through the cloud, it is still 10°C. From the sounding, clouds should have been inferred to be from about 4,500 feet (base of the rapid humidity increase) to 500-hPa and a second layer from 20,000 feet up. In view of the rapid falling of the cloud free gap between 15,000 and 21,000 feet that followed as the warm front approached, the agreement between reported and inferred conditions is good.

Figure 4-17 shows a middle cloud layer with no precipitation reaching the surface. This is a case of a cloud in the 500-hPa surface with no precipitation reaching the surface; the nearest rain reaching the surface was in Tennessee. The evidence from the sounding for placing the cloud base at 12,200 feet is strong, yet the base is inexplicably reported at 15,700 feet. The reported cloud base of 15,000 feet was probably not representative, since altostratus, with bases 11,000 to 14,000 feet, was reported for most stations over Ohio and West Virginia.

Figure 4-18 shows layered clouds with their intermediate clear layers not showing in the humidity trace. There is good agreement between the sounding and the aircraft report. The clear layer between 6,000 and 6,500 feet is not indicated on the sounding. Thin, clear layers, as well as thin cloud layers, usually cannot be recognized on the humidity trace.

Comparisons between soundings and cloud reports provide us with the following rules:

1. A cloud base is almost always found in a layer, indicated by the sounding, where the dewpoint depression decreases.

2. You should not always associate a cloud with a layer of decreasing dewpoint, but only when the decrease leads to minimum dewpoint depressions from 6°C to 0°C. However, at temperatures below -25°C, dewpoint depressions in clouds are often higher than 6°C.

3. The dewpoint depression in a cloud is, on the average, smaller in clouds that have higher temperatures. typical dewpoint depressions are 1°C to 2°C at temperatures of 0°C and above, and 4°C between -10°C and -20°C.

4. The base of a cloud should be located at the base of the layer of decreasing dewpoint depression, if the decrease is sharp.

5. If a layer of decreasing dewpoint depression is followed by a layer of a stronger decrease, the cloud base should be associated with the base of the strongest decrease.

6. The top of a cloud layer is usually indicated by an increase in dewpoint depression. Once a cloud base is determined, the cloud is extended up to a level where a significant increase in dewpoint depression starts. The gradual increase of dewpoint depression with height in a cloud is not significant.

In addition to the above analysis, another study was made to determine how reliable the dewpoint depression is as an indicator of clouds. The results are summarized in figure 4-19. Each graph shows the percent probability of the existence of a cloud layer in January for different values of dewpoint depression. On each graph one curve shows the probability of clear or

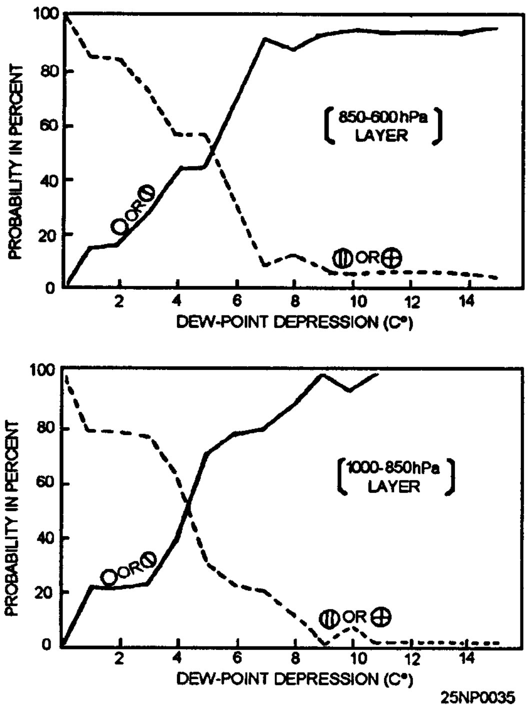

Figure 4-19.-Percent probability of existence of cloud layer bases for different values of dewpoint depression (degrees C). Solid lines represent probability of clear or scattered conditions; dashed lines, the probability of broken or overcast conditions with the cloud layer bases between 1,000-hPa and 600-hPa.

scattered conditions as a function of the dewpoint depression; the other curve shows that of broken or overcast conditions. Separate graphs are based on 1,027 observations, which are enough to indicate the order of magnitude of the dewpoint depressions at the base of winter cloud layers. Minor irregularities in the curves were not smoothed out because it is not certain that they are all due to insufficient data. The graphs are applicable without reference to the synoptic situation. For a given winter sounding, you can estimate from the graph the probability of different sky cover conditions with cloud bases between 1,000-hPa and 600-hPa for layers of given minimum dewpoint depressions.

HUMIDITY FIELD IN THE VICINITY OF FRONTAL SYSTEMS

Studies of the humidity field throughout frontal zones indicate there is a tongue of dry air extending downward in the vicinity of the front, and sloping in the same direction as the front. One study found that such a dry tongue was more or less well developed for all frontal zones investigated. This dry tongue was best developed near warm fronts; it extended, on the average, down to 700-hPa in cold fronts and to 800-hPa in warm fronts. In about half the fronts, the driest air was found within the frontal zone itself on occasion it was found on both the cold and warm sides of the zone. About half the flights through this area showed a sharp transition from moist to dry air, and the change in frost point on these flights averaged about 20°C in 35 miles. Some flights gave changes of more than 20°C in 20 miles.

As a frontal cloud deck is approached, the dewpoint or frost point depression starts diminishing rapidly. At distances beyond 10 to 15 nm, this variation is much less. You should keep this fact in mind when attempting to locate the edge of a cloud deck from rawinsonde data. Linear extrapolation or interpolation of dewpoint depressions cannot be expected to yield good results. For instance, when one station shows a dewpoint depression of 10°C and the neighboring station shows saturation, the frontal cloud may be anywhere between them, except within about 10 nm of the driest station.

Since the frontal cloud masses at midtropospheric levels is usually surrounded by relatively dry air, it is possible to locate the edge of the cloud mass from humidity data on constant pressure charts. This is so because the typical change in dewpoint depression in going from the cloud edge into cloudfree air is considerably greater than the average error in the reported dewpoint depression.

500-hPa ANALYSIS OF DEWPOINT DEPRESSION

Figure 4-20 shows an analysis of the 500-hPa dewpoint depression field superimposed upon an analysis of areas of continuous precipitation, and of areas of overcast middle clouds. The 500-hPa dewpoint depression isopleths were drawn independently of the surface data. The analysis shows the following:

1. The regions of high humidity at the 500-hPa level coincide well with the areas of middle clouds and the areas of precipitation.

2. The regions of high humidity at the 500-hPa level are separated from the extensive dry regions by strong humidity gradients. These gradients are, in all probability, much stronger than those shown on this analysis.

3. A dewpoint depression of 4°C or less is characteristic of the larger areas of continuous

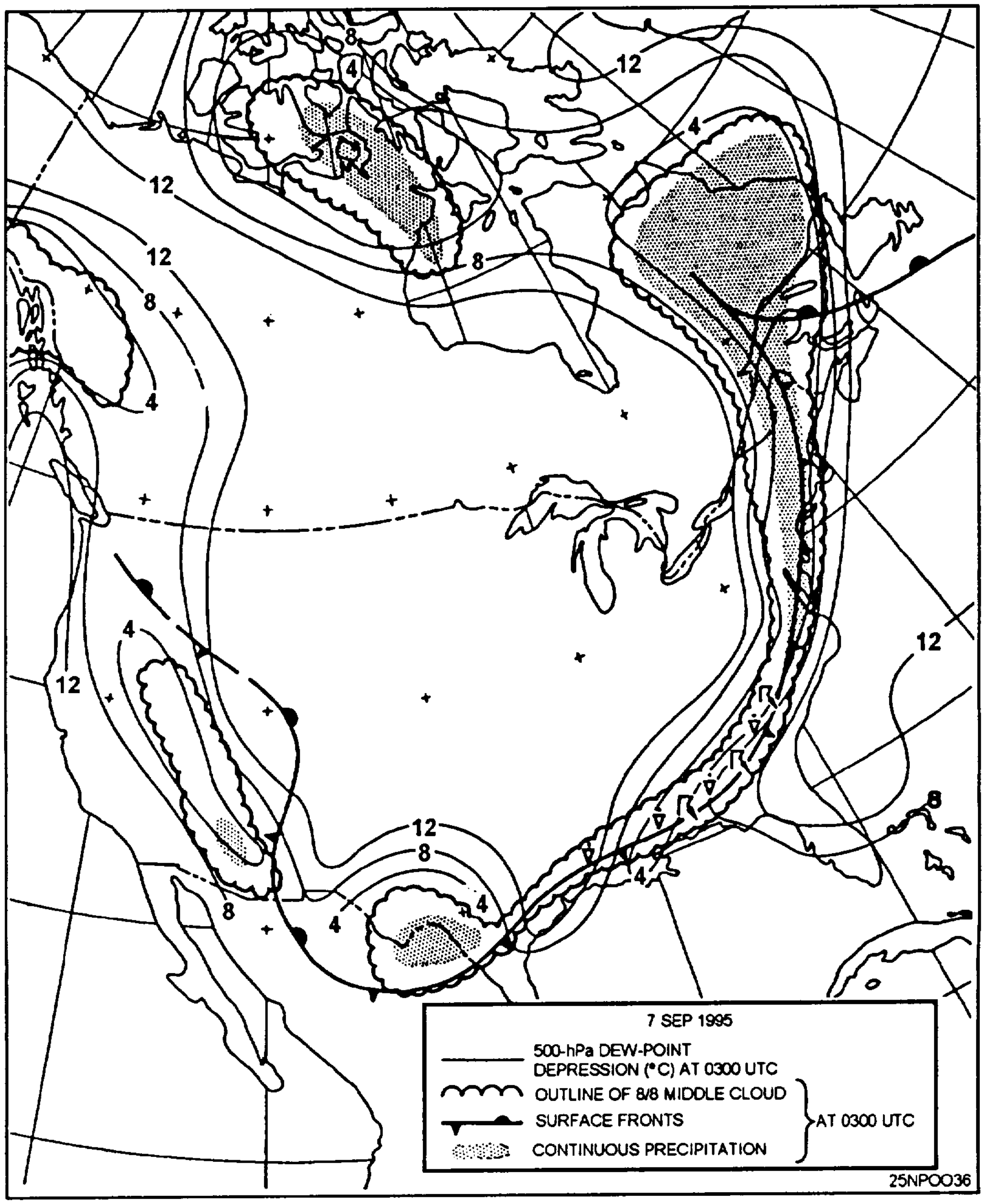

Figure 4-20.-Surface fronts, areas of continuous precipitation, areas covered by 8/8 middle clouds, and isolines of 500-hPa dewpoint depression at 0300 UTC, 7 September 1995.

precipitation, and also of the larger areas of overcast middle clouds.

Since the 500-hPa level dewpoint depression analysis agrees well with the surface analysis of middle cloud and precipitation, the possibility exists of replacing or supplementing one of these analyses with the other.

The characteristics of the 500-hPa level dewpoint depression analysis, outlined above, make it a valuable adjunct to the surface analysis. These analyses can be compared and, by cross-checking, each can be completed with greater accuracy than if they were done independently.

THREE-DIMENSIONAL HUMIDITY ANALYSIS—THE MOIST LAYER

Using a single level (for example, the 500-hPa level dewpoint depression analysis) to find probable cloud areas does not indicate clouds above or below that level. For example, if the top of a cloud system reached only to 16,000 feet, and there was dry air above at 500-hPa (approximately 18,000 feet), you wouldn't suspect, from the 500hPa analysis, the existence of clouds below the 500-hPa level.

However, an analysis of the extension of the moist layers in three dimensions can be obtained simply by scrutinizing individual RAOBs. Those selected should be in the general vicinity of, and the area 500 to 1,200 miles upstream of, the area of interest, depending on the forecast period. The heights of the bases and tops can be indicated, though there is little advantage in indicating a dry layer 2,000 to 3,000 feet thick sandwiched between thicker moist layers. Usually, it is sufficient to indicate the entire moist layer, without bothering about any finer stratums. A survey of the cloud field is made easier by writing the heights of the bases and tops in different colors.

A moist layer for the sake of simplicity may be defined as a layer having a frost point depression of 3°C or less (i.e., a dewpoint depression of 4°C at –10°C; 5°C at -20°C; 6°C at –30°C).

PRECIPITATION AND CLOUDS

The type and intensity of precipitation observed at the surface is related to the thickness of the cloud aloft, and particularly to the temperatures in the upper portion of the cloud.

The results of a study relating cloud-top temperatures to precipitation type and intensity are as follows:

- From aircraft ascents through stratiform clouds, along with simultaneous surface observations of precipitation, it was found that 87 percent of the cases where drizzle occurred, it fell from clouds whose cloud-top temperatures were warmer than –5°C. The frequency of rain or snow increased markedly when the cloud-top temperature fell below –12°C.

- When continuous rain or snow fell, the temperature of the coldest part of the cloud was below –12°C in 95 percent of the cases.

- Intermittent rain was mostly associated with cold cloud-top temperatures.

- When intermittent rain was reported at the surface, the cloud-top temperature was colder than –12°C in 81 percent of the cases, and colder than –20°C in 63 percent of the cases. From this, it appears that when minor snow (continuous or intermittent) reaches the ground from stratiform clouds, the clouds (solid or layered) extend in most cases to heights where the temperature is well below –12°C, or even –20°C.

This rule cannot be reversed. When rain or snow is not observed at the surface, middle clouds may well be present in regions where the temperature is below -12°C or –20°C. Whether or not precipitation reaches the ground will depend on the cloud thickness, height of the cloud base, and the dryness of the air below the base.

INDICATIONS OF CIRRUS CLOUDS IN RAOB

Cirrus clouds form at temperatures of –40°C or colder. At these temperatures, as soon as the air is brought to saturation, the condensate immediately freezes. The ice crystals often descend in altitude slowly, to levels that have air temperatures of –30°C, and persist if the humidity below the formation level is high enough to support saturation. In general, cirrus clouds are found in layers that are saturated, or supersaturated, with respect to ice at temperatures colder than 0°C.

OBSERVATION AND FORMATION OF CIRRUS

Cirrus, or cirriform clouds, are divided into three general groups: cirrus (proper), cirrostratus, and cirrocumulus. Cirrus clouds, detached or patchy, usually do not create a serious operational problem. Cirrostratus and extensive cirrus haze, however, may be troublesome in high-level jet operations, aerial photography, interception, rocket tracking, and guided missile navigational systems. Therefore, a definite requirement for cirrus cloud forecasting exists.

The initial formation of cirrus clouds normally requires that cooling take place to saturation, and to have temperatures near –40°C. Under these conditions, water droplets are first formed, but most of them immediately freeze. The resulting ice crystals persist as long as the humidity remains near saturation with respect to ice. There is some evidence that the speed of the cooling, and the kind and abundance of freezing nuclei, may have an important effect on the form and occurrence of cirrus clouds. Slow ascent starts crystallization at humidities substantially below saturation; this is presumably the case in extensive cirrostratus clouds associated with warm frontal altostratus clouds. If slow ascent occurs in air that has insufficient freezing nuclei, a widespread haze may result, which at –30° to –40°C is predominantly composed of water droplets. In the case of more rapid cooling, there is a tendency for the initial condensation to contain a higher proportion of water droplets, which leads to a "mixed cloud' that will convert to ice or snow in time. Presumably, dense cirrus, fine cirrus, cirrocumulus, and anvil cirrus clouds are of this type. It is assumed that fine cirrus clouds (proper) are formed in shallow layers that are undergoing rapid convection due to advection of colder air at the top of the shear layer.

On the other hand fine cirrus and cirrostratus clouds are so often associated, and cirrostratus clouds are so often reported by pilots as developing from the merging of fine cirrus clouds, that there is a question whether the process of formation in cirrus and cirrostratus clouds are essentially different. Nevertheless, the prevailing crystal types in cirrus and cirrostratus clouds seem to differ, though this may not be universal, or may merely represent different stages in cirrus cloud evolution.

Horizontal risibilities within cirrostratus clouds are generally between 500 feet and 2 nm. Thin cirrus haze, invisible from the ground, often reduces the visibility to 3 nm.

A rule of thumb for forecasting or estimating the visibility within thin cirrus or other high cloud (temperatures below –30°C) follows:

Visibility = 1/2 nm times dewpoint depression in degrees C. For example; temperature is –35°C, dewpoint is –38°C, and visibility = 1/2 x 3 = 1 1/2 nm. This rule has been used successfully only in the Arctic where poor visibility in apparently cloudfree air is often encountered.

THE CIRRUS CLOUD FORECASTING PROBLEM

Many forecasters have attempted to forecast cirrus clouds by using frontal or cyclone models. This procedure is not always satisfactory. There are a number of parameters, both surface and aloft, that have been correlated with cirrus cloud formation. A few of the more prominent parameters are mentioned in the following text.

Surface Frontal Systems

Frontal and cyclone models have been developed that embody an idealized cloud distribution. In these models, the cirrus clouds are lowering and thickening to form altostratus clouds, which indicates an advancing warm front.

Fronts Aloft

Above 500-hPa the concepts of air masses and fronts have little application. Most of the fine cirrus clouds observed ahead of and above warm fronts or lows initially form independent of the frontal middle cloud shield, though later it may trail downward to join the altocumulus and altostratus cloud shields. With precipitation occurring in advance of a warm front, a 60-percent probability exists that cirrus clouds are occurring above. Cirrus clouds observed with the cold front cloud shield either originate from cumulonimbus along and behind the front or from convergence in the vicinity of the upper trough. In many cases there is no post cold front cirrus clouds, probably due to marked subsidence aloft.

Contour Patterns Aloft

One forecasting rule used widely states that "the ridge line at 20,000 feet, about 500 hpa, preceding a warm front marks the forward edge of the cirrus cloud shield."

For a typical 500-hPa wave pattern, the following information applies:

- No extensive cirrostratus clouds will occur before the surface ridge line arrives.
- Extensive cirrostratus clouds follow the passage of the surface ridge line.
- No middle clouds appear before the arrival of the 500-hPa ridge line.
- Middle clouds tend to obscure the cirrus clouds after passage of the 500-hPa ridge line.
- When the 500-hPa wave has a small amplitude, the cirrus cloud arrival is delayed and the clouds are thinner.
- The greater the 500-hPa convergence from trough to ridge, the more cirrus clouds between the surface and 500-hPa ridge lines.

Cirrus Clouds in Relation to the Tropopause

Experiences of pilots have confirmed that the tops of most cirrus clouds are at or below the tropopause. In the midlatitudes, the tops of most cirrus cloud layers are at or within several thousand feet of the polar tropopause. Patchy cirrus clouds are found between the polar tropopause and the tropical tropopause. A small percentage of cirrus clouds, and sometimes extensive cirrostratus, may be observed in the lower stratosphere above the polar tropopause, but mainly below the level of the jetstream core. The cirrus clouds of the equatorial zone also generally extend to the tropopause. There is a general tendency for the mean height of the bases to increase from high to low latitudes more or less paralleling the mean tropopause height, ranging from 24,000 feet at 70°to 80°atitude to 35,000 to 4,000 feet or higher in the vicinity of the equator. The thickness of individual cirrus cloud layers are generally about 800 feet in the midlatitudes. The mean thickness of cirrus clouds tends to increase from high to low latitudes. In polar continental regions in winter, cirrus clouds are virtually based at the surface. In the midlatitudes and in the tropics, there is little seasonal variation.

Cirrus Clouds in Relation to the Jetstream

A discussion of cloud types associated with the jetstream is contained in the *AG2* TRAMAN, volume 1. In addition to this information, we will discuss a few studies pertaining to cloud types. All of these studies agree that most of the more extensive and dense clouds clouds are on the equatorward of the jet axis. The observed frequency of high clouds poleward of the jet axis can be accounted for as the upper reaches of a cold front, or cold lows, not directly related to the jetstream. In some parts of a trough, these high clouds may tend to be dense, and in other areas thin.

PREDICTION OF SNOW VS RAIN

LEARNING OBJECTIVES: Evaluate the surface and upper-level synoptic situations in determining the form of precipitation in your forecast.

Typically, an inch or so of precipitation in the form of rain will cause no serious inconvenience. On the other hand the same amount of precipitation in the form of snow, sleet, or freezing rain can seriously interfere with naval operations. In such cases, the snow versus rain problem may become a factor of operational significance.

Sleet and freezing rain, which often may occur in the intermediate period between snow and rain, are generally grouped with snow in our discussion. Any decision arrived at for the snow versus rain problem would, naturally, have to be modified, dependant on your geographical location. This should be easily accomplished through a local study of the optimum conditions. The various techniques and systems presented here will often complement each other. The approach used here is a discussion of the general synoptic patterns and the thermal relationship; that is, the use of temperatures at the surface and aloft, and the presentation of an objective technique to distinguish the types of precipitation.

GEOGRAPHICAL AND SEASONAL CONSIDERATIONS

The forecasting problem of snow versus rain arises, naturally, during the colder months of the year. In midwinter when the problem is most serious in the northern states, the southern states may not be concerned.

PHYSICAL NATURE OF THE PROBLEM

The type of precipitation that reaches the ground in a borderline situation is essentially dependent on two conditions. There must be a stratum of above-freezing temperatures between the ground and the level at which precipitation is forming, and this stratum must be sufficiently deep to melt all of the falling snow prior to striking the surface. Thus, a correct prediction of rain or snow at a given location depends largely on the accuracy with which the vertical distribution of the temperature, especially the height of the freezing level, can be predicted. On the average, it is generally satisfactory to assume that the freezing level must be at least 1,200 feet above the surface to ensure that most of the snow will melt before reaching the surface.

Effects of Advection

In the lower troposphere, above the surface, horizontal advection is usually the dominant factor affecting local temperature changes. In most precipitation situations, particularly in borderline situations, warm air advection and upward motion are occurring simultaneously, giving rise to the fact that warming generally accompanies precipitation. However, this effect is frequently offset when there is weak warm advection, or even cold advection, in the cold air mass in the lower layers.

In situations where precipitation is occurring in association with a cold upper low, upward motion is accompanied by little, if any, warm advection. In such borderline cases, precipitation may persist as snow, or tend to turn to snow, due to cooling, as a result of upward motion or advection.

Nonadiabatic Effects

The most important of the nonadiabatic effects taking place during the precipitation process is the cooling, which takes place due to evaporation as the precipitation falls through unsaturated air between the clouds and the surface. This effect is especially pronounced when very dry air is present in the lower levels, with wet-bulb temperatures at or below freezing. Then, even if the dry-bulb temperature is above freezing in a layer deeper than 1,200 feet in the lower levels, the precipitation may still fall as snow, since the evaporation of the snow will lower the temperatures in the layer between the cloud and the surface until the below-freezing wet-bulb temperatures are approached.

The actual cooling that occurs during the period when evaporation is taking place may often be on the order of 5° to 10°F within an hour. After the low-level stratum becomes saturated, evaporation practically ceases, and advection brings a rise in temperature in the low levels. However, reheating often comes too late to bring a quick change to rain since the temperatures may have dropped several degrees below freezing, and much snow may have already fallen. The lower levels may be kept cool through the transfer of any horizontally transported heat to the colder, snow-covered surface.

Melting of Snow

Melting snow descending through layers that are above freezing is another process which cools a layer. To obtain substantial temperature changes due to melting, it is necessary to have heavy amounts of precipitation falling, and very little warm air advection. As cooling proceeds, the temperature of the entire stratum will reach freezing, so that a heavy rainstorm could transform into a heavy snowstorm,

Incidents of substantial lowering of the freezing level due to melting are relatively rare. The combination of heavy rain, and little, if any, warm advection is an infrequent occurrence.

Combined Effects

The combined effects of horizontal temperature advection, vertical motion, and cooling due to evaporation are well summarized by observations of the behavior of the bright band on radar (approximately 3,000 ft). Observers have found that within the first 1 1/2 hours after the onset of precipitation, the bright band lowers by about 500 to 1,000 feet. This is attributable primarily to evaporational cooling, and probably secondary to melting. Since evaporational cooling ceases as saturation is reached, warm air advection, partially offset by upward motion, again becomes dominant, and the bright band ascends to near its original level. The bright band will ascend to its original level approximately 3 hours after the onset of precipitation, and may ascend a few thousand additional feet.

Other nonadiabatic effects, such as radiation and heat exchange with the surface, probably play a relatively smaller role in the snow-rain problem. However, it is likely that the state of the underlying surface (snow-covered land versus open water) may determine whether the lower layers would be above or below freezing. Occasionally, along a seacoast in winter, heat from the open water keeps temperatures offshore above freezing in the lower levels. Along the east coast of the United States, for example, coastal areas may have rain, while a few miles inland snow predominates. This situation is associated with low-level onshore flow, which is typical of the flow associated with many east coast cyclones. Actually, this situation cannot be classified as a purely nonadiabatic effect since the warmer ocean air is being advected on shore.

GENERAL SYNOPTIC CONSIDERATIONS

The snow versus rain problem usually depends upon relatively small-scale synoptic considerations, such as the exact track of the surface disturbance, whether the wind at a coastal station has an onshore component, the position of the warm front, and the orientation of a ridge east of the low.

In the larger sense, the snow versus rain zone is directly related to the position of the polar front. The location of the polar front is, in turn, closely related to the position of the belt of strong winds in the middle and upper troposphere. When the westerlies extend farther to the south, storm tracks are similarly affected, and the snow-rain zone may be farther to the south. As the westerlies shift northward of their normal position, the storm tracks develop across Canada. Concurrent with this northward shift, the United States has above normal temperatures, and the snow-rain problem exists farther to the north.

With a high zonal index situation aloft, the snow-rain zone will extend in a narrower belt, often well ahead of the surface perturbation and will undergo little north-south displacement, Those areas with precipitation occurring will not undergo a change from one form to another since there is relatively little advection of warm or cold air with a high zonal condition.

When the upper-level wave is of large or increasing amplitude (low zonal index), it is difficult to generalize about the characteristics of the snow versus rain problem without considering the surface perturbation.

Up to this point, we have discussed the snow-rain pattern in association with an active low of the classical type, The rate of precipitation accumulation here is rapid, and the transition period of freezing rain or sleet is short, usually on the order of a few hours or less. Another situation in which there is frequently a snow versus rain problem is that of a quasi-stationary front in the southern states, with a, broad west-southwest to southwest flow aloft, and a weak surface low. The precipitation area in this case tends to become elongated in the direction of the upper-level current. The precipitation rate may be slow, but it occurs over a longer period. Often a broad area of sleet and freezing rain exists between belts of snow and rain, leading to a serious icing condition over an extensive region for a period of several hours or more. This pattern of precipitation changes either as an upper trough approaches from the west and initiates cyclogenesis on the front or as the flow aloft veers and precipitation ceases.

FORECASTING TECHNIQUES AND AIDS

Approaches to the snow versus rain forecasting problem have generally fallen into three broad categories. The first category depends on the use of observed flow patterns and parameters to predict the prevalent form of precipitation for periods as much as 36 hours in advance. The second category consists of studies relating local parameters to the occurrence of rain or snow at a particular station, or area. In this approach, it is assumed thermal parameters will be obtainable from prognoses. This approach tends to have its greatest accuracy for periods of 12 hours, or less, since longer periods of temperature predictions for the boundary zone between rain and snow are very difficult to make with sufficient precision. A third category used involves the use of one of the many objective techniques available. A number of stations have developed objective local techniques. The method presented here is applicable to the eastern half of the United States. Thus, the general procedure in making a snow versus rain forecast at present is to use a synoptic method for periods up to 24 or 36 hours, and then consider the expected behavior of thermal parameters over the area to obtain more precision for periods of about 12 hours or less.

A number of methods based on synoptic flow patterns applicable to the United States are described in the U.S. Department of Commerce's publication. *The Prediction of Snow vs Rain, Forecasting Guide No. 2.* These methods are mostly local in application and are beyond the scope of this manual.

Prognostic charts from the National Meteorological Center and other sources should be used whenever and wherever available, not only to determine the occurrence and extent of precipitation, but for the prediction of the applicable thermal parameters as well.

Methods Employing Local Thermal Parameters

The following text discusses methods of employing surface temperature, upper-level temperatures, 1000- to 700-hPa and 1000- to 500-hPa thicknesses, the height of the freezing level, and combined parameters for the prediction of snow versus rain. All of these are interdependent, and should be considered simultaneously.

SURFACE TEMPERATURE.— Surface temperature considered by itself is not an effective criterion. Its use in the snow versus rain problem has generally been used in combination with other thermal parameters. One study for the Northeastern United States found that at 35°F snow and rain occurred with equal frequency, and by using 35°F as the critical value (predict snow at 35°F and below, rain above 35°F), 85 percent of the original cases could be classified, Another study based on data from stations in England suggested a critical temperature of 34.2°F, and found that snow rarely occurs at temperatures higher than 39°F. However, it is obvious from these studies that even though surface temperature is of some value in predicting snow versus rain, it is often inadequate. Thus, most forecasters look to upper-level temperatures as a further aid to the problem.

UPPER-LEVEL TEMPERATURES.— Two studies of the Northeastern United States found that temperatures at the 850-hPa level proved to be a good discriminating parameter, and that including the surface temperature did not make any significant contribution. The discriminating temperatures at the 850-hPa level were -2° to -4°C. Another study found that the area bounded by the 0°C isotherm at 850-hPa level and the 32°F isotherm at the surface, when superimposed upon the precipitation area, separated the snow-rain precipitation shield in a high percentages of cases. A range of -2° to -4°C at the 850-hPa level should be used along coastal areas, and also behind deep cold lows, At mountain stations a higher level would have to be used.

A technique that uses temperatures at mandatory levels (surface, 1000-, 850-, 700-, and 500-hPa, etc.) is advantageous because of the availability of charts at these levels. There is, however, the occasional problem where temperature inversions are located near the 850- or 700-hPa levels, so that the temperature of one level may not be indicative of the layer above or below. This difficulty can be overcome by using thickness, which is a measure of the mean temperature of the layer.

THICKNESS.— The National Weather Service has examined both the 1,000- to 700-hPa and 1,000- to 500-hPa thickness limits for the eastern half of the United States.

A generalized study of 1,000- to 500-hPa thickness as a predictor of precipitation forms in the United States was made by A. J. Wagner, More complete details on this study can be found in *The Prediction of Snow vs Rain, Forecasting Guide No. 2.*

Wagner's data was taken from a study of 40 locations in the United States for the colder months of a 2-year period. Cases were limited to surface temperatures between 10°F and 50°F. The form of precipitation in each case was considered as belonging in one of two categories-frozen which includes snow, sleet, granular snow, and snow crystals; and unfrozen, which includes rain, rain and snow mixed, drizzle, and freezing rain and drizzle.

Equal probability, or critical thickness values, were obtained from the data at each location. From this study it was clear that the critical thickness values increase with increasing altitude. This altitude relationship is attributable to the fact that a sizable portion of the thickness stratum is nonexistent for high-altitude stations, and obviously does not participate in the melting process. To compensate for this, the equal probability thickness values must increase with station elevation. For higher altitude stations, thickness values between the 850- to 500-hPa or 700- to 500-hPa stratums, as appropriate, should prove to be better dated to precipitation form.

The Wagner equal probability chart is reproduced in figure 4-21.

Wagner's study also indicates that the form of precipitation can be specified with a certainty of 75 percent at plus or minus 30 meters from the equal probability value, increasing to 90 percent certainty at plus or minus 90 meters from this value. Stability is the parameter that accounts for the variability of precipitation for a given thickness at a given point. This fact is taken into account in the following reamer: if the forecast precipitation is due to a warm front that is more stable than usual, the line separating rain from frozen precipitation is shifted toward higher thickness values. Over the Great Lakes, where snow occurs in unstable, or stable conditions, the equal probability thickness is lower than that shown in figure 4-21 for snow showers, and higher than that shown in figure 4-21 for warm frontal snow.

HEIGHT OF THE FREEZING LEVEL.— The height of the freezing level is one of the most critical thermal parameters in determining whether snow can reach the ground. It was pointed out earlier that theoretical and observational evidence indicates that a freezing level averaging 1,200 feet or more above the surface is usually required to ensure that most of the snow will melt before reaching the surface. This figure of 1,200 feet can thus be considered as a critical or equal probability value of the freezing level.

COMBINED THERMAL PARAMETERS.— From the foregoing discussion, you can conclude that no one method, when used alone, is a good discriminator in the snow versus rain forecasting problem. Therefore, you should use a combination of the surface temperature, height of the freezing level, 850-hPa temperature, and the 1,000-to 700-hPa and /or 1,000- to 500-hPa thicknesses to arrive at the forecast, There is generally a high correlation between the 850-hPa temperature, and the 1,000- to 700-hPa thickness and between the 700-hPa temperature and the 1,000- to 500-hPa thickness. Certainly an accurate temperature forecast for these two levels would yield an approximate thickness value for discriminating purposes.

Additional Snow versus Rain Techniques

The determining factor in the form of precipitation in this study was found to be the distribution of temperature and moisture between the surface and the 700-hPa level at the time of the beginning of precipitation. The median level of 850-hPa was studied in conjunction with the precipitation area and the 32°F

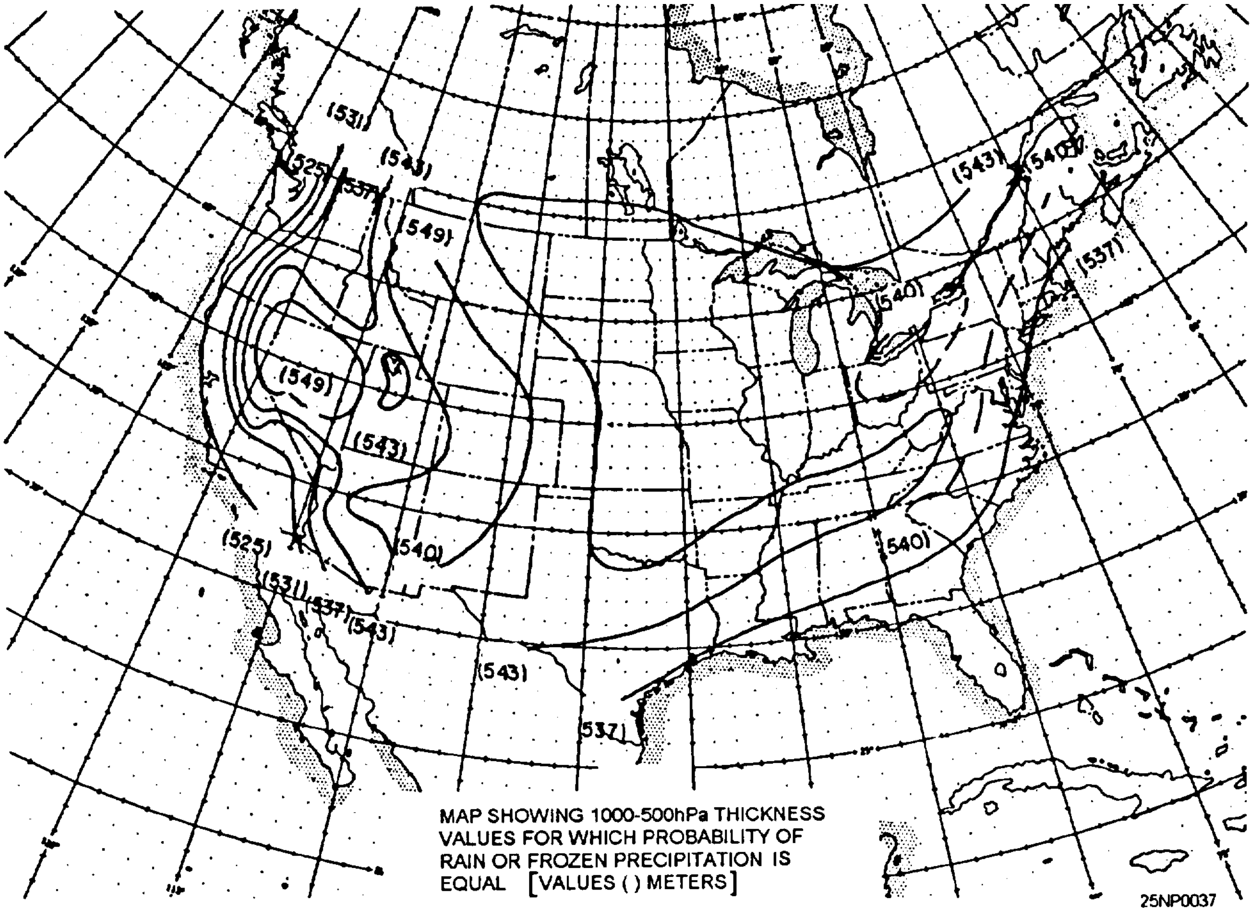

Figure 4-21.-Map showing 1,000- to 500-hPa thickness values for which probability of rain or frozen precipitation is equal (after Wagner).

isotherm sketched on the surface chart. This method presents an objective and practical method by which the forecaster can make a decision on whether the precipitation in winter will be rain, snow, freezing rain, sleet, or some combination of these.

The following objective techniques can be applied to the land areas south of 50° north latitude and east of a line drawn through Williston, North Dakota; Rapid City, South Dakota; Goodland, Kansas; and Amarillo, Texas.

The area outlined by the 0°C isotherm at 850-hPa and the 32°F isotherm on the surface chart, when superimposed upon the precipitation area, generally separates the forms of precipitation. Most of the pure rain was found on the warm side of the 32°F isotherm, and most of the pure snow on the cold side of 0°C isotherm, with intermediate types falling generally within the enclosed area between these two isotherms. It was also observed that in a large majority of situations, evaporation and condensation was a sizable factor, both at 850-hPa and at the surface level in its affect upon temperature. With this in mind, the wet-bulb temperature was selected for investigation because of its conservative properties with respect to evaporation and condensation, and also because of its ease of computation directly from the temperature and dewpoint. The surface chart is used for computations of the 1,000-hPa level because the surface chart approximates the 1,000-hPa level for most stations during a snow situation; therefore, little error is introduced. *IT MUST BE REMEMBERED THAT ALL PREDICTIONS ARE BASED ON FORECAST VALUES.*

MOVEMENT OF THE 850-hPa 0°C ISOTHERM.— A reasonably good approximation for forecasting the 0°C isotherm at the 850-hPa level can be made subjectively by use of a combination of extrapolation and advection, considerations of synoptic developments, and the rules listed in the following paragraphs. (See figure 4-22, views (A) and (B), for typical warm and cold air advection patterns at 850-hPa.)

The following rules for the movement of the 24-hour, 850-hPa level temperature change areas have been devised

- Maximum cooling takes place between the 850-hPa contour trough and the 850-hPa isotherm ridge east of the trough.

- Maximum warming takes place between the 850-hPa contour ridge and the 850-hPa isotherm trough east of the contour ridge.

- Changes are slight with ill-defined isotherm/contour patterns.

- Usually, little change occurs when isotherms and contours are in phase at the 850-hPa level.

- The temperature falls at the 850-hPa level tend to replace height falls at the 700-hPa level in an average of 24 hours. Conversely, temperature rises replace height rises.

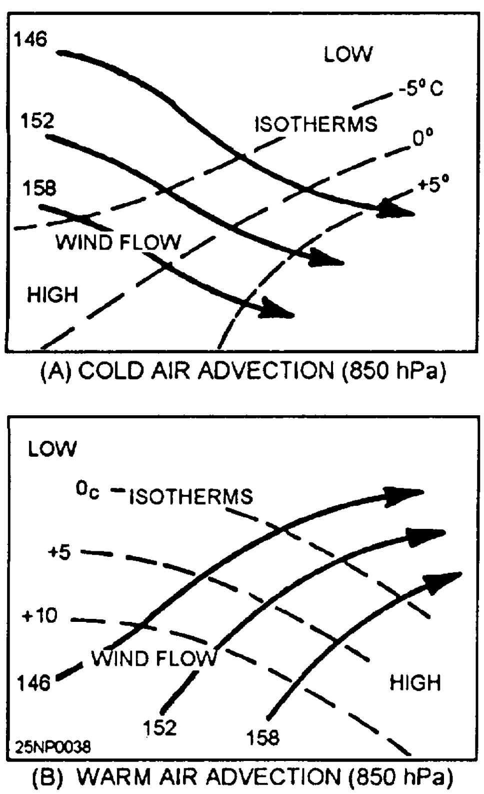

Figure 4-22.-Typical cold and warm air advection patterns at 850-hPa. (A) Cold; (B) Warm.

- With filling troughs or northeastward moving lows, despite northwest flow behind the trough, 850-hPa level isotherms are seldom displaced southward, but follow the trough toward the east or northeast

- Always predict temperature falls immediately following a trough passage.

- Do not forecast temperature rises of more than 10 to 2°F in areas of light or sparse precipitation in the forward areas of the trough. If the area of precipitation is widespread and moderate or heavy, forecast no temperature rise.

- With eastward moving systems under normal winter conditions (trough at the 700-hPa level moving east about 10° per day), a distance of 400 nautical miles to the west is a good point to locate the temperature to be expcted at the forecasting point 24 hours later. A good 850-hPa temperature advection speed seems to be about 75 percent of the 700-hPa trough displacement.

The following is step-by-step procedures for moving the 850-hPa 0°C isotherm:

1. Extrapolate the movement of the thermal ridge and troughs for 12 and 24 hours. If poorly defined, this step may be omitted. The amplitude of the thermal wave may be increased or decreased subjectively if, during the past 12 hours, there has been a corresponding increase or decrease in the height of the contours at 500-hPa.

2. The thermal wave patterns will maintain their approximate relative position with the 850-hPa level contour troughs and ridges. Therefore, the 12- and 24-hour prognostic positions of the contour troughs and ridges should be made, and the extrapolated positions of the thermal points checked against this contour prognosis. Adjustments of these thermal points should be made.

3. Select points on the 0°C isotherm that lie between the thermal ridge and trough as follows: one or two points in the apparent warm advection area, and one or two points in the apparent cold advection area. Apply the following rules to these selected points.

- Warm advection area. If the points lie in a near saturated or precipitation area, they will remain practically stationary with respect to the contour trough. If the points lie in a nonsaturated area, but one that is expected to become saturated or to lie in precipitation area, then it will remain stationary or move upwind slightly to approximate y the prognostic position of the 0°C wet-bulb. If the point does not fall in the above two categories, it will be advected with about 50 percent of the wind component normal to the isotherm. Note in all three cases above, the movement is related to the contour pattern.

- Cold Advection area. Advect the point with approximately 75 to 80 percent of the wind component normal to it.

In the case of a slow moving, closed low at the 850-hPa level, the 0°C isotherm will move eastward with respect to the closed low as cold air is advected around the low.

MOVEMENT OF THE 850-hPa 0°C WET-BULB ISOTHERM.— The wet-bulb temperature can be forecast by the above procedures and rules. Remember that the wet-bulb temperature is dependent upon dewpoint as well as the temperature. The dewpoint will be advected with the winds at nearly the full velocity, whereas the temperature under nonsaturated conditions moves slower. The following observations with respect to the 0°C wet-bulb isotherm *after* saturation is reached may help:

- The 0°C wet-bulb isotherm does not move far offshore in the Gulf and the Atlantic because of upward vertical motion in the cold air over the warmer water.

- If the 0°C wet-bulb isotherm lies in a ribbon of closely packed isotherms, movement is slow.

- Extrapolation works well on troughs and ridges.

After the forecast of the surface and 850-hPa level temperature and dewpoint values are made, you are ready to convert these values to their respective wet-bulb temperature. The following procedures are recommended

- Use figure 4-23, views (A) and (B), to compute the wet-bulb temperatures for the 1,000- and 850-hPa levels, respectively. (The surface chart may be used for the 1,000-hPa level.) Admittedly, the wet-bulb temperatures at just these two levels do not give a complete picture of the actual distribution of moisture and temperature, and error is introduced when values are changing rapidly, but these are values the forecaster can work with and predict with reasonable accuracy.

- Refer to figure 4-24. From the surface wet-bulb temperature at the bottom, go up vertically until you intersect the computed 850-hPa level wet-bulb temperature to the left. This intersection indicates the form of precipitation that can be expected. A necessary assumption for use of this graph is that the wet-bulb temperatures at these two levels can be predicted with reasonable accuracy. Known factors affecting the wet-bulb temperature at any particular station should be carefully considered before entering the graph. Some of the known factors are elevation, proximity to warm bodies of water, known layers of warm air above or below the 850-hPa level, etc. Area "A" on the graph calls for a rain forecast, area "B" for a freezing rain forecast, and area "C" for a snow forecast. Area "D" is not so clear cut because it is an overlap portion of the graph; however, wet snow or rain and snow mixed predominate in this area. Sleet occurring by itself for more than 1 or 2 hours is rare, and should be forecast with caution.

FORECASTING THE AREA OF MAXIMUM SNOWFALL

The intent of this section is to introduce the patterns associated with maximum snowfall and to present techniques for predicting the areas where snowstorms are likely to appear.

Synoptic Types

There are four distinct types of synoptic patterns with associated maximum snow area.

BLIZZARD TYPE.— The synoptic situation features an occluding low. In the majority of cases, the "wrapped around" high pressure and ridges are present. The track of the low is north of 40°N, and its speed, which initially may be average or about 25 knots, decreases into the slow category during the occluding process. In practically all cases, a cold closed low at the 500-hPa level is present and captures the surface low in 24 to 36 hours.

The area of maximum snowfall lies to the left of the track. At any particular position, the area is located from due north to west of the low-pressure center. When this type occurs on the east coast with its large temperature contrast and high moisture availability, heavy snowfall may occur. The western edge of the maximum area is limited by the 700-hPa level trough, or low center, and the end of all snow occurs with the passage of the 500-hPa level trough or low center.

MAJOR STORM AND NONOCCLUDING LOWS.— The synoptic situation consists of a nonoccluding wave-type low. The track of the low or wave is south of 40° latitude, and its speed is at least the average of 25 knots, often falling into the fast-moving category. The upper-air picture is one of fast-moving troughs, generally open, but on occasion could have a

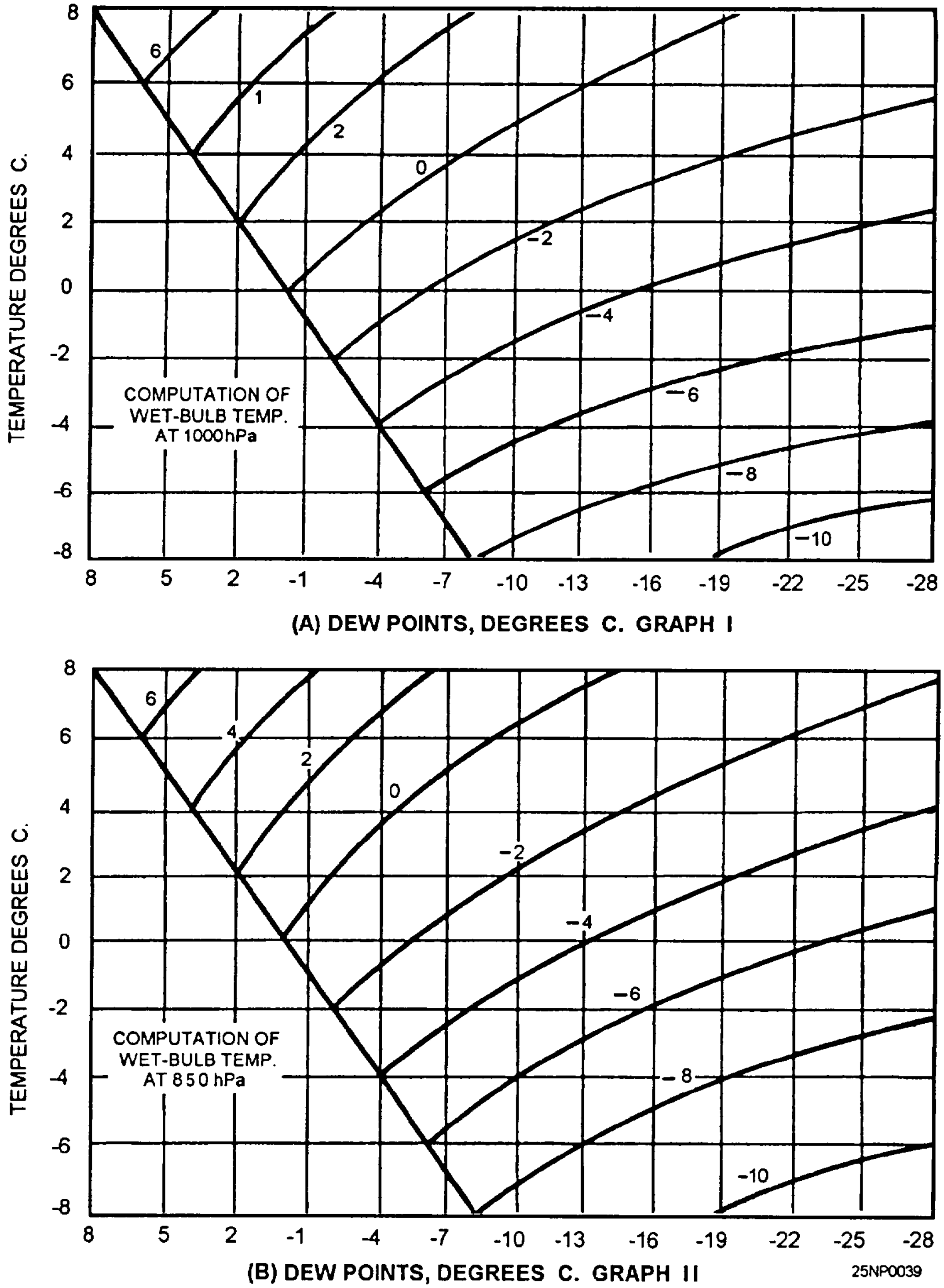

Figure 4-23.-Graphs for computing wet-bulb temperatures. (A) Computation of 1,000-hPa level wet-bulb temperature; (B) computation of 850-hPa level wet-buib temperature.

minor closed center for one or two maps in the bottom of the trough.

The area of maximum snowfall lies in the cold air to the left of the track and usually describes a narrow belt oriented east-west or northeast-southwest about 100 to 200 miles wide. At any position, the area is parallel to the warm front and north of the low-pressure center. Within the maximum area, the rate of snowfall is variable from one case to another, depending upon available moisture, amount of vertical shear, etc. However, it is not uncommon for heavy snow (1 inch

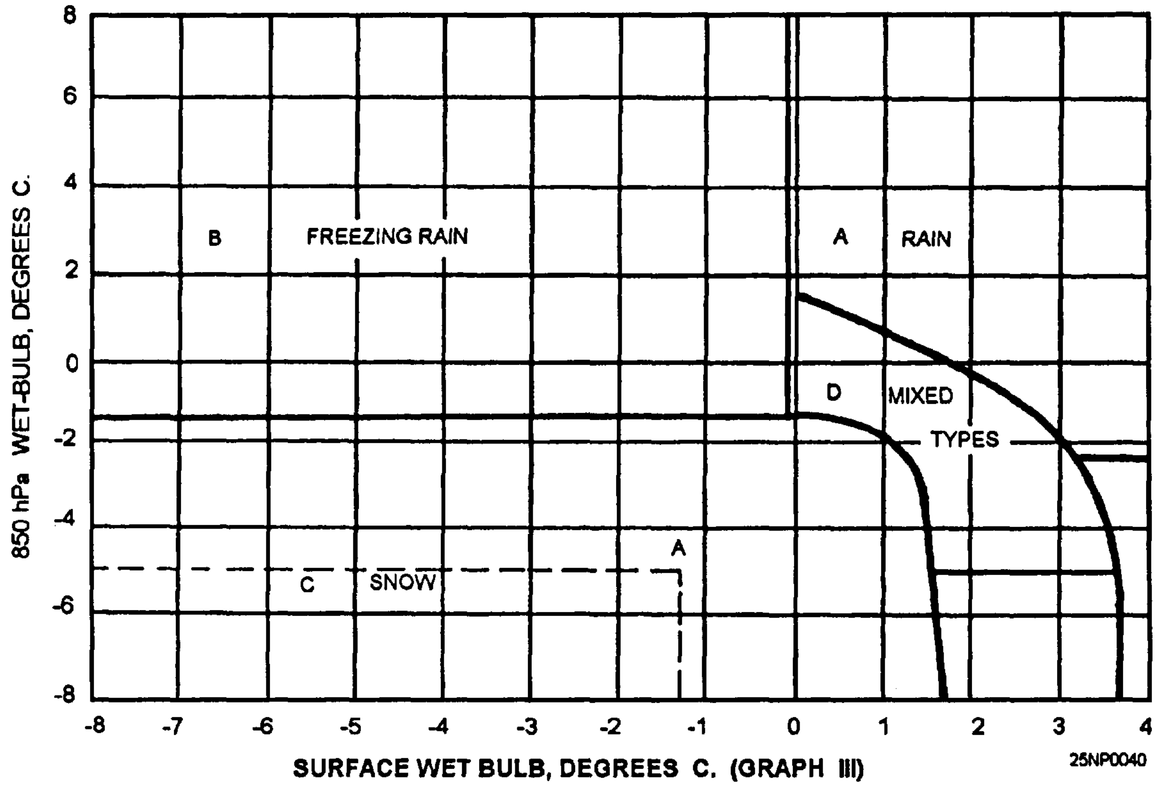

Figure 4-24.-Graph for delineating the form of precipitation.

per hour or greater) to occur. You must remember that even though heavy snow occurs, the duration is short, This means that a location could lie in the maximum snow area only 4 to 8 hours; whereas, in the case of the blizzard type, it usually remains in the area in excess of 10 hours.

WARM ADVECTION TYPE.— This type was separated from the other types because of the absence of an active low in the vicinity of the maximum snow area. A blocking high or ridge is located ahead of a sharp warm front. The overrunning warm air is a steady current from the south to southwest. The area of maximum snowfall is a narrow band parallel to the warm front and moves north or northeast. A rate of fall of moderate to heavy for a 6- to 12-hour duration may occur. The usual history is a transition to freezing rain, and then rain.

POST-COLD FRONTAL TYPE.— The synoptic situation consists of a sharp cold front oriented nearly north-south in a deep trough. A minor wave may form on the front and rapidly travel to the north or northeast, Strong cold advection from the surface to the 850-hPa level is present west of the front. The troughs at the 700-hPa and 500-hPa levels are sharp and displaced to the west of the surface trough 200 to 300 miles. Ample moisture is available at the 850-hPa and 700-hPa levels. This type of heavy snow may occur once or twice per season.

The area of maximum snowfall is located between the 85MPa and 700-hPa troughs, where moisture at both levels is available, The rate of fall is moderate, although for a brief period of an hour or less, it may be heavy. The duration is short, on the order of 2 to 4 hours at anyone location. The area as a whole generates and dissipates in a 12- to 18-hour period. The normal history is one of a general area of light snow within the first 200 miles of a strong outbreak of cold air. After the cold air moves far enough south and the cold front becomes oriented more north to south and begins moving steadily eastward, the troughs aloft and moisture distribution reach an ideal state, and a maximum snow area appears. After 12 to 18 hours, the advection of dry air at the 700-hPa level decreases the rate or fall in the area, and soon thereafter, the area, as a whole, dissipates.

Locating the Area of Maximum Snowfall

The various parameters and characteristics that may be of benefit in locating areas of maximum snowfall are discussed in the following text.

TEMPERATURE.— The 0°C (-3°C east coast) isotherm at the 850-hPa level is used as a basis for snow-rain areas, This isotherm should be carefully analyzed by using all data at 850 hpa. It should then be checked against the surface chart. Keep in mind the following two points:

1. In areas of precipitation, locations reporting snow should lie on the cold side of the 0°C (–3°C east coast) isotherm; for locations reporting mixed types of precipitation (e.g., rain and snow, sleet and snow), the 0°C (–3°C) isotherm will lie very close to or through the location.

2. In areas of no precipitation, the 0°C (–3°C east coast) isotherm will roughly parallel the 32°F isotherm at the surface. In cloudy areas, the separation will be small, and in clear areas, the separation will be larger.

At the 850-hPa level, the 0°C wet-bulb temperature should be sketched in, particularly in the area where precipitation may be anticipated within the next 12 to 24 hours. This line will serve as the first approximation of the future position of the 0°C isotherm.

MOISTURE.— At the 850-hPa level, the –5°C dewpoint line is used as the basic defining line; at the 700-hPa level, the –10°C dewpoint line is used as the basic defining line. The area at 850 hpa that lies within the overlap of the 0°C isotherm and the –5°C dewpoint line is the first approximation of the maximum snowfall area. All locations within this area have temperatures less than 0°C and spreads of 5°C or less. This area is further refined by superimposing the sketched –10°C dewpoint line at 700 hPa upon the area. Now the final area is defined by the 0°C isotherm and the overlapped minimum dewpoint lines from both levels. This final area becomes the area where moderate or heavy snow will be reported, depending upon the particular synoptic situation. See figure 4-25,

MOVEMENT.— The first basic rule for moving the area of maximum snowfall is that it maintains the same relative position to the other synoptic features of the 850-hPa level and surface charts. However, in order to forecast the expansion or contraction of the area, it is necessary to forecast the lines that define it. The 0°C isotherm should be forecast according to the roles set forth in the section treating this particular phase. The moisture lines may be advected with the winds. The 0°C isotherm should also be moved with rules stated previously in this chapter.

The area of maximum snowfall can be forecast for 12 hours with considerable accuracy, and for 24 hours with fair accuracy, provided a reasonable amount of care is exercised according to rules and subjective ideas mentioned previously.

TEMPERATURE

LEARNING OBJECTIVES: Analyze synoptic features in determining temperature forecasts.

Temperature ranks among the most important forecast elements. Temperatures are not only important in the planning and execution of operational exercises, but also are of keen interest to all of us in everyday life.

FACTORS AFFECTING TEMPERATURES

Many factors are involved in the forecasting of temperatures. These factors include air mass characteristics, frontal positions and movement, amount and type of cloudiness, season, nature and position of pressure systems, and local conditions.

Temperature, which is subject to marked changes from day to night, is not considered a conservative property of an air mass. Too, it does not always have a uniform lapse rate from the surface up through the atmosphere. This means that the surface air temperature will not be representative because of the existence of inversions, which may be a condition particularly prevalent at night. Usually, the noonday surface air temperature is fairly representative,

Let's look at factors that cause temperature variations. These factors include insolation and terrestrial radiation, lapse rate, advection, vertical heat transport, and evaporation and condensation.

Insolation and Radiation

In forecasting temperatures, insolation and terrestrial radiation are two very important factors. Low latitudes, for instance, receive more heat during the day than stations at high latitudes. More daytime heat can be expected in the summer months than in the winter months, since during the summer months the sun's rays are more direct and reach the earth for a longer period of time. Normally, there is a net gain of heat during the day and a net loss at night. Consequently, the maximum temperature is usually reached during the day, and the minimum at night. Cloudiness will affect insolation and terrestrial radiation. Temperature forecasts must be made only after the amount of cloudiness is determined. Clouds reduce insolation and terrestrial radiation,

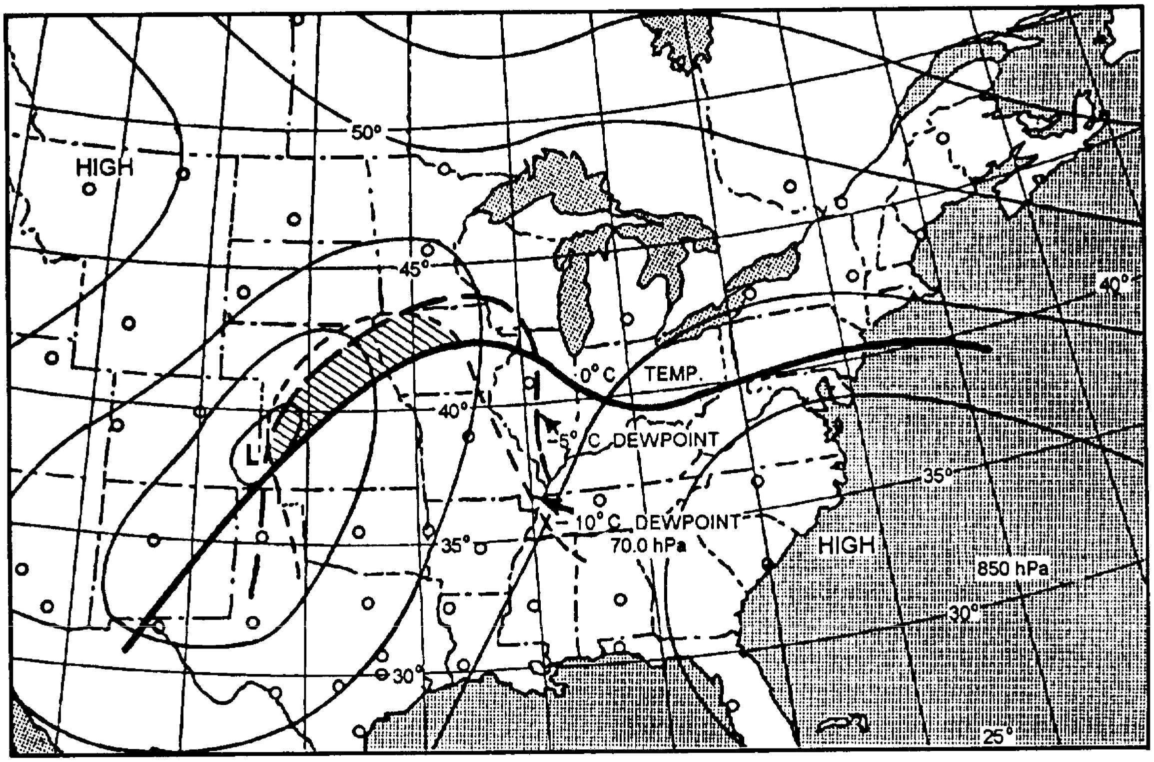

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Figure 4-25.-Illustration of the location of the maximum snow area. The low center moved to Iowa in 24 hours, and the maximum snow area spread northeast along the area 50 to 75 miles either side of a line through Minneapolis to Houghton, Michigan.

causing daytime temperature readings to be relatively lower than normally expected, and nighttime temperatures to be relatively higher. The stability of the lapse rate has a marked effect on insolation and terrestrial radiation. With a stable lapse rate, there is less vertical extent to heat; therefore, surface heating takes place more rapidly. With an unstable lapse rate, the opposite is true. If there is an inversion, there is less cooling, since the surface temperature is lower than that of the inversion layer; that is, at some point the energy radiated by the surface is balanced by that radiated by the inversion layer.

Advection

One of the biggest factors affecting temperature is the advection of air. Advection is particularly marked in its effect on temperature with frontal passage. If a frontal passage is expected during the forecast period, the temperature must be considered. Advection within an air mass may also be important. This is particularly true of sea and land breezes and mountain breezes. They affect the maximum and minimum temperatures and their time of occurrence.

Vertical Heat Transport

Vertical heat transport is a temperature factor. It is considerably affected by the windspeed. With strong wind there is less heating and cooling than with light wind or a calm because the heat energy gained or lost is distributed through a deeper layer when the turbulence is greater.

Evaporation and Condensation

Evaporation and condensation affect the temperature of an air mass. When cool rainfalls through a warmer air mass, evaporation takes place, taking heat from the air. This often affects the maximum temperature on a summer day on which afternoon thundershowers occur. The temperature may be affected at the surface by condensation to a small extent during fog formation, raising the temperature a degree or so because of the latent heat of condensation.

FORECASTING SPECIAL SITUATIONS

The surface and aloft situations that are indicative of the onset of cold waves and heat waves are discussed in the following text.

Cold Wave

A forecast of a cold wave gives warning of an impending severe change to much colder temperatures. In the United States, it is defined as a net temperature drop of 20°F or more in 24 hours to a prescribed minimum that varies with geographical location and time of the year. Some of the prerequisites for a cold wave over the United States are continental Polar, or Arctic air with temperatures below average over west central Canada, movement of a low eastward from the Continental Divide that ushers in the cold wave, and large pressure tendencies on the order of 3 to 4 hPa occurring behind the cold front. Aloft, a ridge of high pressure develops over the western portion of the United States or just off the coast. An increase in intensity of the southwesterly flow over the eastern Pacific frequently precedes the intensifying of the ridge. Frequently, retrogression of the long waves takes place. In any case, strong northerly to northwesterly flow is established aloft and sets the continental Polar or Arctic air mass in motion. When two polar outbreaks rapidly follow one another, the second outbreak usually moves faster and overspreads the Central States. It also penetrates farther southward than the first cold wave. In such cases, the resistance of the southerly winds ahead of the second front is shallow. At middle and upper levels, winds remain west to northwest, and the long wave trough is situated near 80° west.

Most cold waves do not persist. Temperatures moderate after about 48 hours. Sometimes, however, the upper ridge over the western portion of the United States and the trough over the eastern portion are quasi-stationary, and a large supply of very cold air remains in Canada. Then, we experience successive outbreaks with northwest steering that hold temperatures well below normal for as long as 2 weeks.

Heat Wave

In the summer months, heat wave forecasts furnish a warning that very unpleasant conditions are impending. The definition of a heat wave varies from place to place. For example, in the Chicago area, a heat wave is said to exist when the temperature rise above 90°F on 3 successive days. In addition, there are many summer days that do not quite reach this requirement, but are highly unpleasant because of humidity.

Heat waves develop over the Midwestern and eastern portions of the United States when along wave trough stagnates over the Rockies or the Plains states, and along wave ridge lies over or just off the east coast. The belt of westerlies are centered far to the north in Canada. At the surface we observe a sluggish and poorly organized low-pressure system over the Great Plains or Rocky Mountains. Pressure usually is above normal over the South Atlantic, and frequently the Middle Atlantic states. An exception occurs when the amplitude of the long wave pattern aloft becomes very great. Then, several anticyclonic centers may develop in the eastern ridge, both at upper levels and at the surface. Frequently, they are seen first at 500 hPa. Between these highs we see formation of east-west shear lines situated in the vicinity of 38° to 40°N. North of this line winds blow from the northeast and bring cool air from the Hudson Bay into the northern part of the United States. A general heat wave continues until the long wave train begins to move.

SUMMARY

In this chapter we discussed condensation and precipitation producing processes. Following a discussion on condensation and precipitation producing processes, we then covered condensation and precipitation dissipation processes. Forecasting of frontal clouds and weather was then discussed, including the topics of frontal cloudiness and precipitation, air mass cloudiness and precipitation, vertical motion and weather, vorticity and precipitation, and middle clouds in relation to the jetstream. We then covered short-range extrapolation techniques, which included use of the nephanalysis, frontal precipitation, lowering of ceilings in continuous rain areas, the trend chart as an aid, and the time-liner as an aid. A discussion of cloud layer analysis and forecasting was then presented along with the importance of RAOB use in cloud analysis and identification, the humidity field, a 500-hPa level analysis of the dewpoint depression, a three-dimensional analysis of the moist layer, precipitation and clouds, and cirrus indications. A discussion of the prediction of snow versus rain followed. Topics presented were geographical and seasonal considerations, the physical nature of the problem, general synoptic considerations, forecasting techniques, and areas of maximum snowfall. The last topics of discussion were factors affecting temperature, and the forecasting of temperatures during special situations.