CHAPTER 5

FORECASTING SEVERE WEATHER FEATURES

The paramount responsibilities of the forecaster include providing forecasts of severe weather conditions and timely warnings to aircraft and ships to ensure the safety of their operations, as well as the safety of their personnel.

This chapter discusses some of these phenomena and methods that may be used to forecast severe weather conditions.

THUNDERSTORMS

LEARNING OBJECTIVES: Recognize phenomena associated with thunderstorm activity. Forecast the movement and intensity of thunderstorms.

The thunderstorm represents one of the most formidable weather hazards in temperate and tropical zones. The turbulence, high winds, heavy rain, and occasionally hail that accompany thunderstorms are a definite threat to the safety of flight and to the security of naval installations. It is important that the forecaster be acquainted with the structure of thunderstorms and their associated weather, as well as the knowledge to accurately predict their formation and movement.

The *AG2* TRAMAN, volume 1, covered thunderstorm formation and movement. Therefore, in this chapter, we will discuss, in more detail, the weather phenomena associated with thunderstorms and various methods of forecasting their intensity and movement.

THUNDERSTORM TURBULENCE AND WEATHER

Tlmnderstorms are characterized by turbulence, moderate to extreme updrafts and downdrafts, hail, icing, lightning, precipitation, and, under most severe conditions (in certain areas), tornados.

Turbulence (Drafts and Gusts)

Downdrafts and updrafts are currents of air that may be continuous over many thousands of feet in the vertical, and horizontally as large as the extent of the thunderstorm. The velocity of the downdrafts and updrafts is relatively constant as contrasted to gusts. Gusts are primarily responsible for the bumpiness (turbulence) normally encountered in cumuliform clouds. A downdraft or updraft maybe compared to a river flowing at a fairly constant rate, whereas a gust is comparable to an eddy or other type of random motion of water in a river.

Studies of the structure of the thunderstorm cell indicate that during the cumulus stage of development, the updrafts may cover a horizontal area as large as 6 miles. In the cumulus stage, the updraft may extend from below the cloud base to the cloud top, a height greater than 25,000 feet. During the mature stage, the updrafts cease in the lower levels of the cloud, although they continue in the upper levels where cloud tops may exceed 60,000 feet. These drafts are of considerable importance in aviation because of the change in altitude that may occur when an aircraft flies through them.

In general, the maximum number of high velocity gusts are found at altitudes of 5,000 to 10,000 feet below the top of the thunderstorm cloud, while the least severe turbulence is encountered near the base of the thunderstorm. The characteristic response of an aircraft intercepting a series of gusts is a number of sharp accelerations or “bumps” without an accompanying change in altitude. The degree of bumpiness or turbulence experienced in flight is related to both the number of such abrupt changes encountered in a given distance and the strength of the individual changes.

Hail

Hail is regarded as one of the worst hazards of flying in thunderstorms. It usually occurs during the mature stage of cells that have updrafts of more than average intensity, and is found with the greatest frequency between 10,000 and 15,000 feet. As a general rule, the greater the vertical extent of the thunderstorm, the more likely hail will occur.

Although encounters by aircraft with large hail are not too common, hail can severely damage an aircraft in a very few seconds. The general conclusion regarding hail is that most midlatitude storms contain hail sometime during their cycle. Most hail will occur during the mature stage. In subtropical and tropical thunderstorms, hail seldom reaches the ground. It is generally believed that these thunderstorms contain less hail aloft than do midlatitude storms.

Rain

Thunderstorms contain considerable quantities of moisture that may or may not be falling to the ground as rain. These water droplets may be suspended in, or moving with, the updrafts. Rain is encountered below the freezing level in almost all penetrations of fully developed thunderstorms. Above the freezing level, however, there is a sharp decline in the frequency of rain.

There seems to be a definite correlation between turbulence and precipitation. The intensity of turbulence, in most cases, varies directly with the intensity of precipitation. This relationship indicates that most rain or snow in thunderstorms is held aloft by updrafts.

Icing

Where the air temperatures are at or below freezing, icing should be expected in flights through thunderstorms. In general, icing is associated with temperatures from 0° to –20°C. Most severe icing occurs from 0°C to –10°C. The heaviest icing conditions usually occur in that region above the freezing level where the cloud droplets have not yet turned to ice crystals. When the thunderstorm is in the cumulus stage, severe icing may occur at any point above the freezing level. However, because of the formation of ice crystals at high levels and the removal of liquid water by precipitation, icing conditions are usually somewhat less in the mature and dissipating stage.

THUNDERSTORM ELECTRICITY AND LIGHTNING

The thunderstorm changes the normal electrical field, in which the earth is negative with respect to the air above it, by making the upper portion of the thunderstorm cloud positive and the lower portion negative. This negative charge then induces a positive charge on the ground. The distribution of the electric charges in a typical thunderstorm is shown in figure 5-1. The lightning first occurs between the upper positive charge area and the negative charge area immediately below it. Lightning discharges are considered to occur most frequently in the area roughly bounded by the 0°C and the -9°C temperature levels. However, this does not mean that all discharges are confined to this region, for as the thunderstorm develops, lightning discharges may occur in other areas, and from cloud to cloud, as well as cloud to ground. Lightning can do considerable damage to aircraft, especially to radio equipment.

THUNDERSTORMS IN RELATION TO THE WIND FIELD

During all stages of a cell, air is being brought into the cloud through the sides of the cloud. This process is known as *entrainment.* A cell entrains air at a rate of 100 percent per 500 hPa; that is, it doubles its mass in an ascent of 500 hPa. The factor of entrainment is important in establishing a lapse rate within the cloud that is greater than the moist adiabatic rate, and in maintaining the downdraft.

When there is a marked increase with height in the horizontal wind speed, the mature stage of the cell may be prolonged. In addition, the increasing speed of the wind with height produces considerable tilt to the updraft of the cell, and in fact, to the visible cloud itself. Thus, the falling precipitation passes through only a small section of the rising air; it falls thereafter through the relatively still air next to the updraft, perhaps even outside the cell boundary. Therefore, since the drag of the falling water is not imposed on the rising air currents within the thunderstorm cell, the updraft can continue until its source of energy is exhausted. Tilting of the thunderstorm explains why hail is sometimes encountered in a cloudless area just ahead of the storm.

RADAR DETECTION

Radar, either surface or airborne, is the best aid in detecting thunderstorms and charting their movements. A thunderstorm's size, direction of movement, shape and height, as well as other significant features, can be determined from a radar presentation. Radar interpretation is mentioned in chapter 12 of this manual, and for a more detailed discussion, refer to the *Federal Meteorological Handbook No. 7,* Part B, and *the Federal Meteorological Handbook No. 11,* Part B.

THUNDERSTORM FLIGHT HAZARDS

Thunderstorms are often accompanied by extreme fluctuations in ceiling and visibility. Every thunderstorm has turbulence, sustained updrafts and downdrafts, precipitation, and lightning. Icing conditions, though quite localized, are quite common in thunderstorms, and many contain hail. The flight

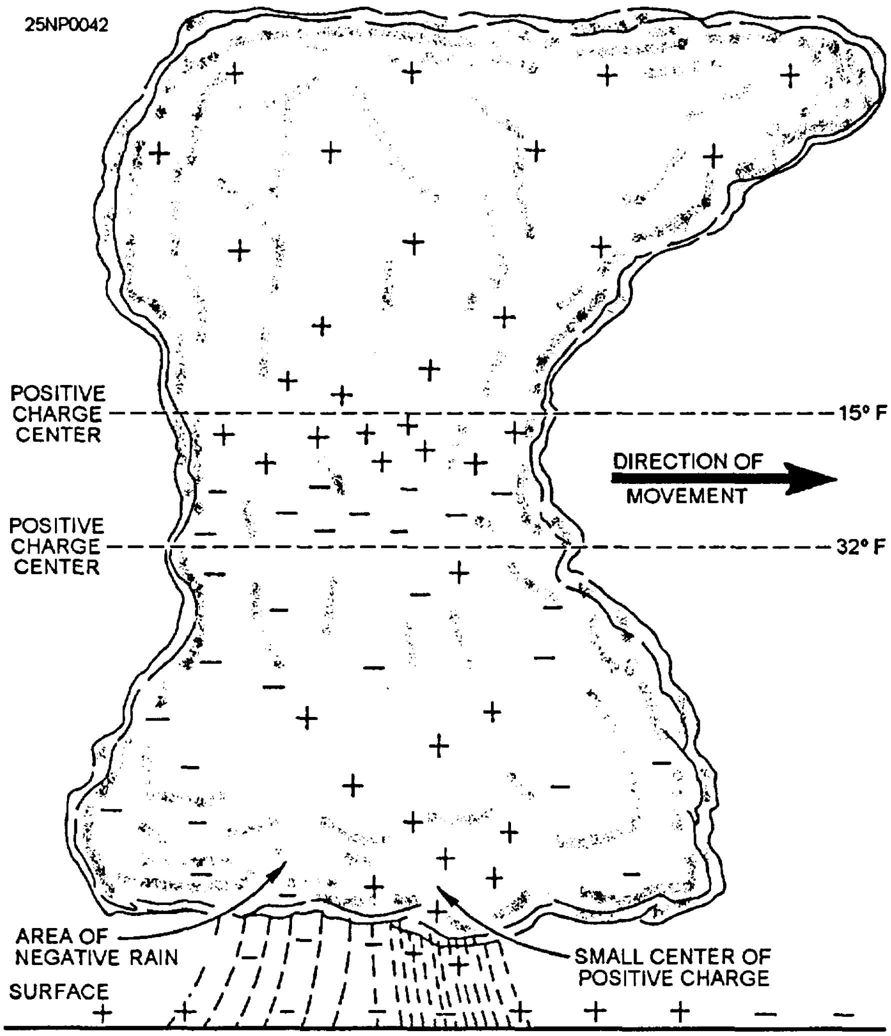

Figure 5-1.-Location of electric charges inside a typical thunderstorm cell.

conditions listed below are generally representative of many (but not necessarily all) thunderstorms.

- The chance of severe or extreme turbulence within thunderstorms is greatest at higher altitudes, with most cases of severe and extreme turbulence about 8,000 to 15,000 feet above ground level (AGL). The least turbulence may be expected when flying at or just below the base of the main thunderstorm cloud. (The latter rule would not be true over rough terrain or in mountainous areas where strong eddy currents produced by strong surface winds would extend the turbulence up to a higher level.)

- The heaviest turbulence is closely associated with the areas of heaviest rain.

- The strongest updrafts are found at heights of about 10,000 feet AGL or more; in extreme cases, updrafts in excess of 65 feet per second occur. Downdrafts are less severe, but downdrafts on the order of 20 feet per second are quite common.

- The probability of lightning strikes occurring is greatest near or slightly above the freezing level.

Because of the potential hazards of flying in a thunderstorm, it is obviously nothing short of folly for pilots to attempt to fly in thunderstorms, unless operationally necessary.

THUNDERSTORM SURFACE PHENOMENA

The rapid change in wind direction and speed immediately before thunderstorm passage is a significant surface hazard associated with thunderstorm activity. The strong winds that accompany thunderstorm passage are the result of horizontal spreading of downdraft currents from within the storm as it approaches the ground.

Figure 5-2 shows the nature of the wind outflow and indicates how it is formed from the settling dome of cold air that accompanies the rain core during the mature stage of the thunderstorm. The arrival of this outflow results in a radical and abrupt change in the wind speed and direction. It is an important consideration for aircraft that are landing or taking off.

Wind speeds at the leading edge of the thunderstorm are ordinarily far greater than those at the trailing edge. The initial wind surge observed at the surface is known as the first gust. The speed of the first gust is normally the highest recorded during thunderstorm passage, and it may vary as much as 180 degrees in direction from the surface wind direction that previously existed. The mass of cooled air spreads out from downdrafts of neighboring thunderstorms (especially in squall lines), and often becomes organized into a small, high-pressure area called a *bubble high,* which persists for some time as an entity that can sometimes be seen on the surface chart. These highs may be a mechanism for controlling the direction in which new cells form.

The speed of the thunderstorm winds depends upon a number of factors, but local surface winds reaching 50 to 75 miles per hour for a short time are not uncommon. Because thunderstorm winds can extend several miles in advance of the thunderstorm itself, the thunderstorm wind is a highly important consideration for pilots preparing to land or take off in advance of a storm's arrival. Also, many thunderstorm winds are strong enough to do considerable structural damage, and are capable of overturning or otherwise damaging even medium-sized aircraft that are parked and not adequately secured.

The outflow of air ahead of the thunderstorm sets up considerable low-level turbulence. Over relatively level ground, most of the significant turbulence associated with the outrush of air is within a few hundred feet of the ground, but it extends to progressively higher levels as the roughness of the terrain increases.

THUNDERSTORM ALTIMETRY

During the passage of a thunderstorm, rapid and marked surface pressure variations generally occur. These variations usually occur in a particular sequence characterized as follows.

- An abrupt fall in surface pressure as the storm approaches.
- An abrupt rise in surface pressure associated with rain showers as the storm moves overhead (often associated with the first gust).
- A gradual return to normal surface pressure after thunderstorm passage, and the rain ceases.

Such pressure changes may result in significant pressure altitude errors on landing.

Of greater concern to the pilot are pressure altitude readings that are too high. If a pilot used an altimeter setting computed during the maximum pressure, and then landed after the pressure had fallen, the altimeter still could read 60 feet or more above the true altitude after landing.

Here is where you, as a forecaster, can make certain that timely and accurate altimeter settings are furnished to the tower for transmission to pilots during

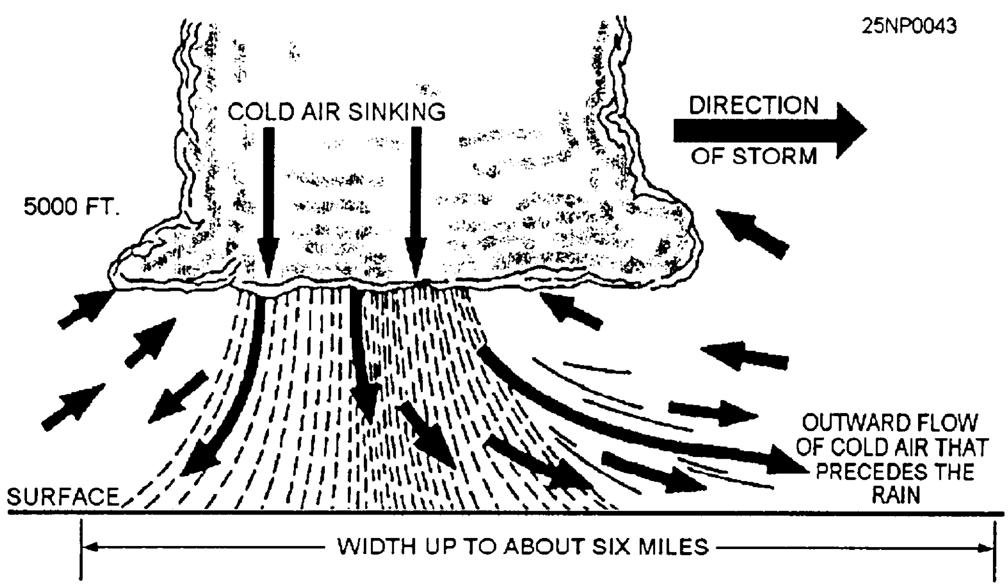

Figure 5-2.-Cold dome of air beneath a thunderstorm cell in the mature stage. Arrows represent deviation of windflow. Dashed lines indicate rainfall.

thunderstorm conditions. If the data are old and inaccurate, an aircraft mishap could result.

THUNDERSTORM FORECASTING

The standard method of forecasting air mass thunderstorms has long consisted primarily of an analysis of rawinsonde data with particular emphasis on the so-called "positive areas."

Many times conditions are favorable for thunderstorm development, with a large positive energy area showing up on the sounding, with no ensuing thunderstorm activity. At other times, thunderstorms may occur when they are not forecasted. Clearly, factors other than instability are important, and, at times, of overriding importance.

A number of thunderstorm forecasting methods have been developed, but many of these are beyond the scope of this manual. The forecasting of convective clouds by using variations of the parcel method are covered in this section. Further, a method for the prediction of these storms that enables the forecaster to arrive at a fairly accurate, reasonably objective forecast will be discussed.

For a more detailed discussion of the determinations of instability, stability, the convective condensation level (CCL), the level of free convection (LFC), and the lifting condensation level (LCL), refer to the *AG2* TRAMAN, volume 2.

THE PARCEL METHOD

The temperature of a minute parcel of air is assumed to change adiabatically as the parcel is displaced vertically from its original position. If, after vertical displacement, the parcel has a higher virtual temperature than the surrounding atmosphere, the parcel is subjected to a positive buoyancy force and will be further accelerated upwards; conversely, if its virtual temperature has become lower than that of the surrounding air, the parcel will be denser, and eventually return to its initial or equilibrium position,

Formation of Clouds by Heating From Below

The first step is to determine the convection temperature or the surface temperature that must be reached to start the formation of convection clouds by solar heating of the surface air layer. The procedure is to first determine the CCL on the plotted sounding and, from the CCL point on the T curve, proceed downward along the dry adiabat to the surface pressure isobar. The temperature read at this intersection is the convection temperature.

Figure 5-3 shows an illustration of forecasting afternoon convective cloudiness from a plotted sounding. The dewpoint curve was not plotted to avoid confusion. The dashed line with arrowheads indicates the path the parcel of air would follow under these conditions. You can see that the sounding was modified at various times during the day.

To determine the possibility of thunderstorms by the use of this method and from an analysis of the sounding, the following conditions must exist:

- Sufficient heating must occur.
- The positive area must exceed the negative area. The greater the excess, the greater the possibility of thunderstorms.
- The parcel must rise to the ice crystal level. Generally, this level should be –10°C and below.
- There must be sufficient moisture in the lower troposphere. This is the most important single factor in thunderstorm formation.
- Climatic and seasonal conditions should be favorable.
- A weak inversion (or none at all) should be present in the lower levels.
- An approximate height of the cloud top may be determined by assuming that the top of the cloud will extend beyond the top of the positive area by a distance equal to one-third of the height of the positive area.

Formation of Clouds by Mechanical Lifting

When using this method, it is assumed that the type lifting will be either orographic or frontal, Here we will be concerned with the Lifting Condensation Level (LCL) and the Level of Free Convection (LFC).

The LCL is the height at which a parcel of air becomes saturated when it is lifted dry adiabatically. The LCL for a surface parcel is always found at, or below, the CCL. The LFC is the height at which a parcel of air is lifted dry adiabatically until saturated, and thereafter would first become warmer than the surrounding air. The parcel will then continue to rise freely above this level until it becomes colder than the surrounding air.

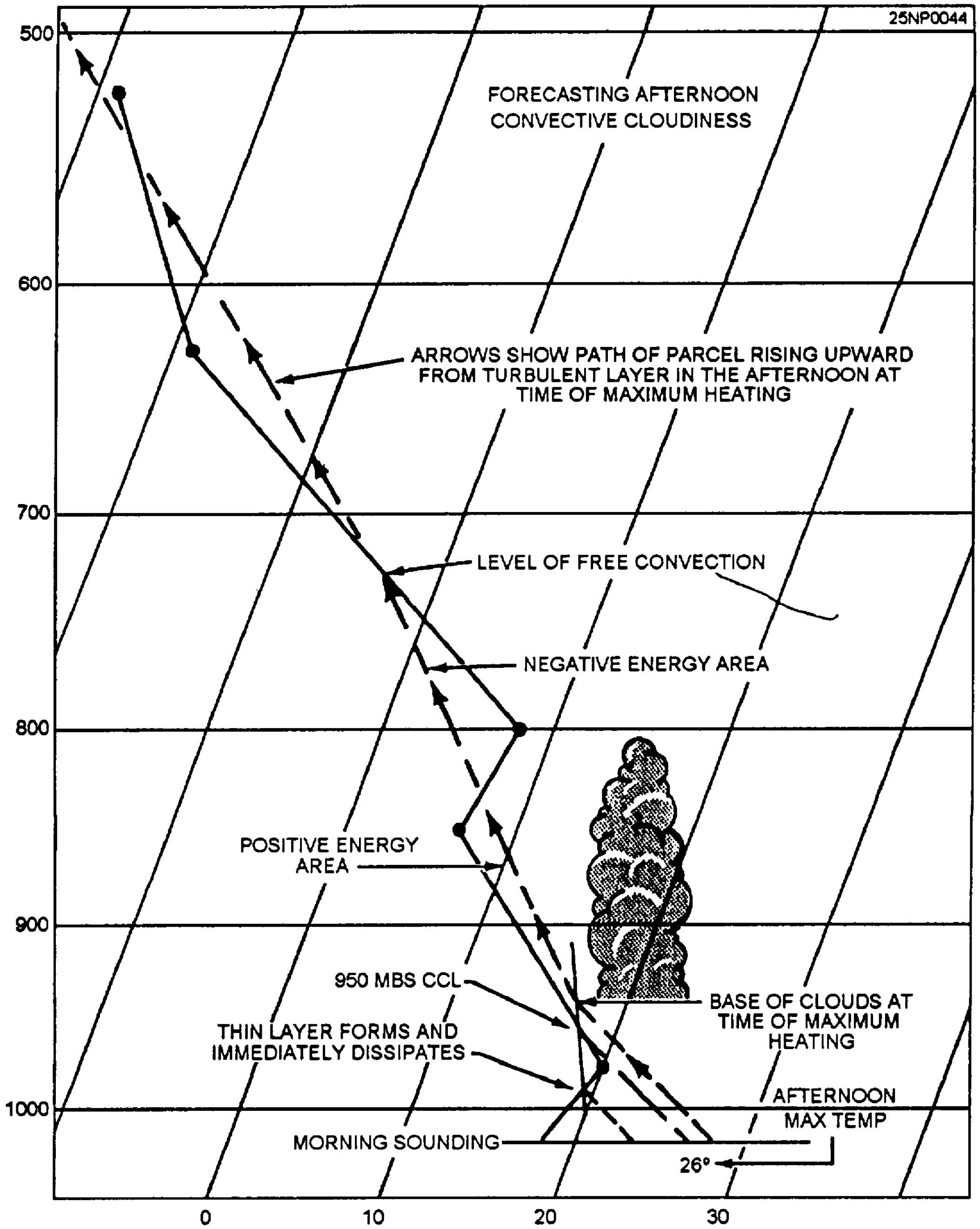

Figure 5-3.-Forecasting afternoon convective cloudiness.

Figure 5-4 shows the formation of clouds due to mechanical lifting. This figure shows the formation of a stratified layer of clouds above the LCL, and to the LFC. At the LFC and above, the clouds would be turbulent. The tops of the clouds extend beyond the top of the positive area due to overshooting, just as clouds formed due to heating.

To determine the possibility of thunderstorms from this method, the following conditions should be met:

- The positive area must exceed the negative area; the greater the excess, the greater the possibility of thunderstorms,

- There must be sufficient lifting for the parcel to reach the LFC, The frontal slope or the orographic barriers can be used to determine how much lifting can be expected,

- The parcel must reach the ice crystal level (–10°C and below).

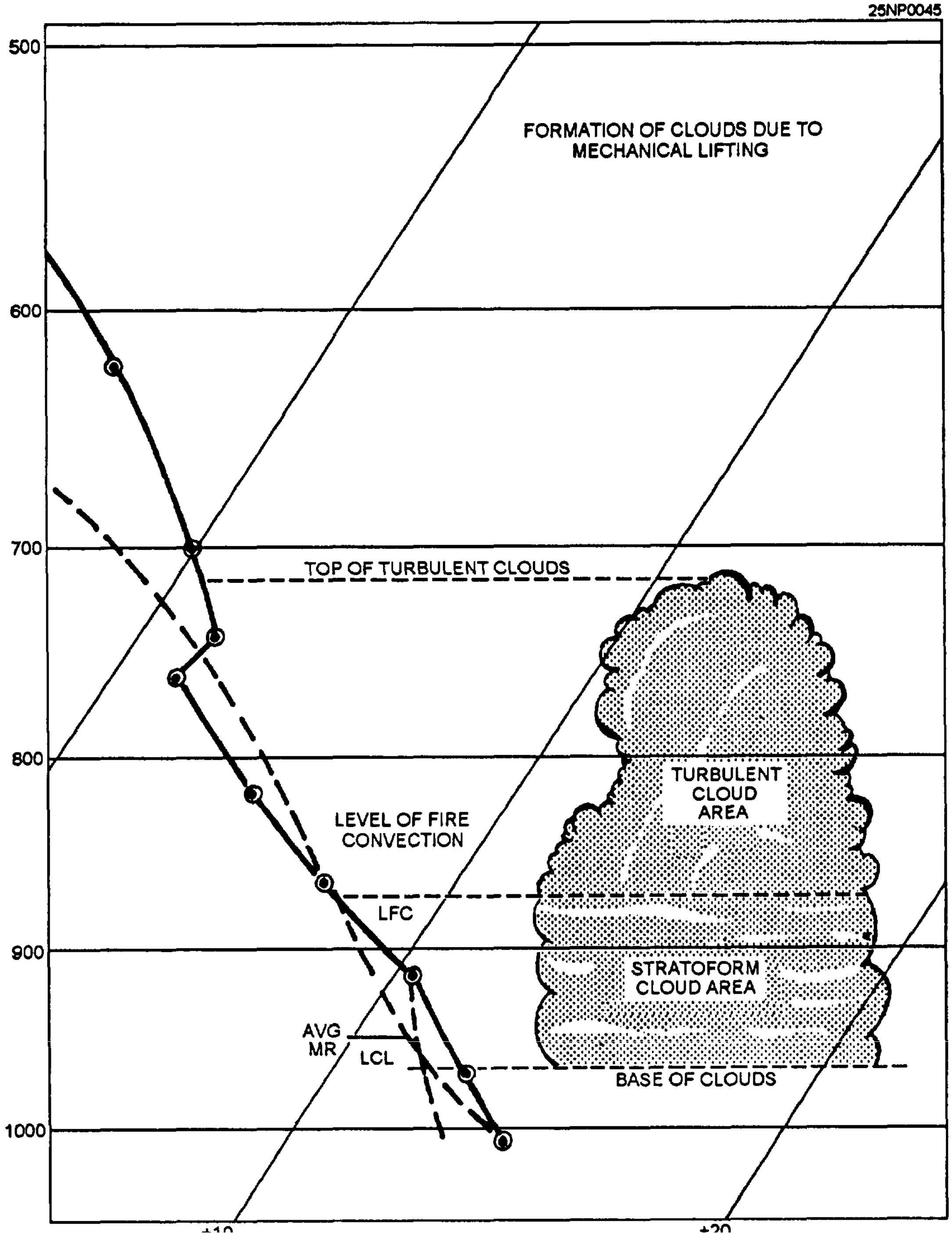

Figure 5-4.-Formation of clouds due to mechanical lifting.

- Even though the positive area does not exceed the negative area, cloudiness occurs after the parcel passes the LCL, and precipitation may occur after the parcel passes the ice crystal level.

One main advantage of this method is that it can be done quickly and with relative ease. A major disadvantage is that it assumes that the parcel does not change its environment, and that it overestimates or underestimates the stability and instability conditions.

INSTABILITY INDICATIONS FROM THE WET-BULB CURVE

A layer of the atmosphere is potentially unstable if the potential wet-bulb temperature decreases with altitude. Potential instability refers to a layer that is lifted as a whole. The wet-bulb temperature may be found by lifting each individual point on the sounding dry adiabatically to saturation, and then back to its original level moist adiabatically, By connecting the points on a sounding, a wet-bulb curve can be constructed.

If the wet-bulb curve slopes to the right with increasing altitude, the potential wet-bulb temperature increases with height, and the layer is potentially stable. If it slopes to the left with increasing height, more than the saturation adiabats, the layer is potentially unstable. If none of the potential curves intersect the sounding, thunderstorms are not likely to occur.

AIR-MASS THUNDERSTORMS

This method is a reasonably simple method for forecasting air-mass thunderstorms in the Eastern United States. It does require prediction of short-range changes in the vertical distribution of temperature and moisture.

- The first consideration is to eliminate those areas whose soundings disclosed one or more of the following moisture inadequacies:

- The Dewpoint depression is 13°C or more at any level from 850 through 700 hPa.

- The Dewpoint depression sum is 28°C or more at 700- and 600-hPa levels.

- There is dry or cool advection at low levels.

- The surface dewpoint is 60°F or less at 0730 local with no substantial increase expected before early afternoon.

- There is a lapse rate of 21°C or less from 850 to 500 hPa.

- There is a freezing level below 12,000 feet in an unstable cyclonic flow, producing only light showers.

After eliminating all soundings that meet one or more of the previous six conditions, you should use the following parameters to make the forecast.

- The lapse rate between 850 and 500 hPa.

- The sum of the dewpoint depressions at 700 and 600 hPa in degrees C.

NOTE: The lapse rate is the difference in temperature between these two pressure levels. For example, if the temperature at 850 hpa was 15°C and at 500hPa it was –10°C, the difference would equal 25°.

These two computations are used as arguments for the graph in figure 5-5.

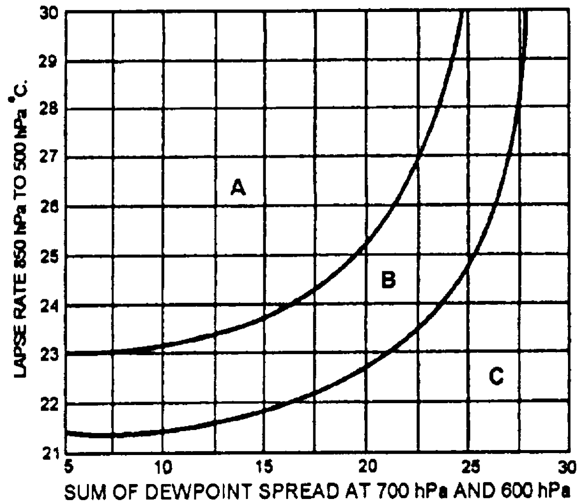

Figure 5-5.-Local area thunderstorm graph. Area "A" is isolated thunderstorms, with a 12 to 1 chance of at least one rain gauge in dense network receiving rain. Area "B" is scattered thunderstorms, with a 4 to 1 chance of reported rain. Area "C" is no rain.

One further condition for the development of thunderstorms is the absence of large anticyclonic wind shear, which is measured at 850 hPa. No horizontal shear at this level may exceed 20 knots in 250 miles measured toward the low-pressure area from the sounding station. Figure 5-6 illustrates how this measurement is made.

STABILITY INDEXES AS AN INDICATION OF INSTABILITY

The overall stability or instability of a rawinsonde sounding is sometimes conveniently expressed in the form of a single numerical value called the stability index. Such indexes have been introduced mainly as aids in connection with particular forecasting techniques. Most of the indexes take the form of a difference in temperature, dewpoint, wet-bulb temperature, or potential temperature in height or pressure between two arbitrarily chosen surfaces. These indexes are generally useful only when combined, either objectively or subjectively, with other data and synoptic considerations. Used alone, they are less valuable than when plotted on stability index charts and analyzed for large areas. In this respect they have the value of alerting the forecaster to those soundings, routes, or areas that should be more closely examined by other procedures.

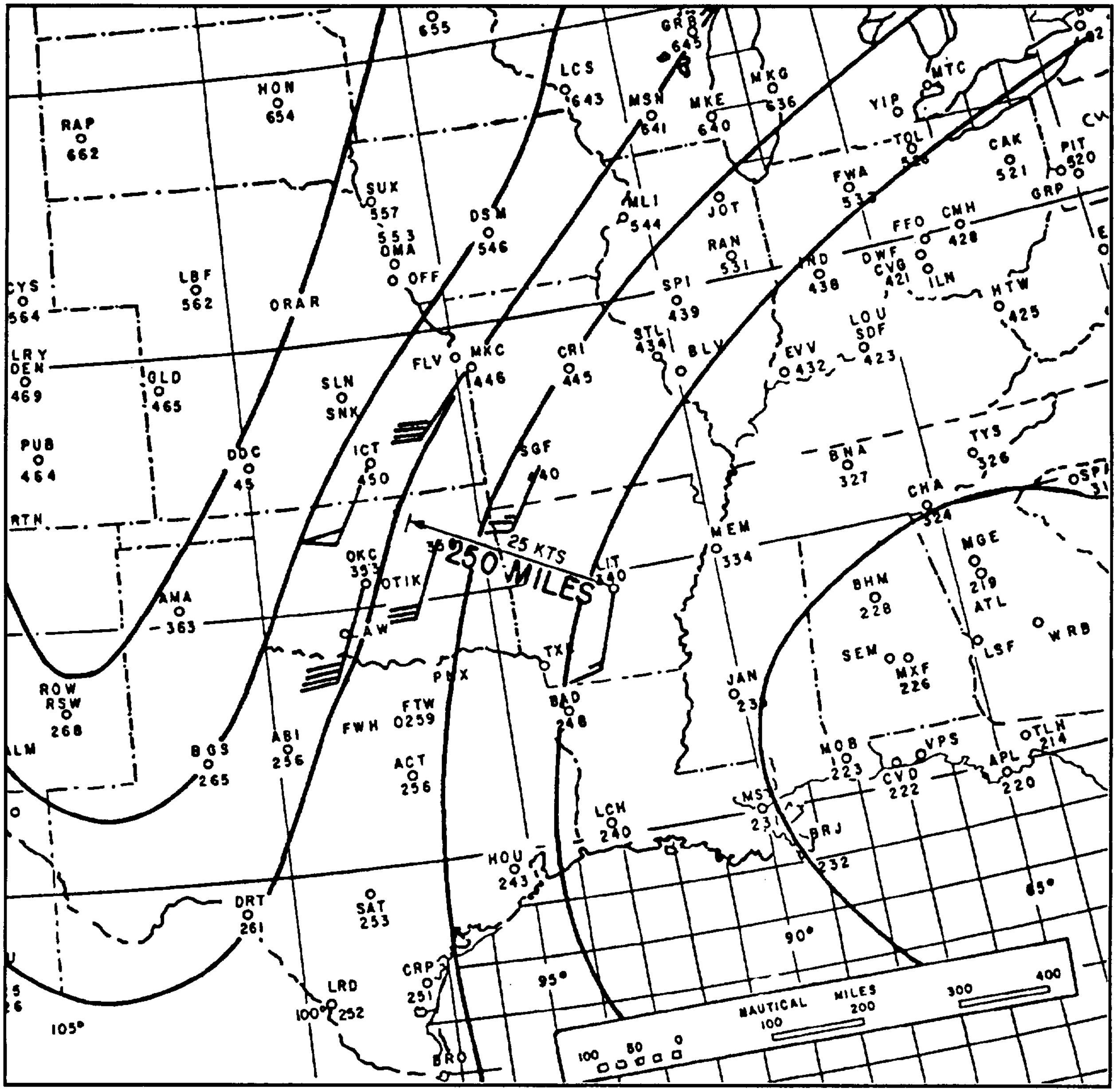

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Figure 5-6.-Example of anticyclonic shear and curvature at 850 hPa preventing thunderstorms, with otherwise favorable air mass conditions present in the Little Rock area.

There are a number of methods that maybe used for determining stability or instability. Among these methods are the Showalter Stability Index (SSI), the Lifted Index (LI), the Fawbush-Miller Stability Index (FMI), and the Martin Index (MI). Only the SSI method will be discussed in this TRAMAN.

The National Weather Service currently uses the (LI) method for producing their facsimile products. All current methods are discussed in detail in NAVAIR 50- IP-5, *Use of the Skew T, Log P Diagram in Analysis and Forecasting.*

For more information on the SSI, the forecasting of peak wind gusts, and the forecasting of hail, refer to the *AG2* TRAMAN, volume 2, unit 6. These subjects are covered in an abbreviated form in this chapter.

The SSI is the most widely used of the various types of indexes.

Figure 5-7 shows the computation of the SSI.

For forecasting purposes, the significance of the index values to forecasting is as follows:

- When the index is +3°C or less, showers are probable, and some thunderstorms may be expected in the area.
- The chance of thunderstorms increases rapidly for index values in the range of +1°C to -2°C.
- Index values of -3°C or less are associated with severe thunderstorms.
- When the value of the index is below –6°C, the forecaster should consider the possibility of tornado occurrence. However, the forecasting value of all index categories must, in each case, be evaluated in the light of the moisture content of the air and of other synoptic conditions.

FORECASTING MOVEMENT OF THUNDERSTORMS

Radar can be an invaluable aid in determining the speed and the direction of movement of the thunderstorm. Sometimes it is desirable and necessary to estimate the movement from winds aloft. There is no completely reliable relationship between speed or direction of the winds aloft in forecasting thunderstorm movement. However, one study reveals that there is a marked tendency for large convective rainstorms to move to the right of the wind direction in the mean cloud layer from 850 to 500 hPa (or the 700-hPa wind) with a deviation of about 25 degrees. There appears to be little correlation between wind speed and speed of movement at any level, although the same study mentioned above revealed that 82 percent of the storms move within plus or minus 10 knots of a mean speed of 32 knots.

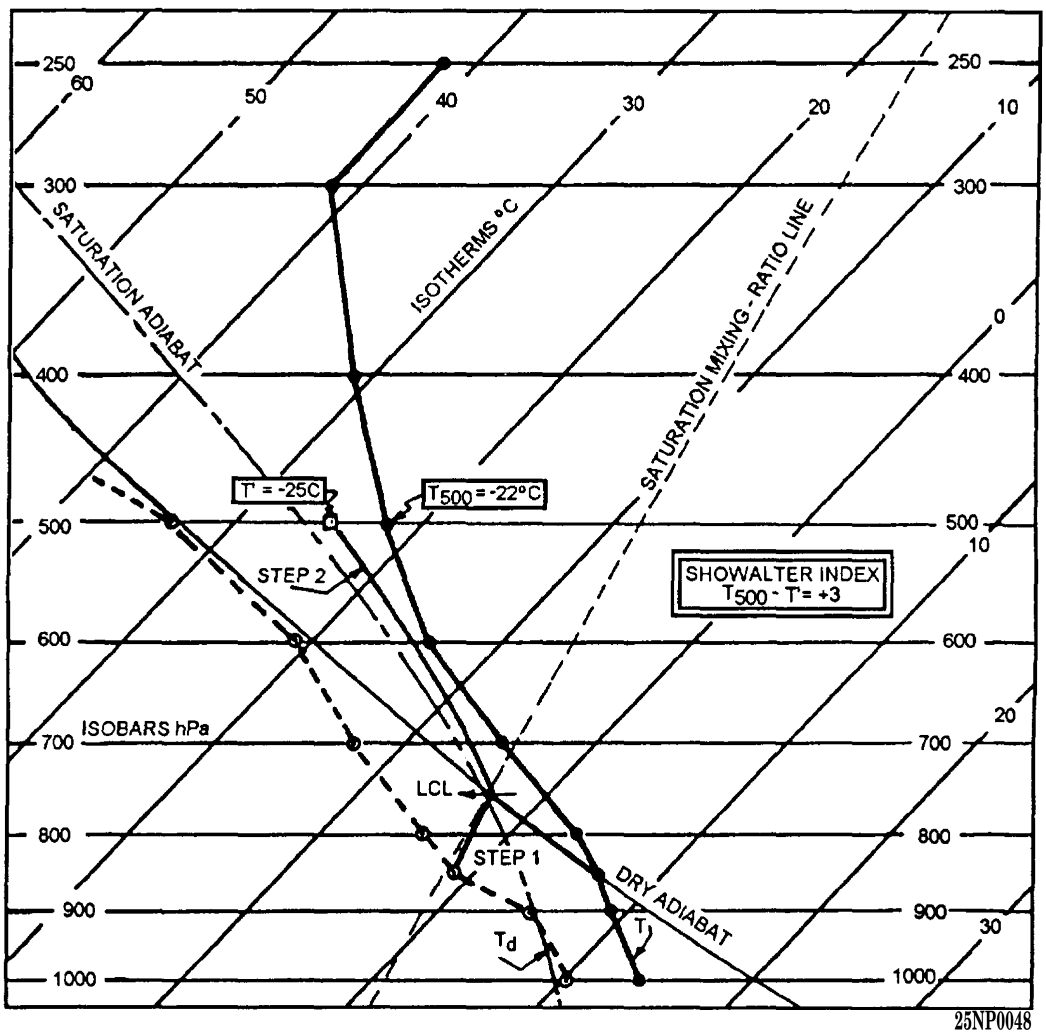

Figure 5-7.-Computation of the Showalter Stability Index (SSI).

FORECASTING MAXIMUM GUSTS WITH NONFRONTAL THUNDERSTORMS

Maximum gusts associated with thunderstorms occur over a very small portion of the area in which the thunderstorm exists, and usually occur immediately before the storm's passage. Nevertheless, the possibility of damage to aircraft and installations on the surface is so great that every available means should be used to make the best and most accurate forecast possible to forewarn the agencies concerned.

Estimation of Gusts From Climatology and Storm Intensity

The forecaster is aware that the season of the year and the station location have a great bearing upon the maximum winds to be expected. Certain areas of the United States, and the world, have a history of severe thunderstorm occurrence with associated strong winds during the most favorable seasons of the year. For this reason you should have the thunderstorm climatology for your station, as well as for the general area, available as to time of occurrence, season of occurrence, and the associated conditions.

The storm's intensity and reports from neighboring stations can give you a good indication of conditions to expect at your station.

Forecasting Peak Wind Gusts Using the USAF Method

Refer to the AG2 TRAMAN, volume 2, unit 6, for an explanation of two methods for forecasting maximum wind gusts of convective origin. One involves the use of the Dry Stability Index (T_1) and the other the downrush temperature subtracted from the dry-bulb temperature (T_2).

FORECASTING HAIL

Hail, like the maximum wind gusts in thunderstorms, usually takes place in a narrow shaft that is seldom wider than a mile or 2 and usually less than a mile wide. The occurrence of hail in thunderstorms was discussed earlier in this chapter. However, a few additional facts concerning the occurrence and frequency of hail in flight should be discussed at this point.

- Since hail is normally associated with thunderstorms, the season of the maximum occurrence of hail is coincident with the season of maximum occurrence of thunderstorms.

- When the storm is large and well developed, an assumption should be made that it contains hail.

- Encounters of hail below 10,000 feet show the hail distribution to the equally divided between the clear air alongside the thunderstorm, in the rain area beneath the storm, and within the thunderstorm itself.

- From 10,000 to 20,000 feet, the percentages range from 40 percent in the clear air alongside the storm to 60 percent in the storm, with 82 percent of the encounters outside the storm beneath the overhanging cloud.

- Above 20,000 feet, the percentages reflect 80 percent of the hail is encountered in the storm with 20 percent in the clear air beneath the anvil or other cloud extending from the storm.

Climatology is of vital importance in predicting hail occurrence, as well as hail size. Good estimations of the size of hail can be gleaned from reports of the storm passage over nearby upstream stations. Here too, modifying influences must be taken into account.

Hail Frequency as Related to Storm Intensity and Height

Fifty percent hail frequencies may be expected in storms exceeding 46,000 feet, based on radar reports. With a maximum echo height of 52,000 feet, a 67 percent hail frequency can be expected; and only 33 percent at 35,00 feet, Mean echo heights are 42,000 feet for hail and 36,000 feet for rain.

A Yes-No Hail Forecasting Technique

The technique presented in this section of the chapter is an objective method of hail forecasting. It uses the parameters of the ratio of cloud depth below the freezing level and height of the freezing level. The data used in this study were derived from 70 severe convective storms (34 hail-producing and 36 nonhail-producing storms) over the Midwestern States. Severe thunderstorms, as used in the development of this technique, were defined as those thunderstorms causing measurable property damage due to strong winds, lightning, or heavy rain. Severe thunderstorms with accompanying hail were defined as those thunderstorms accompanied by hail where hail was listed as the prime cause of property damage, even though other phenomena may also have occurred. All tornadoes were excluded from consideration to avoid confusion.

METHOD OF ANALYSIS.— Plot representative upper air soundings (0000 UTC and 1200 UTC) on the Skew T diagram. Then analyze the following parameters:

- Convective condensation level (CCL).

- Equilibrium level (EL). The EL is found at the top of the positive area on the sounding where the temperature curve and the saturation adiabat through the CCL again intersect. This gives a measure of the extent of the cloud's vertical development, and thus an estimate of its top or maximum height.

- Freezing level. This is defined as the height of the zero degree isotherm.

NOTE: All three heights are expressed in units of hectopascals (hPa).

Next determine the following two parameters:

1. The ratio of the cloud depth below the freezing level (distance in hectopascals from the CCL to the freezing level) to the cloud's estimated vertical development (distance from the CCL to the EL in hectopascals) is defined as the cloud depth ratio. For example, if the CCL was at 760 hpa, the freezing level at 620 hPa, and the EL at 220 hpa, then the cloud depth ratio would be computed as follows:

$$\text{Cloud depth ration} = \frac{140}{540} = .26$$

2. Height of the freezing level in hectopascals (620 hpa).

With step 1 as the vertical axis and step 2 as the horizontal axis, use figure 5-8 for occurrence or nonoccurrence of hail. In this case, hail would be forecast if the value falls in the hail forecast area.

EVALUATION.— Using the data in 70 dependent cases, the correct percent for prediction of hail or no hail was 83 percent. This technique combines the two parameters relating hail to convective activity into a single predictor. Although the data used in this study were from the Midwest, the application need not be confined to that area. With some modification of this diagram, this method could serve as a basis for a local forecasting tool for other areas.

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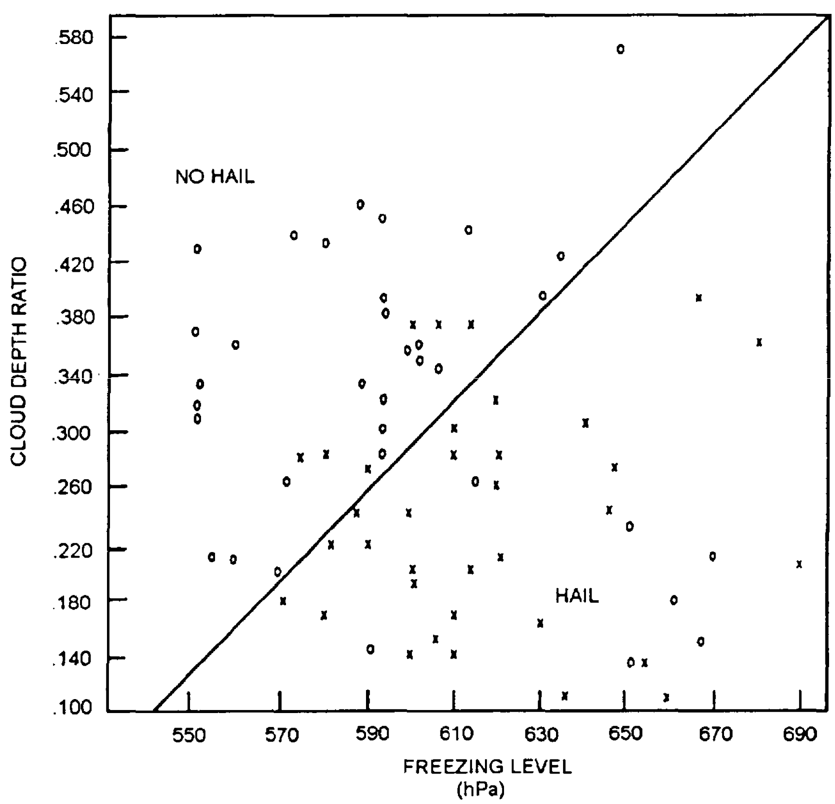

Figure 5-8.-Scattered diagram showing the distribution of selected hail occurrences at certain Midwestern stations during the spring and summer of 1994 and 1995. The freezing level is plotted against the cloud depth ratio. (X = hail reported. 0 = no hail reported.)

Hail Size Forcasting

Once the forecaster has determined that the probability of hail exists, as previously outlined, the next logical question will relate to the size of the hail that may be anticipated. The following text discusses the method that uses the Skew T Log P diagram.

The first step in forecasting hail is to determine the convective condensation level (CCL). This parameter is evaluated on the adiabatic chart by finding the mean mixing ratio in the moist layer of the lowest 150 hPa. and following this saturation, mixing ratio line to its intersection with the sounding dry-bulb temperature curve. Next, the moist adiabat through the CCL is traced up to the pressure level where the dry-bulb temperature is -5°C. This pressure level, the dry-bulb temperature curve, and the moist adiabat through the CCL form a triangle, outlining a positive area. Figure 5-9 illustrates this procedure. The horizontal coordinate in figure 5-9 is the length of the horizontal side of the triangle in degrees Celsius. The length is measured from the pressure at the base of the triangle.

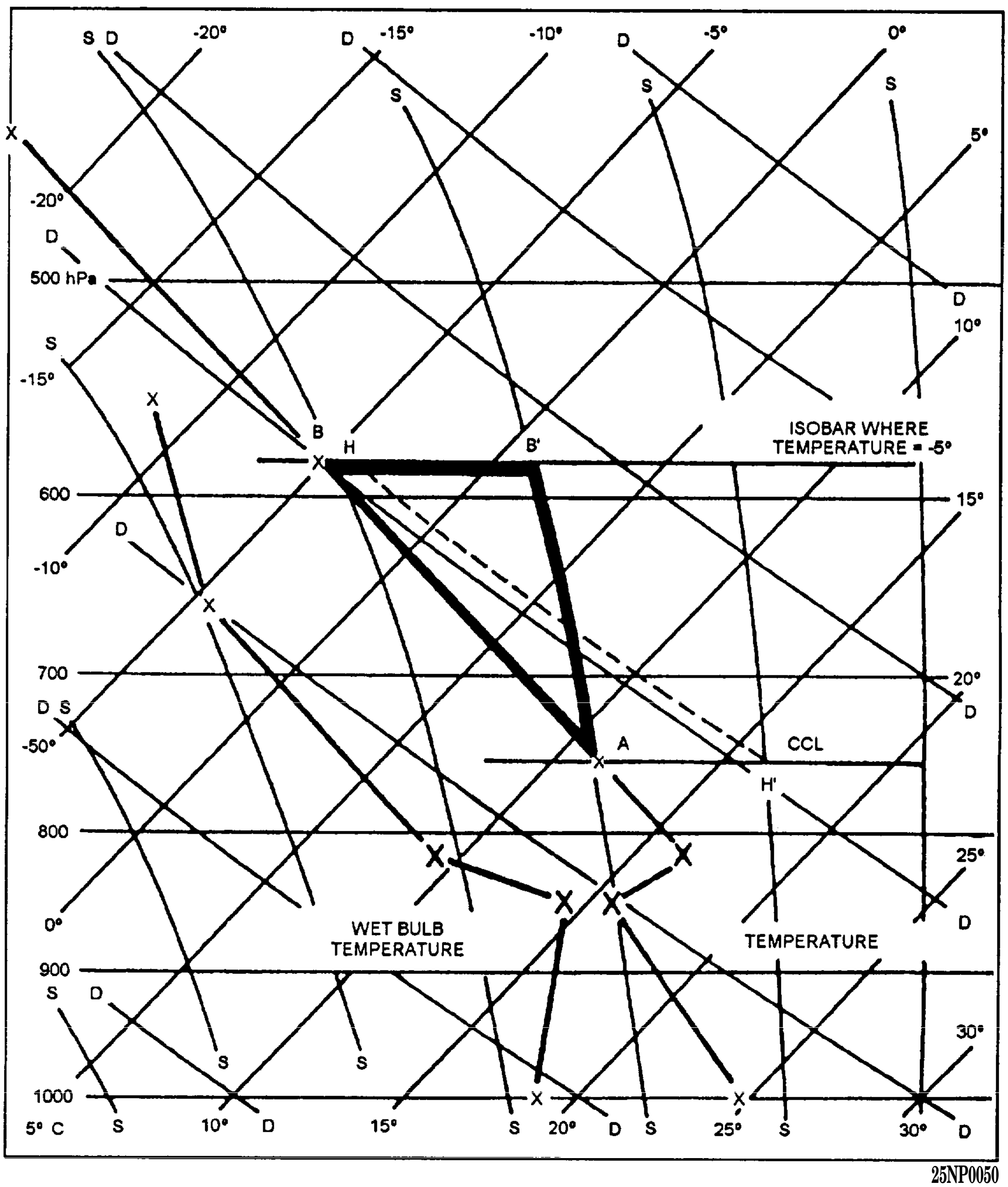

Figure 5-9.-Example of hail size forecast sounding.

These computations for the horizontal length in degrees and altitude in degrees are used on the graph in figure 5-10, and for the forecast of hailstone diameter.

EXAMPLE OF TECHNIQUE.— In the sounding shown in figure 5-9, the CCL is point A. The moist adiabat from the CCL to the pressure level, where the temperature is -5°C, is the line AB´. The isobar from the point where the air temperature is –5°C to its intersection with the moist adiabat is the line BB´. The dry adiabat from the isobar BB´ through the triangle to the pressure of the CCL is the line HH´. The base of the triangle in degrees Celsius is 6°C (from plus 1 to minus 5). The length of the dry adiabat through the triangle is 21°C (from minus 4 to plus 17).

EVALUATION.— The value on the graph in figure 5-10, with a horizontal coordinate of 6 and a vertical coordinate of 21, is a forecast of 1-inch hail.

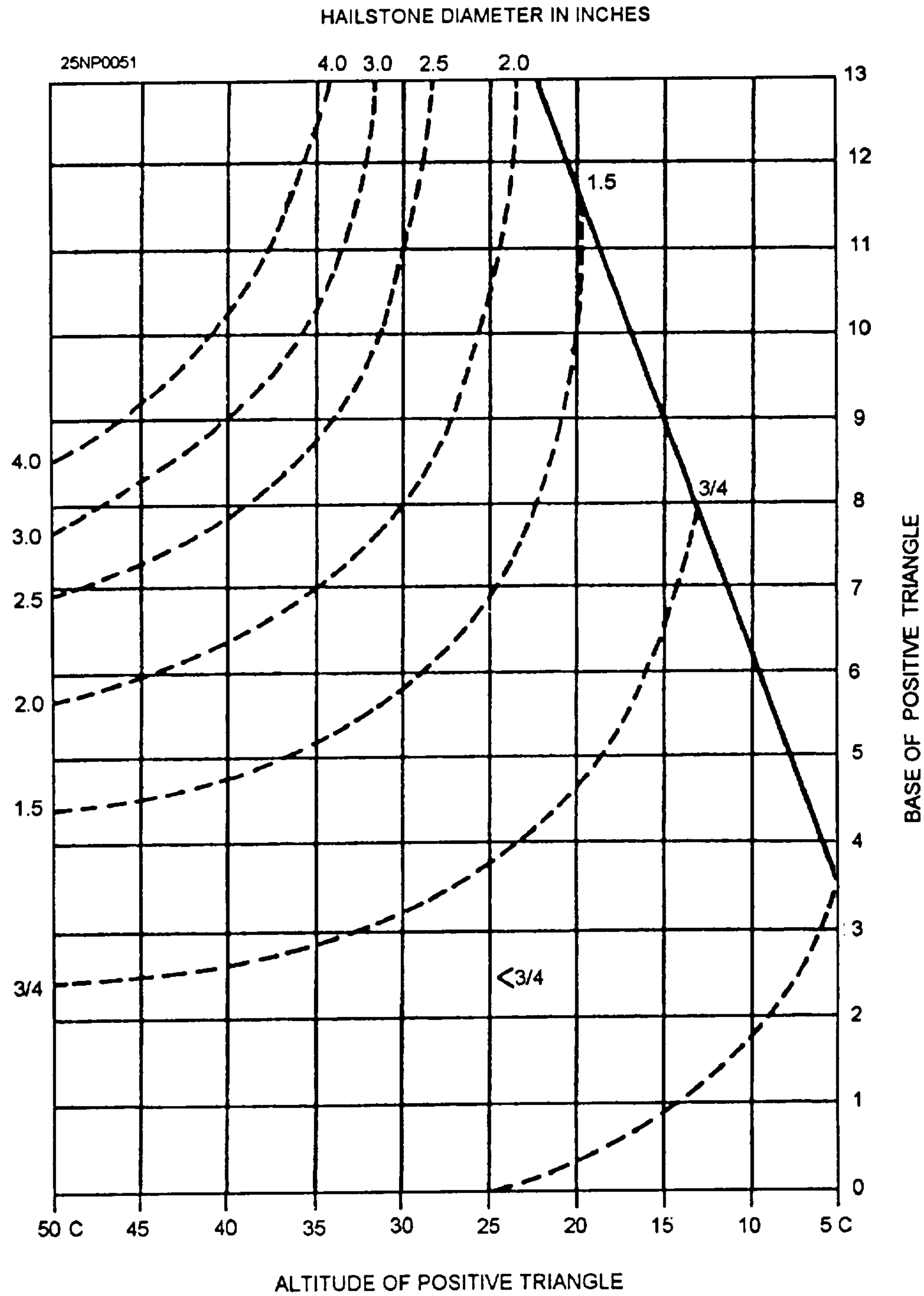

Figure 5-10.-Fawbush-Miller Hail Graph showing forecast hailstone diameter in inches.

To more accurately forecast hail size in conjunction with thunderstorms along the Gulf Coast or in any air mass where the Wet-Bulb-Zero height is above 10,500 feet, it is necessary to refer to the graph in figure 5-11. The hail size derived from figure 5-10 is entered on the horizontal coordinate of figure 5-11, and the corrected hail size read off is compatible with the height of the Wet-Bulb-Zero temperature.

A Thunderstorm Checklist

Regardless of where you are forecasting, the factors and parameters favorable for thunderstorm, hail, and gust forecasting should be systematized into some sort of checklist to determine the likelihood and probability of thunderstorms and the attendant weather. Figure 5-12 is a suggested format and checklist to ensure that all parameters have been given due consideration.

TORNADOES

LEARNING OBJECTIVES: Analyze areas conducive to tornadic activity. Identify the three general types of tornadoes. Recognize the difference between tornadoes and waterspouts.

Tornadoes are violently rotating columns of air extending downward from a cumulonimbus cloud. They are nearly always observed as funnel clouds. Their relatively small size ranks them as second in the severity of the damage they cause, with tropical cyclones ranking first. They occur only in certain areas of the world, and are most frequent in the United States in the area bounded by the Rockies to the west and the Appalachians to the east. Tornadoes also occur during certain preferred seasons of the year in the United States,

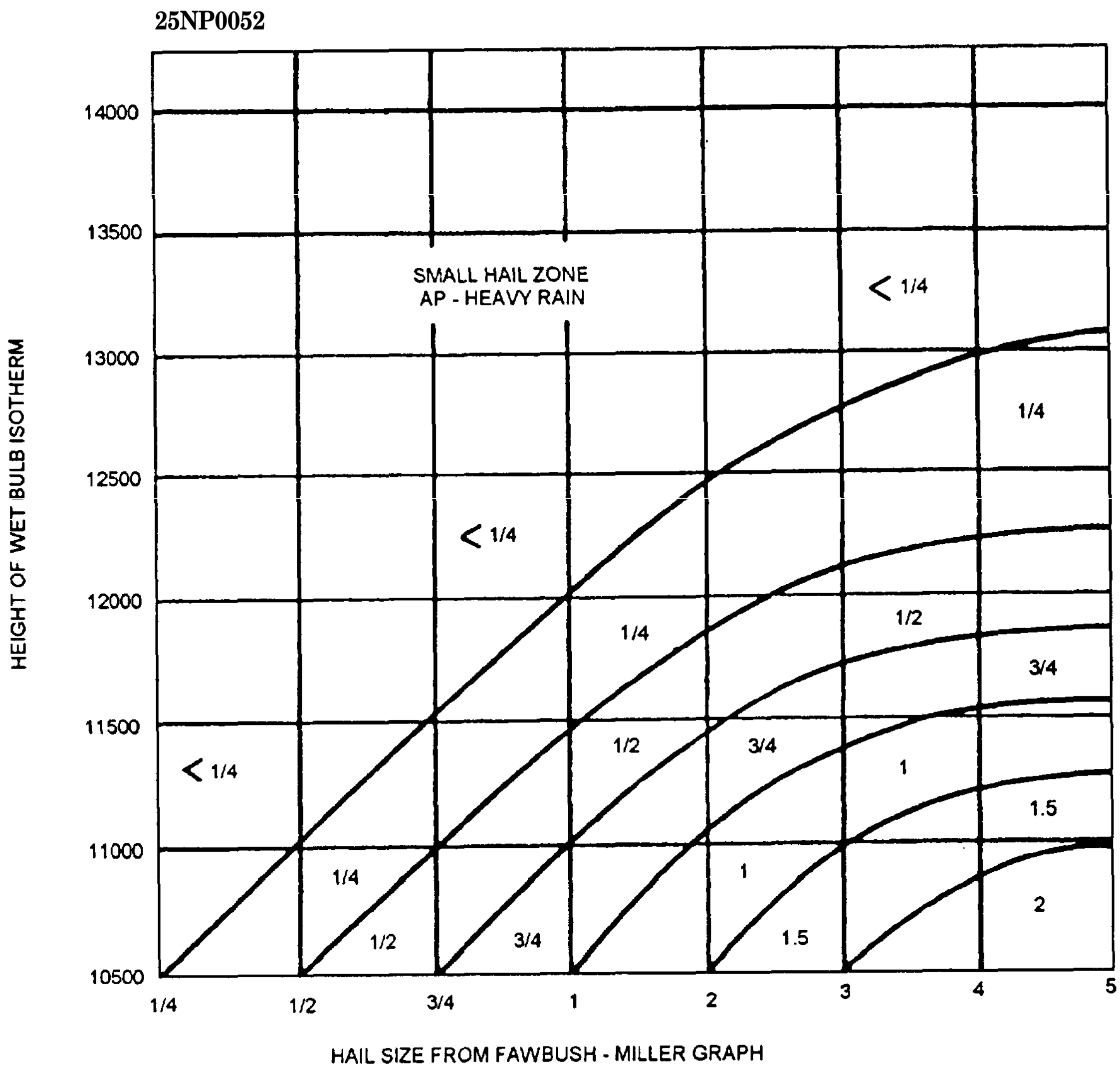

Figure 5-11.-Hail size at surface expected from tropical air mass thunderstorm.

WORKSHEET

SUGGESTED CHECKLIST FOR THE DETERMINATION OF AIR MASS THUNDERSTORM ACTIVITY

I. Analyze the Skew T Logo P Diagram using the closest sounding & parcel methods

1. Determine the following

a.	Convective condensation level (CCL)	5,000'
b.	Temperature necessary for convection	+28°C(82.4°F)
c.	Maximum temperature forecast	+30°C (86°F)
d.	Is temperature necessary for convection expected to be reached?	YES
e.	Inversions present? (Stg/Weak/Mdt/Height)	NONE
f.	Inversions strong enough to prevent or retard convection activity	NO
g.	Positive energy area favorable/unfavorable	FAV
h.	Does positive energy area extend well above the freezing level? (Preferably above the -10° C isotherm)	YES
i.	Does moist layer extend to 10,000 ft?	YES
j.	At least 3 gr/kg moisture 12,000 ft	YES(4.5)
k.	Conditionally unstable air extent to 16,000'	YES
l.	Stability index (fav/unfav)	-1 (FAV)
m.	Were thunderstorms or clouds of vertical development present previous day? If so, has there been any change at lower or upper levels to retard further development today	YES - NO CHANGE
n.	Is month and time of year favorable climatologically for thunderstorm development ?	YES(6.4 PER MO.)

2. Forecast from consideration of a thru n. AFTN TSTMS

Note: The most important single predictor of daytime thunderstorms is moisture in the lower troposphere in the morning.

Determination of f. is of prime consideration. You must decide if heating and lift is sufficient to overcome the inversion.

25NP0053

Figure 5-12.-Suggested checklist for the determination of air mass thunderstorm activity.

25NP0054

II. Forecast from using the Bailey Graph SCTD TSTMS

III. Final forecast from all of the above considerations SCTD AFTN

AIR MASS TSTMS BETWEEN 14-1800 LOCAL

IV. Gusts

1. T_1 Manual Method 40 KT

2. Estimated from reports or expected intensity of storm 38 KT

3. Final Forecast 39 KTS

V. Hail (SURFACE)

1. Yes-No Method YES

2. Fawbush-Miller Method ½-1"
Note if wet bulb zero above 10,500 ft.

3. Estimated ½" 4. Forecast ½"

Figure 5-12.-Suggested checklist for the determination of air mass thunderstorm activity—Continued.

with their most frequent occurrence during May. The season of occurrence varies with the locality. In the United States, 80 percent of the tornadoes have occurred between noon and 2100 local time.

Complete details on the forecasting of these phenomena are beyond the scope of this training manual. You should consult the *U.S. Department of Commerce, Forecasting Guide No. 1, Forecasting Tornadoes and Severe Thunderstorms,* and the many other excellent texts and publications for a complete understanding of this problem. The senior Aerographer's Mate should have a basic understanding of the factors leading to the formation of such severe phenomena, to recognize potential situations, and to be able to forecast such phenomena.

SURFACE THERMAL PATTERNS AS A FORECASTING AID

This method indicates that thermal tongues averaging 50 miles or less in width are favored locations for tornado activity within the area of convective storm activity. These thermal tongues can be located quite readily on the surface synoptic chart. Use the following technique to supplement the existing tornado forecast.

The procedures to locate areas of potentially severe convective activity with reference to the synoptic surface thermal pattern are as follows:

1. Draw isotherms for every 2°C to locate thermal tongues.

2. Within the general area in which convective storms are forecast, locate the axis of all pronounced

thermal tongues oriented nearly parallel to the gradient flow.

3. On this axis, locate the point with the greatest temperature gradient within 50 to 100 miles to the right, and normal to the flow.

4. From this reference point, a rectangle is constructed with its left side along the axis of the thermal tongue by locating comer points on the axis 25 miles upstream and 125 miles downstream from the reference point. The rectangle is 150 miles long and 50 miles wide.

5. This is the forecast area for possible tornado or funnel cloud development.

TORNADO TYPES

There are three distinct tornadic types over the United States. They are the Great Plains type, the Gulf Coast type, and West Coast type.

Great Plains Type

This type of tornado will generally form on the squall line in advance of a fast moving cold front, hence its prediction involves timely forecasting of the squall line formation along, or in advance of, a cold front, upper cold front, or trough. Conditions must favor a downrush of air from aloft.

Gulf Coast Type

In contrast to the air mass type (Great Plains type), tornadoes also form in equatorial type air masses that are moist to great heights. Such storms are most common on the coasts of the Gulf of Mexico and produce the waterspouts often reported over Florida. Tornadoes are triggered in this air mass primarily by lifting at the intersection of a thunderstorm line with a warm front, and less frequently by frontal and prefrontal squall lines.

West Coast Type

Tornadoes also form in relatively cold moist air. This air mass tornado is the Pacific or West Coast type. It is responsible for waterspouts on the West Coast. Tornadoes in this type of air mass are normally in a rather extensive cloudy area with scattered rain showers and isolated thunderstorms. Clouds are mostly stratocumulus. Favorable situations for tornado development in this air mass type include the rear of Maritime Polar (mP) cold fronts—well cooled air behind squall lines.

WATERSPOUTS

Waterspouts fall into two classes-tornadoes over water and fair weather waterspouts. The fair weather waterspout is comparable to a dust devil. It may rotate in either direction, whereas the other type of waterspout rotates cyclonically. In general, waterspouts are not as strong as tornadoes, in spite of the large moisture source and the reduced frictiion of the underlying surface. The water surface beneath a waterspout is either raised or lowered, depending on whether it is affected more by the atmospheric pressure reduction or the wind force. There is less inflow and upflow of air in a waterspout than in a tornado. The waterspout does not lift a significant amount of water from the surface. Ships passing through waterspouts have mostly encountered fresh water.

FORECASTING FOG AND STRATUS

LEARNING OBJECTIVES: Be familiar with the effects of air-mass stability on fog formation. Identify the procedures used in the forecasting of fog. Recognize conditions favorable for the formation of the various types of fog. Calculate fog parameters by using the Skew T Log P Diagram.

Fog and stratus clouds are hazardous conditions for both aircraft and ship operations. You will frequently be called upon to forecast formation, lifting, or dissipation of these phenomena. To provide the best information available, we will discuss the various factors that influence the formation and dissipation of fog and stratus.

EFFECT OF AIR MASS STABILITY ON FOG

Fog and stratus are typical phenomena of a warm air mass. Since a warm air mass is warmer than the underlying surface, it is stable, especially in the lower layers.

Through the use of upper air soundings, measurements can be made of temperature and relative humidity, from which stability characteristics can be determined. Refer to the publication, *Use of the Skew T, Log P in Analysis and Forecasting,* NAVAIR 50-IP-5, for complete information on analyzing upper air soundings.

GENERAL PROCEDURE FOR FORECASTING FOG

The synoptic situation, time of year, climatology of the station, air-mass stability, amount of cooling expected, strength of the wind, dewpoint-temperature spread, and trajectory of the air over favorable types of underlying surfaces are basic considerations you should take into account when forecasting fog.

Consideration of Geography and Climatology

Certain areas are more favorable climatologically for fog formation during certain periods of the year than others. All available information pertaining to climatology should be compiled for your station or operating area to determine the times and periods most favorable for fog formation.

You should also determine the location of the station with respect to air drainage or upslope conditions. Next, determine the type of fog to which your location would be exposed. For example, inland stations would be more likely to have radiation fog and shore coastal stations advection fog. A determination should then be made of air trajectories favorable for fog formation at your station.

Frontal Fog

Frontal fogs are associated with the forecasting of the movement of fronts and their attendant precipitation areas. For example, fogs can form in advance of a warm front, in the warm air section behind the warm front (when the warm air dewpoint is higher than the cold air temperature), or behind a slow moving cold front when the air becomes saturated.

Air-mass Fog

The first step in the forecasting of air-mass fogs is to determine the trajectory of the air mass and estimate the changes that are expected to occur during the night. If the air mass has been heated during the day and there was no fog the preceding morning, no marked cloud cover during the day, and no overwater trajectory, fog will not form during the night. However, during fall and winter, nights are long and days are short, and conditions are generally stable. When a fog situation has been in existence, the same conditions tend to remain night after night, and the heating during the day is insufficient to effectively raise the temperature above saturation. Also, determine if the air has had a path over extensive bodies of water, and whether this path was sufficient to raise the humidity or lower the temperature sufficiently to form fog. Then construct nomograms, tables, etc., by using dewpoint depression against time of fog formation for various seasons and winds; modify these in the light of each particular synoptic situation.

FACTORS TO BE CONSIDERED IN FOG AND STRATUS FORMATION

Wind, saturation of the air mass, nocturnal cooling, and air-mass trajectories have a role in the formation of fog or stratus clouds.

Wind

Wind velocity is an important consideration in the formation of fog and/or low ceiling clouds.

When the temperature and dewpoint are near one another at the surface and eddy currents are 100 feet or more in vertical thickness, adiabatic cooling in the upward portion of the eddy could give the additional cooling needed to bring about saturation. Any additional cooling would place the air in a temporary supersaturated state. The extra moisture will then condense out of the air, producing a low ceiling cloud. Adiabatic heating in the downward portion of the eddy will usually evaporate the cloud particles. If all cloud particles evaporate before reaching the ground, the horizontal visibility should be good. However, if many particles reach the ground before evaporation, the horizontal visibility will be restricted by moderate fog. Clouds that form in eddy areas may at first be patchy and then become identified as *ragged stratus.* If the cloud forms into a solid layer, it will be a layer of stratus. When conditionally unstable air is present in the eddy, or if the frictional eddy currents are severe enough, stratocumulus clouds will form in the area. See figure. 5-13.

Saturation of the Air Mass

The saturation curve in figure 5-14 shows the amount of moisture in grams per kilogram the air will hold at various temperatures.

The air along the curve is saturated and is at its dewpoint. Any further cooling will yield water as a result of condensation; hence, fog or low ceiling clouds (depending upon the wind velocity) will form.

Nocturnal Cooling

Nocturnal cooling begins after the temperature reaches its maximum during the day. Cooling will continue until sunrise, or shortly thereafter. This cooling affects only the lower limits of the atmosphere. If nocturnal cooling reduces the temperature to a value near the dewpoint, fog or low clouds will develop. The wind velocity and terrain roughness will control the depth of the cooled air. Calm winds will allow a patchy type of ground fog or a shallow, continuous ground fog to form. Winds of 5 to 10 knots will usually allow the fog to thicken vertically. Winds greater than 10 knots will usually cause low stratus or stratocumulus to form. See figures 5-15 and 5-16 for examples of fog and stratus formation.

The amount of cooling at night is dependent on soil composition, vegetation, cloud cover, ceiling, and other factors. Cloud cover based below 10,000 feet has a greenhouse effect on surface temperatures, absorbing some terrestrial radiation and reradiating a portion of this heat energy back to be absorbed by the land. This causes a reduction in nocturnal cooling. Nocturnal cooling between 1530 local and sunrise will vary from as little as 5° to 10° (with an overcast sky condition based around 1,000 feet), to 25° or 30°F with a clear sky or a cloud layer above 10,000 feet. Other factors and exceptions must also be considered. If a front is expected to pass the station during the night, or onshore winds are expected to occur during the night, the amount of cooling expected would have to be modified in light of these developments.

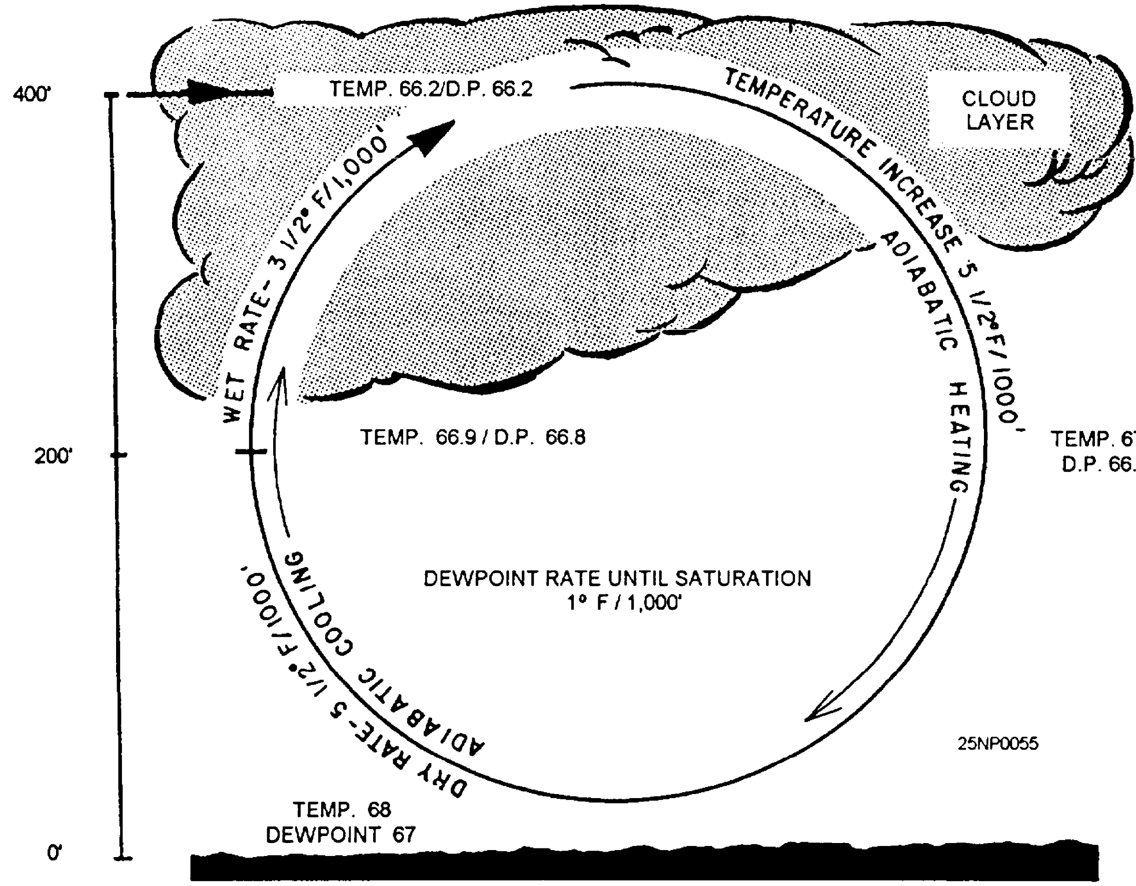

Figure 5-13.-How wind velocity can cause a low cloud layer.

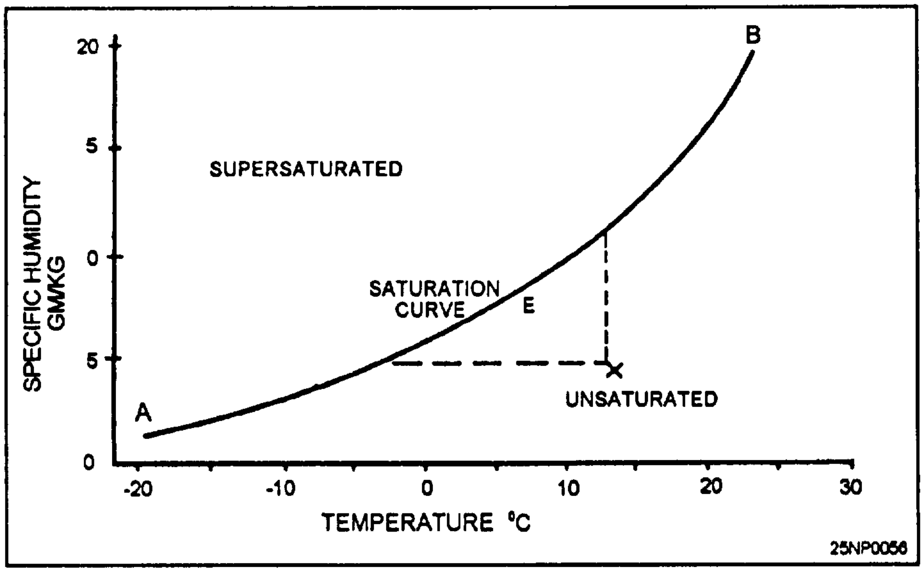

Figure 5-14.-Saturation curve.

An estimate of the formation time of fog, and possibly stratus, can be aided greatly if some type of saturation time chart, such as that illustrated in figure 5-17, can be constructed on which the forecasted temperature versus the forecasted dewpoint can be plotted. To use this diagram, note the maximum temperature and consider the general sky condition from the surface chart, forecasts, or sequence reports, By projecting the temperature and dewpoint temperature, an estimated time of fog formation can be forecast. If smoke is observed in the area, fog will normally form about 1 hour earlier than the formation line indicates on the charts because of the abundance of condensation nuclei.

Air-mass Trajectories

The trajectory of an air mass during the forecast period can be another important factor in fog formation. Warm air moving over a colder surface is a primary fog producer. This can happen when a station is in the warm sector, following a warm frontal passage. Cooling of the air mass takes place, allowing condensation and widespread fog or low stratus to form. To determine the probabilities of condensation behind a warm front, compare the temperature ahead of the front with the dewpoint behind the front. If the temperature ahead of the warm front is lower than the dewpoint behind the front, the air mass behind the front will cool to a temperature near the temperature ahead of the front, causing condensation and the formation of fog or low stratus.

Over water areas, warm air passing over cold water may cause enough cooling to allow condensation and the production of low clouds.

Another instance in which trajectory is important is when cold air moves over a warmer water surface, marsh land, or swamp, producing *steam fog.* In addition, air passing over a wet surface will evaporate a portion of the surface moisture, causing an increase in the dewpoint. Whenever there is a moisture source present, air will evaporate a portion of this moisture, unless the vapor pressure of the air is as great, or greater, than the vapor pressure of the water. A dewpoint increase may be enough to allow large eddy currents, nocturnal cooling, or terrain lifting to complete the saturation process and allow condensation to occur.

CONDITIONS FAVORABLE FOR GROUND OR RADIATION FOG

For the formation of ground or radiation fog, ideally, the air mass should be stable, moist in the lower layers, dry aloft, and under a cloud cover during the day, with clear skies at night. Winds should be light, nights long, and the underlying surface wet.

A stationary, subsiding, high-pressure area furnishes the best requirements for light winds, clear skies, stability, and dry air aloft. If the air in the high has been moving over a body of water, or if it lies over ground previously moistened by an active precipitating front, the wet surface will cause an increase in the dewpoint of the lowest layers of the air. In addition, long

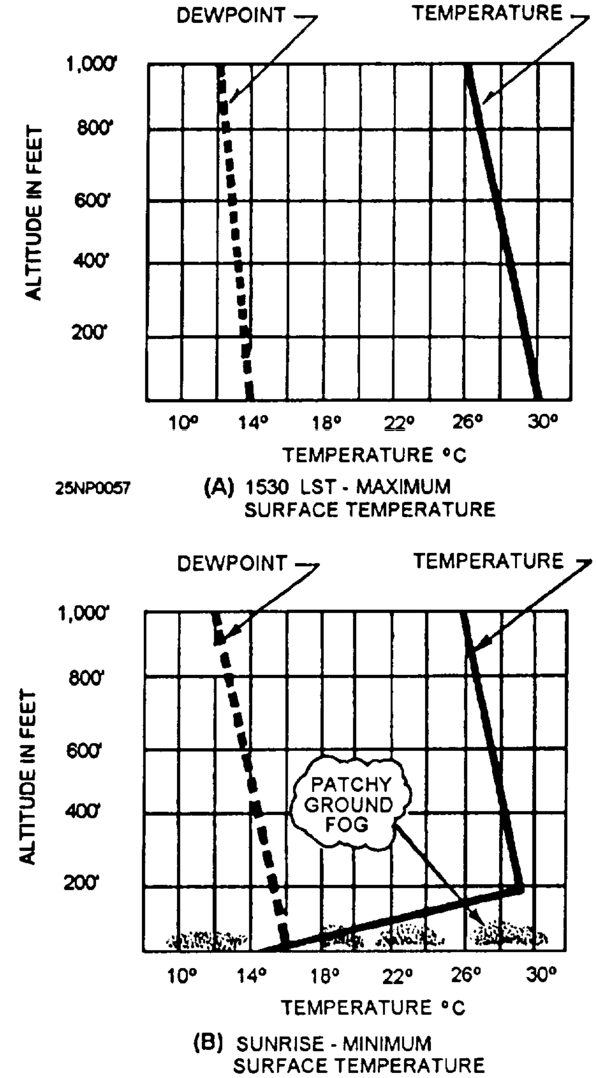

Figure 5-15.-Nocturnal cooling over a land area producing patchy ground fog (calm winds). (A) 1530 Local maximum surface temperature; (B) sunrise-minimum surface temperature.

nights versus short days in fall and winter are favorable for the formation of radiation fog.

CONDITIONS FAVORABLE FOR ADVECTION-RADIATION FOG

Cold and moist are apt descriptions of air masses that form in the late summer and early fall in the western quadrants of the Bermuda High. Cyclogenesis off the east coast of the United States, as well as the southerly flow associated with continental polar highs that have moved out over the ocean, are also cold and moist. If

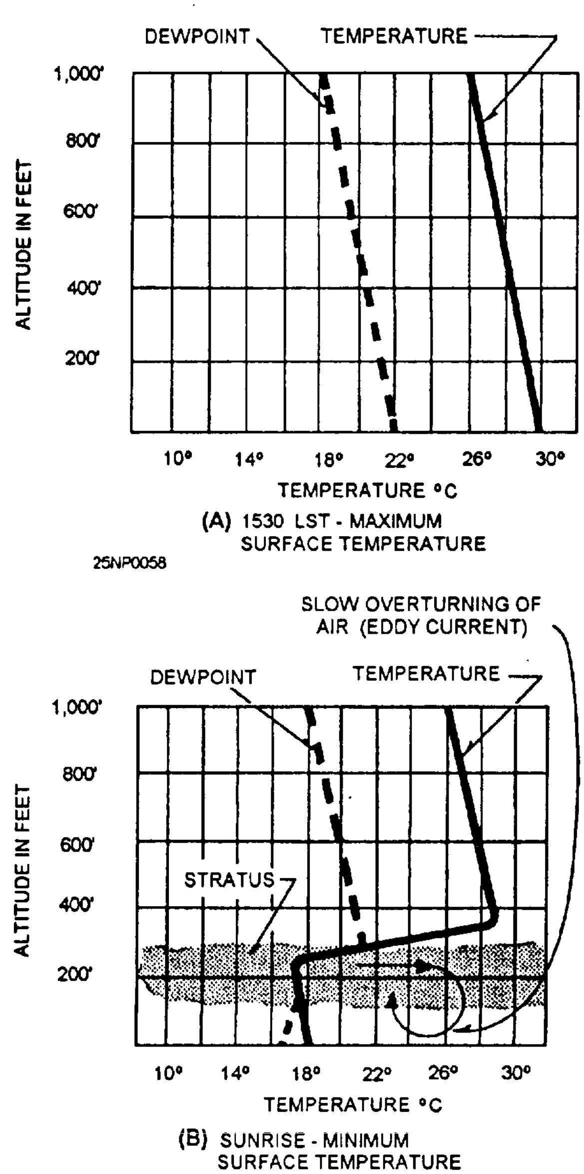

Figure 5-16.-Nocturnal cooling over a land area producing stratus (10- to 15-knot wind). (A) 1530 local maximum surface temperature; (B) sunrise-minimum surface temperature.

this air mass moves inland (replacing warm, dry, land air), it may be cooled to saturation due to radiational cooling during the long autumn nights with consequent formation of fog or stratus. The fog is limited to the coastal areas, extending inland between 150 and 250 miles, depending on the wind speed. On the east coast, it is limited to the region between the Appalachian Mountains and the Atlantic Ocean.

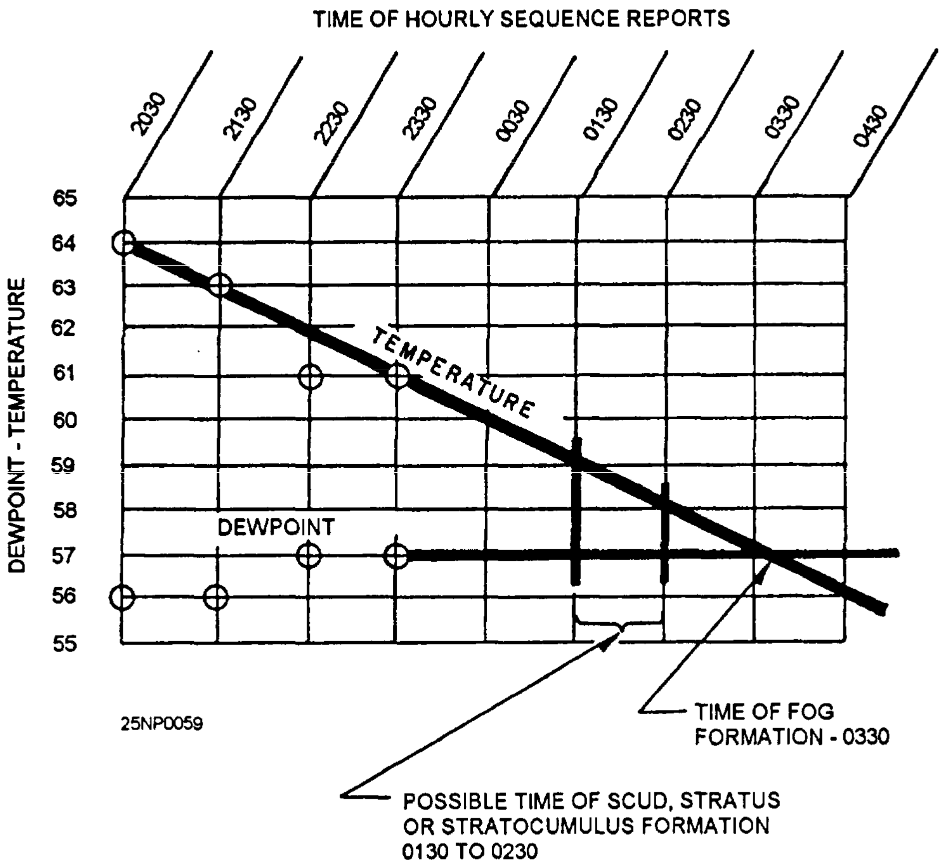

Figure 5-17.-Saturation time chart. (0's indicate actual temperature and dewpoint observations. Straight line is forecast temperature-dewpoint trends.)

In late fall and winter, when continental temperature gradients have intensified, and the land temperature has become colder than the adjacent water, poleward moving air is cooled by advection over colder ground, as well as by radiation. If the air is sufficiently moist, fog or stratus may form. During daytime, heating may dissipate the fog or stratus entirely. If not, the heating, together with the wind, which is advecting the air, sets up a turbulence inversion and stratus or stratocumulus layers form at the base of the inversion. At night if the air is cooled again and the surface pressure gradient is weak, a surface inversion may replace the turbulence inversion, and fog again occurs at the surface. However, if the pressure gradient is strong, cooling will intensify the inversion. Under these conditions stratus or stratocumulus clouds occur just as in the daytime, except with lower cloud bases.

Late fall and winter advection-radiation fogs cart occur any place over the continent that can be reached by maritime air or modified returning continental air. Mainly, this occurs over the eastern half of the United States. However, since tropical air masses do not reach as high a latitude in winter as in summer, the frequency of such fogs are much less in the northern regions of the country. With large, slow-moving, continental warm highs over the eastern half of the country, however, the fogs may extend all the way from the Gulf of Mexico to Canada.

CONDITIONS FAVORABLE FOR UPSLOPE FOG AND STRATUS

Upslope fog and stratus occur in those regions in which the land slopes gradually upward, and those areas accessible to humid, stable air masses. In North America, the areas best meeting these conditions are the Great Plains of the United States and Canada and the Piedmont region east of the Appalachians.

The synoptic conditions necessary for formation of this type of fog or stratus are the presence of humid air and a wind with an upslope component. The stratus is not advected over the station as a solid sheet. It forms gradually overhead. The length of time between the first signs of stratus and a ceiling usually ranges from 1/2 hour to 2 hours; although at times, the stratus may not form a ceiling at all. A useful procedure is to check the hourly observations of surrounding stations, especially those southeastward. If one of these stations starts reporting stratus, the chances of stratus formation at your station are high.

CONDITIONS FAVORABLE FOR FRONTAL FOG

Frontal fogs are of three types: prefrontal (warm front), postiontal (cold front), and frontal passage.

Prefrontal Fog

Prefrontal (warm front) fogs occur in stable continental polar (cP) air masses when precipitating warm air overrides the colder air. The rain raises the dewpoint in the cP air mass sufficiently for fog formation. Generally, the wind speeds are light, and the area most conducive to the formation of this type of fog is one between a nearby secondary low and the primary low-pressure center. The northeastern area of the United States is probably the most prevalent region for this type of fog. Prefrontal fog is also of importance along the Gulf and Atlantic coastal plains, the Midwest, and in the valleys of the Appalachians.

A rule of thumb for forecasting ceiling during prefrontal fog is as follows: If the gradient winds are greater than 25 knots, the ceiling will usually remain 300 feet or higher during the night.

Postfrontal Fog

As with the prefrontal fog, postfrontal (cold front) fogs are caused by falling precipitation. Fogs of this type are common when cold fronts with east-west orientations have become quasi-stationary and the continental polar air behind the front is stable. This type of fog is common in the Midwest. Fog, or stratiform clouds, may be prevalent for considerable distances behind cold fronts if the cold fronts produce precipitation.

Frontal Passage Fog

During the passage of a front, fog may form temporarily if the winds accompanying the front are very light and the two air masses are near saturation. Also, temporary fog may form if the air is suddenly cooled over moist ground with the passage of a precipitating cold front. In low latitudes, fog may form in the summer if the surface is cooled sufficiently by evaporation of rain that fell during a frontal passage, provided that the moisture addition to the air and the cooling are great enough to allow for fog formation.

CONDITIONS FAVORABLE FOR SEA FOG

Sea fogs are advection fogs that form in warm moist air cooled to saturation as it moves over colder water. The colder water may occur as a well-defined current, or as gradual latitudinal cooling. The dewpoint and the temperature undergo a gradual change as the air mass moves over colder and colder water. The surface air temperature falls steadily, and tends to approach the water temperature. The dewpoint also tends to approach the water temperature, but at a slower rate. If the dewpoint of the air mass is initially higher than the coldest water to be crossed, and if the cooling process continues sufficiently long, the temperature of the air ultimately falls to the dewpoint, and fog results. However, if the initial dewpoint is less than the coldest water temperature, the formation of fog is unlikely. Generally, in northward moving air masses or in air masses that have previously traversed a warm ocean current, the dewpoint of the air is initially higher than the cold water temperature to the north, and fog will form, provided sufficient fetch occurs.

The rate of temperature decrease is largely dependent on the speed at which the air mass moves across the sea surface, which, in turn, is dependent both on the spacing of the isotherms and the velocity of the air normal to them.

The dissipation of sea fog requires a change in air mass (a cold front). A movement of sea fog to a warmer land area leads to rapid dissipation. Upon heating, the fog first lifts, forming a stratus deck; then, with further heating, this cloud deck breaks up into a stratocumulus layer, and eventually into convective type clouds or evaporates entirely. An increase in wind velocity can lift sea fog, forming a stratus deck, especially if the air/sea temperature differential is small. Over very cold water, dense sea fog may persist even with high winds.

CONDITIONS FAVORABLE FOR ICE FOG

When the air temperature is below about -25°F, water vapor in the air that condenses into droplets is quickly converted into ice crystals. A suspension of ice crystals based at the surface is called "ice fog." Ice fog occurs mostly in the Arctic regions, and is mainly an artificial fog produced by human activities, It occurs locally over settlements and airfields where hydrocarbon fuels are burned—the burning of hydrocarbon fuels produces water vapor.

When the air temperature is approximately -30°F or lower, ice fog frequently forms very rapidly in the exhaust gases of aircraft, automobiles, or other types of combustion engines.

When there is little or no wind, it is possible for an aircraft to generate enough ice fog during landing or takeoff to cover the runway and a portion of the airfield. Depending on the atmospheric conditions, ice fogs may persist for periods of a few minutes to several days.

There is also a tine arctic mist of ice crystals that persists as a haze over wide expanses of the arctic basin during winter; this fine mist may extend upward through much of the troposphere, similar to a cirrus cloud with the base reaching the ground.

USE OF THE SKEW T LOG P DIAGRAM IN FORECASTING THE FORMATION AND DISSIPATION OF FOG

One of the most accepted methods for forecasting the formation and dissipation of fog makes use of an upper air sounding plotted on the Skew T Log P Diagram. The plotting of an upper air sounding is useful in forecasting both the formation and dissipation of fog, but it can be used more objectively in forecasting fog dissipation.

The use of an upper air sounding to determine the possibility of fog formation must be subjective. A study of the existing lapse rate should be made to determine the stability or instability of the lower layers. The surface layer must be stable before fog can form. If it is not found to be stable, the cooling expected during the forecast period must be considered, and this modification should be applied to the sounding to determine if the layer will be stable with the additional cooling.

The difference between the temperature and the dewpoint must be considered. If the air temperature and the dewpoint are expected to coincide during the period covered by the forecast, a formation of fog is very likely.

The expected wind speed must be considered. If the wind speed is expected to be strong, the cooling will not result in a surface inversion favorable for the formation of fog, but may result in an inversion above the surface, which is favorable for the formation of stratus clouds.

DETERMINATION OF FOG HEIGHT

An upper air sounding taken during the time fog is present will show a surface inversion. The fog will not necessarily extend to the top of the inversion. If the temperature and dewpoint have the same value at the top of the inversion, you can assume that the fog extends to the top of the inversion. However, if they do not have the same value, you can determine the depth of the fog by averaging the mixing ratio at the surface and the mixing ratio at the top of the inversion. The intersection of this average mixing ratio with the temperature curve is the top of the fog layer.

Two methods that may be used to find the height of the top of the fog layer, in feet, are reading the height directly from the pressure-height curve on the Skew T Log P Diagram or by using the dry adiabatic method.

1. In using the pressure-height curve method, locate the point where the temperature curve and the average mixing ratio line intersect on the Skew T Log P Diagram. Move this point horizontally until the pressure-height curve is intersected. Determine the height of the fog layer from the value of the pressure-height curve at this intersection.

2. The dry adiabatic method is based on the fact that the dry adiabatic lapse rate is 1°C per 100 m, or 1°C per 328 ft. Using this method, follow the dry adiabat from the intersection of the average mixing ratio line with the temperature curve to the surface level. Find the temperature difference between the point where the dry adiabat reaches the surface and the point of intersection of the dry adiabat and the average mixing ratio. For example, in figure 5-18, the dry adiabat at the surface is 25°C. The temperature at the intersection of the dry adiabat and the average mixing ratio is 20°C. By applying the dry adiabatic method with a lapse rate of 1°C per 328 ft, we find the height of the top of the fog layer as follows:

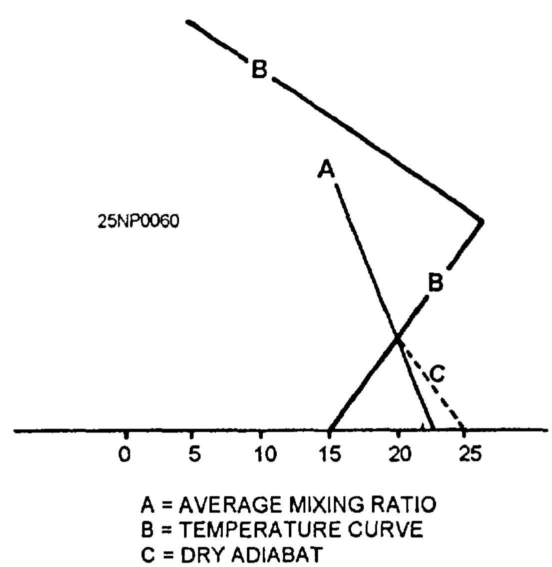

Figure 5-18.-Dry adiabatic method of determining fog height.

$$\frac{1}{328} = \frac{5}{x}$$

$$x = 1{,}640 \text{ ft}$$

DISSIPATION

To determine the surface temperature necessary for the dissipation of fog by using the Skew T Log P Diagram, trace dry adiabatically from the intersection of the average mixing ratio line and the temperature curve to the surface level. The temperature of the dry adiabat at the surface level is the temperature necessary for dissipation. This temperature is known as the CRITICAL TEMPERATURE. This temperature is an approximation, since it assumes no changes will take place in the stratum from the time of observation to the time of dissipation. This temperature should be modified on the basis of local conditions. See figure 5-18.

In considering the dissipation of fog and low clouds, you should consider the rate at which the surface temperature will increase after sunrise. Vertically thick fog, or multiple cloud layers, will slow up the morning heating at the surface. If advection fog is present, the fog may be lifted off the ground to a height where it is classified as stratus. If ground fog is present, the increase in surface air temperature will cause the fog particles to evaporate, thus dissipating the fog. Further heating may evaporate advection fog and low clouds.

FORECASTING ADVECTION FOG OVER THE OCEANS

In the absence of actual temperature and dewpoint data and with a stationary high (a southerly flow is assumed), use the following method to forecast advection fog over the ocean.

1. Pick out the point on an isobar at which the highest sea temperature is present (either from the surface chart or a mean monthly sea temperature chart). Assume that at this point, the air temperature is equal to that of the water and has a dewpoint 2 degrees lower.

2. Find the point on the isobar northward where the water is 2 degrees colder. From this point on, patchy light fog should occur.

3. From a saturation curve chart (fig. 5-14), find how much further cooling would have to occur to give an excess over saturation of 0.4 GM/KG, and also 2.0 GM/KG. The first represents the beginning of moderate fog and the second represents drizzle.

4. As the air continues around the northern ridge of the high, it will reach its lowest temperature, and from then on will be subject to warming. The pattern will then be drizzle until the excess is reduced to 2.0 GM/KG, and moderate fog until 0.4 GM/KG is reached.

If actual water and temperature data are available, use these in preference to climatic mean data. If the high is moving, trajectories will have to be calculated.

The fog is usually less widespread than calculated, and drizzle is less extensive. Also, clearing and lifting on the east side of the high is slightly faster. This method appears to work well in the summer over the Aleutian areas where such fog is frequent.

FORECASTING UPSLOPE FOG

Orographic lifting of the air will cause adiabatic cooling at the dry adiabatic rate of 5.5°F per 1,000 feet. If an adequate amount of lifting occurs, fog or low clouds will form. This process can create challenges for the forecaster.

The procedures for determining the probability of fog or low clouds during nighttime hours at stations having upslope winds are as follows:

1. Forecast the amount of nocturnal cooling,

2. Determine the expected amount of upslope cooling by using the following steps:

a. Determine the approximate number of hours between sunset and sunrise.

b. Estimate the expected wind velocity during the nighttime hours.

c. Multiply a by b. This will give the distance the upslope wind will move during the period of the day when daylight heating cannot counteract upslope cooling.

d. Determine the approximate terrain elevation difference between the station and the distance computed in c. Elevation difference should be in feet. (Example, 2.5 thousand feet.)

e. Multiply the elevation difference by the dry adiabatic rate of cooling. (Example, 2.5 times 5.5 = 13.75°F of upslope cooling.)

3. Add the expected amount of upslope cooling to the expected nocturnal cooling to arrive at the total amount of cooling.

4. Determine the late afternoon temperature dewpoint spread at the station under consideration. If

the expected cooling is greater than the late afternoon spread, either fog or low clouds should be expected. Wind velocity will determine which of the two conditions will form.

FORECASTING STRATUS FORMATION AND DISSIPATION

Fog and stratus forecasting are so closely tied together that many of the fog forecasting rules and conditions previously mentioned also apply to the forecasting of stratus clouds.

Determining the Base and Top of a Stratus Layer

One of the first steps in forecasting the dissipation of stratus is to determine the thickness of the stratus layer. The procedure is an follows:

1. Determine a representative mixing ratio between the surface and the base of the inversion.

2. Project this mixing ratio line upward through the sounding.

3. The intersection of the average mixing ratio line with the temperature curve gives the approximate base and maximum top of the stratus. Point A in figure 5-19 is the base of the stratus layer, and point B is the maximum top of the layer. Point A is the initial base of the layer; but as heating occurs during the morning, the base will lift. Point B represents the maximum top of the stratus layer; although in the very early morning, it might lie closer to the base of the inversion. However, as heating occurs during the day, the top of the stratus layer will also rise and will be approximated by point B. If the temperature and the dewpoint are the same at the top of the inversion, the stratus will extend to this level.

To determine the height of the base and the top of the stratus layer, use either the method previously outlined for fog, or the pressure altitude scale.

Determining Dissipation Temperatures

To determine the temperature necessary for the dissipation of a stratus layer, the following steps are provided:

1. From point A in figure 5-19, follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the temperature required to be reached for stratus dissipation to begin. This is point C.

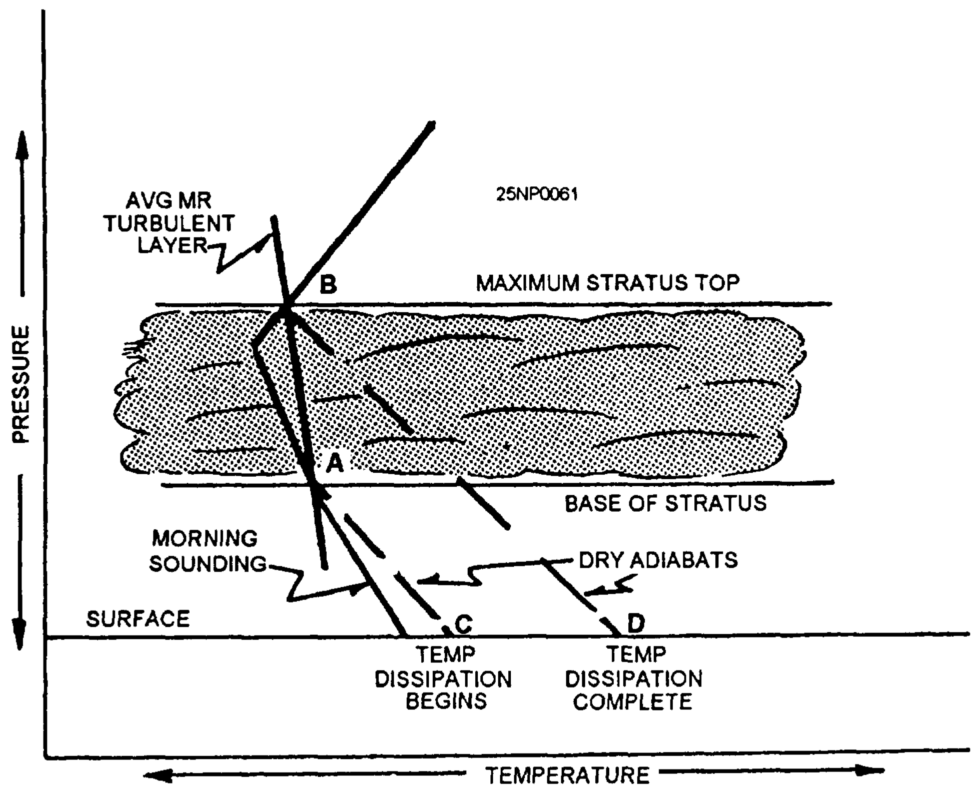

Figure 5-19.-Sounding showing the base and the top of stratus layers. Also note temperature at which dissipation begins and temperature when dissipation is complete.

2. From point B in figure 5-19, follow the dry adiabat to the surface level. The temperature of the dry adiabat at the surface level is the surface temperature required for the dissipation of the stratus layer to be complete. This is point D.

Determining Time of Dissipation

After determination of the temperatures necessary for stratus dissipation to begin and to be completed, a forecast of the time these temperatures will be reached must be made. Estimate the length of time for the required amount of heating to take place; and on the basis of this estimate, the time of dissipation may be forecasted. Remember to take into consideration the absence or the presence of cloud layers above the stratus deck. In addition, consider the trajectory of the air over the station. If the trajectory is from a water surface, temperatures will beheld down for a longer than normal period of time.

One rule of thumb used widely in forecasting the dissipation of the stratus layer is to estimate the thickness of the layer; and if no significant cloud layers are present above and normal heating is expected, forecast the dissipation of the layer with an average of 360 feet per hour of heating. In this way an estimate can be made of the number of hours required to dissipate the layer.

AIRCRAFT ICING

LEARNING OBJECTIVES: Recall factors conducive to aircraft icing. Be familiar with icing hazards at the surface. Analyze aircraft icing forecasts by using synoptic data. Prepare aircraft icing forecasts by using the –8D method.

Aircraft icing is another of the weather hazards to aviation. It is important that the pilot be advised of icing because of the serious effects it may have on aircraft performance. Ice on the airframe decreases lift and increases weight, drag, and stalling speed. In addition, the accumulation of ice on exterior movable surfaces affects the control of the aircraft. If ice begins to form on the blades of the propeller, the propeller's efficiency is decreased, and still further power is demanded of the engine to maintain flight. Today, most aircraft have sufficient reserve power to fly with a heavy load of ice; airframe icing is still a serious problem because it results in greatly increased fuel consumption and decreased range. Further, the possibility always exists that engine-system icing may result in loss of power.

The total effects of aircraft icing are a loss of aerodynamic efficiency; loss of engine power; loss of proper operation of control surfaces, brakes, and landing gear; loss of aircrew's outside vision; false flight instrument indications; and loss of radio communication. For these reasons, it is important that you, the forecaster, be alert and aware of the conditions conducive to ice formation. It is also important that you accurately forecast icing conditions during flight weather briefings.

This chapter will cover icing intensities, icing hazards near the ground, operational aspects of aircraft icing, and icing forecasts. For a discussion of the types of icing, physical factors affecting aircraft icing, and the distribution of icing in the atmosphere, refer to the *AG2* TRAMAN, volume 2, unit 6, as well as *Atmospheric Turbulence and Icing Criteria,* NAVMETOCCOMINST 3140.4, which discusses associated phenomena, as well as a common set of criteria for the reporting of icing.

SUPERCOOLED WATER IN RELATION TO ICING

Two basic conditions must be met for ice to form on an airframe in significant amounts. First, the aircraft surface temperature must be colder than 0°C. Second, supercooled water droplets, or liquid water droplets at subfreezing temperatures, must be present. Water droplets in the free air, unlike bulk water, do not freeze at 0°C. Instead, their freezing temperature varies from an upper limit near –10°C to a lower limit near —40°C. The smaller and purer the droplets, the lower their freezing point. When a supercooled droplet strikes an object, such as the surface of an aircraft, the impact destroys the internal stability of the droplet and raises its freezing temperature. In general, the possibility of icing must be anticipated in any flight through supercooled clouds or liquid precipitation at temperatures below freezing. In addition, frost sometimes forms on an aircraft in clear, humid air if both the aircraft and air are at subfreezing temperatures.

PROCESS OF ICE FORMATION ON AIRCRAFT

The first step in ice formation is when the supercooled droplets strike the surface of the aircraft. As the droplet, or portion of it, freezes, it liberates the heat of fusion. Some of this liberated heat is taken on by the unfrozen portion of the drop; its temperature is thereby increased, while another portion of the heat is conducted away through the surface in which it lies. The unfrozen drop now begins to evaporate due to its increase in temperature, and in the process, it uses up some of the heat, which, in turn, cools the drop. Due to this cooling process by evaporation, the remainder of the drop is frozen. Icing at 0°C will occur only if the air is not saturated because the nonsaturated condition is favorable for evaporation of part of the drop. Evaporation cools the drop below freezing, and then ice formation can take place.

INTENSITIES OF ICING

There are three intensities of aircraft icing—light, moderate, and severe.

Light

The rate of accumulation may create a problem if flight is prolonged in this environment (over 1 hour). Occasional use of deicing/anti-icing equipment removes/prevents accumulation. It does not present a problem if the deicing/anti-icing equipment is used.

Moderate

The rate of accumulation is such that even short encounters become potentially hazardous, and the use of deicing/anti-icing equipment or diversion is necessary.

Severe

The rate of accumulation is such that deicing/anti-icing fails to reduce or control the hazard. Immediate diversion is necessary.

ICING HAZARDS NEAR THE GROUND

Certain icing hazards exist on or near the surface. One hazard results when wet snow is falling during takeoff. This situation can exist when the air temperature at the surface is at or below 0°C. Wet snow sticks tenaciously to aircraft components, and it freezes if the aircraft encounters markedly colder temperatures during takeoff.

If not removed before takeoff, frost, sleet, freezing rain, and snow accumulation on parked aircraft become operational hazards. Another hazard arises from the presence of puddles of water, slush, and/or mud on airfields. When the temperature of the airframe is colder than 0°C, water blown by the propellers or splashed by wheels can form ice on control surfaces and windows. Freezing mud is particularly dangerous because the dirt may clog controls and cloud the windshield.

OPERATIONAL ASPECTS OF AIRCRAFT ICING

Due to the large number of types and different configurations of aircraft, this discussion is limited to general aircraft types, rather than specific models.

Turbojet Aircraft

These high speed aircraft generally cruise at altitudes well above levels where severe icing exists. The greatest problem will be on takeoff, climb, and approach because of the greater probability of encountering supercooled water droplets at low altitudes. Also, the reduced speeds result in a decrease of aerodynamic heating.

Turbojet engines experience icing both externally and internally. All exposed surfaces are subject to external airframe icing.

Internal icing may pose special problems to turbojet aircraft engines. In flights through clouds that contain supercooled water droplets, air intake duct icing is similar towing icing. However, the ducts may ice when skies are clear and temperatures are above freezing. While taxiing and during takeoff and climb, reduced pressure exists in the intake system, which lowers temperatures to a point that condensation and/or sublimation takes place, resulting in ice formation. This temperature change varies considerably with different types of engines. Therefore, if the free air temperature is 10°C or less (especially near the freezing point) and the relative humidity is high, the possibility of induction icing definitely exists.

Turboprop Aircraft

The problems of aircraft icing for this type of aircraft combine those of conventional aircraft and turbojet aircraft. Engine icing problems are similar to those encountered by turbojet aircraft, while propeller icing is similar to that encountered by conventional aircraft.

Propeller icing is a very dangerous form of icing because of the potential for a tremendous loss of power and vibrations. Propeller icing varies along the blade due to the differential velocity of the blade, causing a temperature increase from the hub to the propeller tip. Today's turboprop aircraft have deicers on the propellers. However, these deicers are curative, not preventive, and the danger remains.

Rotary-Wing Aircraft

When helicopters encounter icing conditions, the icing threat is similar, but potentially more hazardous than with fixed-wing aircraft. When icing forms on the rotor blades while hovering, conditions become hazardous because the helicopter is operating near peak operational limits. Icing also affects the tail rotor, control rods and links, and air intakes and filters.

PRELIMINARY CONSIDERATIONS

The first phase in the preparation of an aircraft icing forecast consists of making certain preliminary determinations. These are essential regardless of the technique employed in making the forecast.

Clouds

Determine the present and forecast future distribution, type, and vertical extent of clouds along the flight path. The influence of local effects, such as terrain features, should not be overlooked.

Temperature

Determine those legs of the proposed flight path that will be in clouds colder than 0°C. A reasonable estimate of the freezing level can be made from the data contained on freezing level charts, constant-pressure charts, rawinsonde observations, and airways reports (AIREP) observations, or by extrapolation from surface temperatures.

Precipitation

Check surface reports for precipitation along the proposed flight path, and forecast the precipitation character and pattern during the flight. Special consideration should be given to the possibility of freezing precipitation.

Note that each of the following methods and forecast rules assumes that two basic conditions must exist for the formation of icing. These assumptions are

1. the surface of the aircraft must be colder than 0°C, and

2. supercooled liquid-water droplets, clouds, or precipitation must be present along the fight path.

THE ICING FORECAST

The following text discusses icing intensity forecasts by using upper air data, surface data, and precipitation data, as well as icing formed due to orographic and frontal lifting.

Intensity Forecasts From Upper Air Data

Check upper air charts, pilot reports, and rawinsonde reports for the dewpoint spread at the flight level. Also check the upper air charts for the type of temperature advection along the route. One study, which considered only the dewpoint spread aloft, found that there was an 84 percent probability that there would be no icing if the spread were greater than 3°C, and an 80 percent probability that there would be icing if the spread were less than 3°C.

When the dewpoint spread was 3°C or less in areas of warm air advection at flight level, there was a 67 percent probability of no icing and a 33 percent probability of light or moderate icing. However, with a dewpoint spread of 3°C or less in a cold frontal zone, the probability of icing reached 100 percent. There was also a 100-percent probability y of icing in building cumuliform clouds when the dewpoint spread was 3°C or less. With the spread greater than 3°C, light icing was probable in about 40 percent of the region of cold air advection with a 100-percent probability of no icing in regions of warm or neutral advection. However, it appears on the basis of further experience that a more realistic spread of 4°C at temperatures near –10° to -15°C should be indicative of probable clouds, and that spreads of about 2° or 3°C should be indicative of probable icing. At other temperatures use the values in the following rules:

- If the temperature is:

— 0° to –7°C and the dewpoint spread is greater than 2°C, there is an 80-percent probability of no icing.

— –8° to –15°C and the dewpoint spread is greater than 3°C, forecast no icing with an 80-percent chance of success.

— –16° to –22°C and the dewpoint spread is greater than 1°C, a forecast of no icing would have a 90-percent chance of success.

— Colder than –22°C, forecast no icing regardless of what the dewpoint spread is with 90-percent probability of success.

- If the dewpoint spread is 2°C or less at temperatures of 0° to –7°C, or is 3°C or less at –8° to –15°C:

— In zones of neutral, or weak cold air advection, forecast light icing with a 75-percent probability of success.

— In zones of strong cold air advection, forecast moderate icing with an 80-percent probability of success.

— In areas with vigorous cumulus buildups due to insolational surface heating, forecast moderate icing.

Intensity Forecasts From Surface Chart Data

If upper air data is not available, check the surface charts for locations of cloud shields associated with fronts, low-pressure centers, and precipitation areas along the route.

Intensity Forecasts From Precipitation Data

Within clouds not resulting from frontal activity or orographic lifting, and over areas with steady nonfreezing precipitation, forecast little or no icing. Over areas not experiencing precipitation, but having cumuliform clouds, forecast moderate icing.

Intensity Forecasts Based on Clouds Due To Frontal or Orographic Lifting

Within clouds resulting from frontal, or orographic lifting, neither the presence, nor the absence of precipitation can be used as indicators of icing. The following observations have proven to be accurate:

- Within clouds up to 300 miles ahead of the warm front surface position, forecast moderate icing.
- Within clouds up to 100 miles behind the cold front surface position, forecast severe icing.
- Within clouds over a deep, almost vertical low-pressure center, forecast severe icing.
- In freezing drizzle, below or in clouds, forecast severe icing.
- In freezing rain, below or in clouds, forecast severe icing.

ICING TYPE FORECASTS

The following rules apply to the forecasting of icing types:

- Forecast rime icing when the temperatures at flight level are colder than –15°C, or when between –1° and –15°C in stable stratiform clouds.
- Forecast clear icing when temperatures are between 0° and –8°C in cumuliform clouds and freezing precipitation.
- Forecast mixed rime and clear icing when temperatures are between –9° and –15°C in cumuliform clouds.

EXAMPLE OF ICING FORECASTS USING THE SKEW T LOG P DIAGRAM

It has been previously pointed out how the thickness of clouds, as well as the top of the overcast, may be estimated with accuracy from the dewpoint depression and changes in the lapse rate.

The analysis of cloud type and icing type from the Skew T Log P Diagram is illustrated in figure 5-20. First, look at the general shape of the curve. The most prominent feature is the inversion, which shows a very stable layer between 2,500 and 3,000 feet. Note that the dewpoint depression is less than 1 degree at the base of the inversion. Moisture exists in a visible format this dewpoint depression, so expect a layer of broken or overcast clouds whose base will be approximately 2,000 feet and will be topped by the base of the inversion. The next most prominent feature is the high humidity, as reflected by the dewpoint depressions, at 8,000 feet. Since the dewpoint depression is 2°C, a probability of clouds exists at this level. There is a rapid increase in the dewpoint depression at 17,000 feet, indicating that the top of the cloud layer is at this level. You would then assume the cloud layer existed between 8,000 and 17,000 feet.

The next step is to determine, if possible, the type of clouds in between these levels. If this sounding were plotted on the Skew T Log P Diagram, you would be able to see that the slope of the lapse rate between 8,000 and 12,000 feet is shown to be unstable by comparison to the nearest moist adiabat. The clouds will then display unstable cumuliform characteristics. Above 12,000 feet, the lapse rate is shown by the same type of comparison to be stable, and the clouds there should be altostratus.

Now determine the freezing level. Note that the lapse rate crosses the 0°C temperature line at approximately 9,500 feet. Since you have determined that clouds exist at this level, the temperatures are between 0° and –8°C, and the clouds are cumuliform, forecast clear ice in the unstable cloud up to 12,000 feet. Above 12,000 feet, the clouds are stratiform, so rime icing should be forecast. The intensity of the icing would have to be determined by the considerations given in the previous section of this chapter.

Icing Forecasts Using the –8D Method

When surface charts, upper air charts, synoptic and airway reports, and pilot reports are not clear as to the presence and possibility of icing, it may be determined

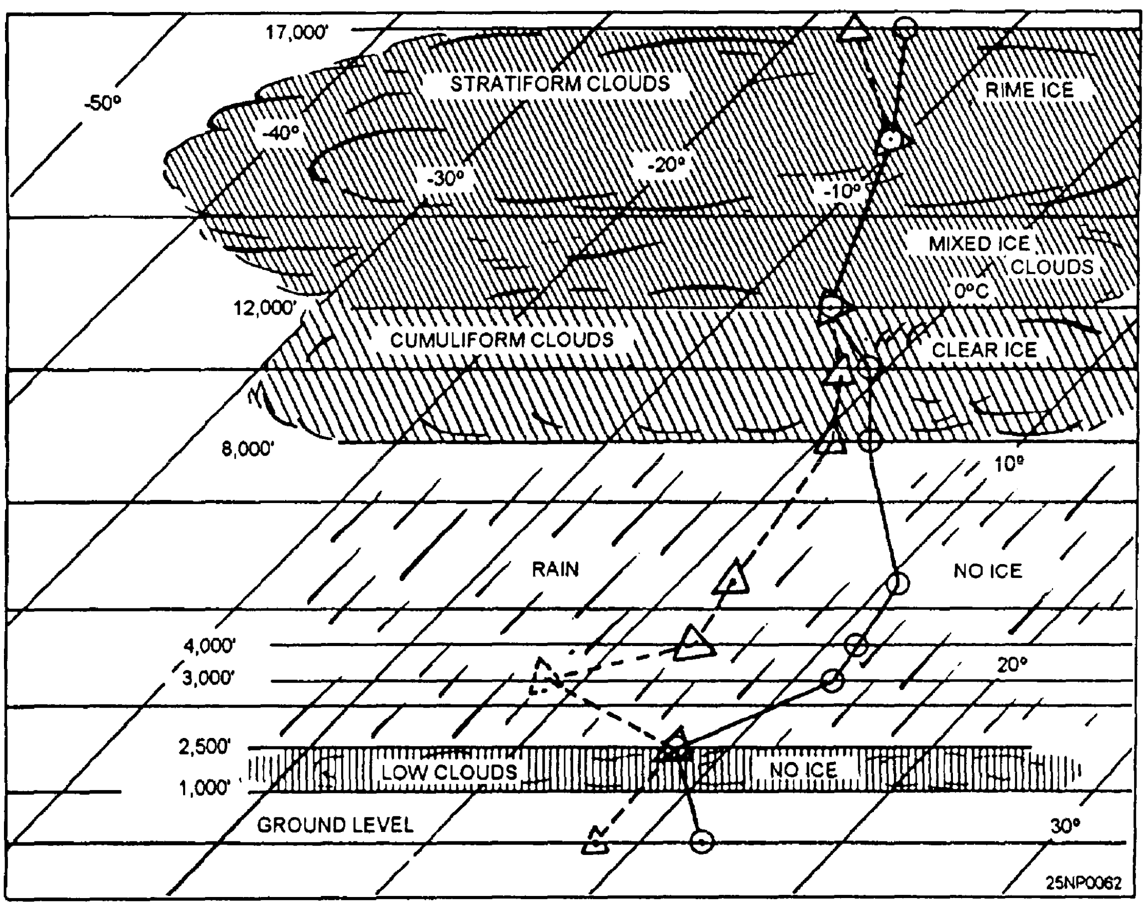

Figure 5-20.-An illustration of the analysis of cloud type and icing type from a Skew T Log P Diagram.

from the Skew T Log P Diagram by using the following 7 steps:

1. Plot the temperature against pressure as determined from a RAOB sounding.

2. Record the temperature and dewpoint in degrees and tenths to the left of each plotted point.

3. Determine the difference (in degrees and tenths) between the temperature and dewpoint for each level. This difference is D, the dewpoint deficit; it is always taken to be positive.

4. Multiply D by –8 and plot the product (which is in degrees Celsius) opposite the corresponding temperature point at the appropriate place.

5. Connect the points plotted by step 4 with a dashed line in the manner illustrated in figure 5-21.

6. The icing layer is outlined by the area enclosed by the temperature curve on the left and the –8D curve on the right. In this outlined area, supersaturation with respect to ice exists. This is the hatched area, as shown in figure 5-21.

7. The intensity of icing is indicated by the size of the area enclosed by the temperature curve and the –8D curve. In addition, the factors given in the following section should be considered when formulating the icing forecast. The cloud type and the precipitation observed at the RAOB time or the forecast time maybe used to determine whether icing is rime or glaze.

Conclusions arrived at by using the-SD method for forecasting icing:

- When the temperature and dewpoint coincide in the RAOB sounding, the –8D curve must fall along the 0°C isotherm. In a subfreezing layer, the air would be saturated with respect to water and supersaturated with respect to ice. Light rime icing would occur in the altostratus/nimbostratus clouds in such a region, and moderate rime icing would occur in cumulonimbus clouds in such a region. Severe clear ice would occur in the stratocumulus virga, cumulus virga, and stratus.

- When the temperature and dewpoint do not coincide but the temperature curve lies to the left of the –8D curve in the subfreezing layer, the layer is supersaturated with respect to cloud droplets. If the clouds in this layer are altostratus, altocumulus, cumulogenitus, or altocumulus virga, only light rime will be encountered. If the clouds are cirrus, cirrocumulus, or cirrostratus, only light hoarfrost will be sublimated on the aircraft. In cloudless regions, there

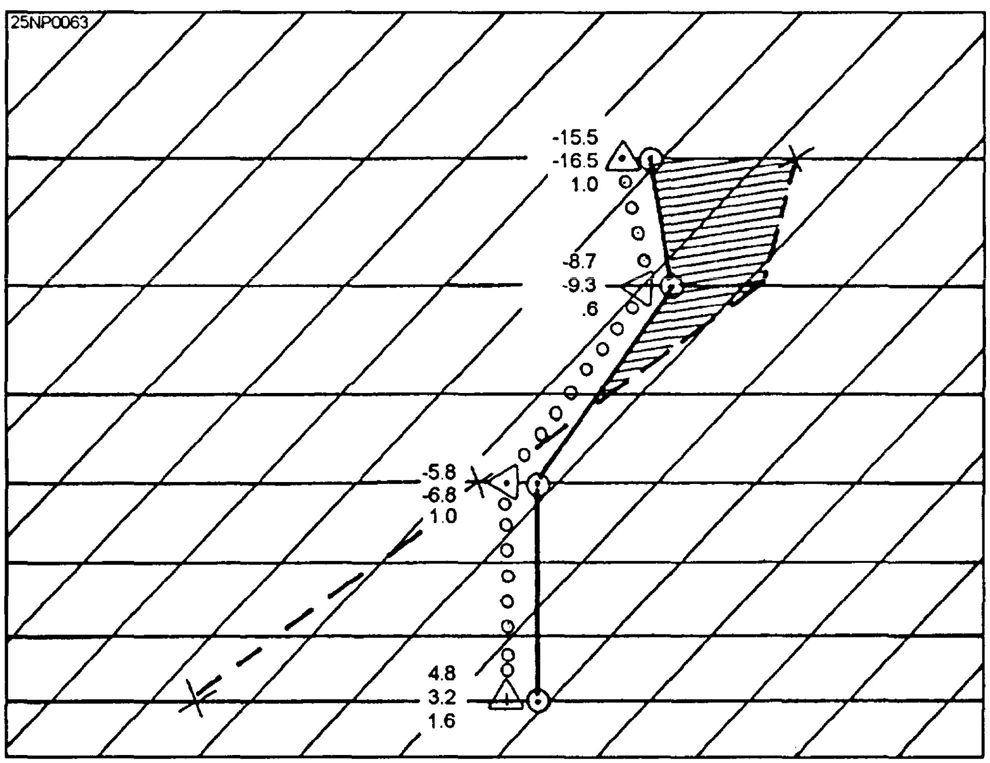

Figure 5-21.-The -8D ice forecast method.

will be no supercooled droplets, but hoarfrost will form on the aircraft through direct sublimation of water vapor.

- When the temperature curve lies to the right of the –8D curve in a subfreezing layer, the layer is subsaturated with respect to both ice and water surface. No icing will occur in this region.

Modification of the Icing Forecast

The final phase is to modify your icing forecast. This is essentially a subjective process. The forecaster should consider the following items: probable intensification or weakening of synoptic features, such as low-pressure centers, fronts, and squall lines during the time interval between the latest analysis and the forecast; local influences, such as geographic location, terrain features, and proximity to ocean coastlines or lake shores, and radar weather observations and pilot reports of icing. The forecaster should be cautious in either underforecasting or overforecasting the amount and intensity of icing. An overforecast results in a reduced payload for the aircraft due to increased fuel load, while an underforecast may result in an operational emergency.

TURBULENCE

LEARNING OBJECTIVES: Recall characteristics associated with turbulence. Determine the four intensities of turbulence. Forecast surface, in-cloud, and clear air turbulence (CAT). Recognize the advantages of the use of Doppler radar in turbulence forecasting.

Turbulence is of major importance to pilots of all types of aircraft; therefore, it is also of importance to the forecaster, whose duty it is to recognize situations where turbulence may exist, and to forecast both the areas and intensity of the turbulence. The following text discusses the classification and intensity of turbulence, the forecasting of turbulence near the ground, the forecasting of turbulence in convective clouds, and the forecasting of clear air turbulence (CAT).

Refer to the *AG2* TRAMAN, volume 2, unit 6, for a discussion of the types and properties of turbulence.

TURBULENCE CHARACTERISTICS

Turbulence may be defined as irregular and instantaneous motions of air that are made up of a number of small eddies that travel in the general air current. Atmospheric turbulence is caused by random fluctuations in the windflow. Given an analyzed wind field with both streamlines and isotachs smoothly drawn, any difference between an actual wind and this smooth field is attributed to turbulence.

To an aircraft in flight, the atmosphere is considered turbulent when irregular whirls or eddies of air affect the motion of the aircraft, and a series of abrupt jolts or bumps is felt by the pilot. Although a large range of sizes of eddies exists in the atmosphere, those causing bumpiness are roughly of the same size as the aircraft dimensions, and usually occur in an irregular sequence imparting sharp translation or angular motions to the aircraft. The intensity of the disturbances to the aircraft varies not only with the intensity of the irregular motions of the atmosphere but also with aircraft characteristics, such as flight speed, weight stability, and size.

Wind Shear and CAT

A relatively tight gradient, either horizontal or vertical, produces churning motions (eddies), which result in turbulence. The greater the change of wind speed and/or direction, the more severe the turbulence. Turbulent flight conditions are often found in the vicinity of the jetstream, where large shears in the horizontal and vertical are found. Since this type of turbulence may occur without any visual warning, it is often referred to as CAT.

The term *clear air turbulence* is misleading because not all high-level turbulence included in this classification occurs in clear air. However, the majority (75 percent) is found in a cloud-free atmosphere. CAT is not necessarily limited to the vicinity of the jetstream; it may occur in isolated regions of the atmosphere. Most frequently, CAT is associated with the jetstream or mountain waves. However, it may also be associated with a closed low aloft, a sharp trough aloft, or an advancing cirrus shield. A narrow zone of wind shear, with its accompanying turbulence, is sometimes encountered by aircraft as it climbs or descends through a temperature inversion. Moderate turbulence may also be encountered momentarily when passing through the wake of another aircraft.

The criteria for each type of CAT areas follows:

- Mountain wave CAT. Winds 25 knots or greater, normal to terrain barriers, and significant surface pressure differences across such barriers.

- Trough CAT. That portion of a trough that has horizontal shear on the order of 25 knots, or more, in 90 nautical miles.

- Closed low aloft CAT. If the flow is merging or splitting, moderate or severe CAT maybe encountered. Also, to the northeast of a cutoff low aloft, significant CAT may be experienced. As with the jetstream CAT, the intensity of this type of turbulence is related to the strength of the shear.

- Wind shear CAT. Those zones in space in which wind speeds are 60 knots or greater, and both horizontal and vertical shear exists, as indicated in table 5-1.

No provision is made for light CAT because light turbulence serves only as a flight nuisance. Any of the above situations can produce moderate to severe CAT. However, the combination of two or more of the above conditions is almost certain to produce severe or even extreme CAT. A jetstream may be combined with a mountain wave or be associated with a merging or splitting low.

Turbulence on the Lee Side of Mountains

When strong winds blow approximately perpendicular to a mountain range, the resulting turbulence may be quite severe. Associated areas of steady updrafts and downdrafts may extend to heights from 2 to 20 times the height of the mountain peaks. Under these conditions when the air is stable, large waves tend to form on the lee-side of the mountains, and may extend 150 to 300 miles downwind. They are referred to as *mountain waves.* Some pilots have reported that flow in these waves is often remarkably smooth, while others have reported severe turbulence. The structure and characteristics of the *mountain wave* were presented in volume 2 of the *AG2* TRAMAN. Refer to figure 6-1-5 in volume 2 for an illustration of a *mountain wave.*

The windflow normal to the mountain produces a primary wave, and, generally less intense, additional waves farther downwind. The characteristic cloud patterns may or may not be present to identify the wave. The pilot, for the most part, is concerned with the primary wave because of its more intense action and proximity to the high mountainous terrain. Severe turbulence frequently can be found 150 to 300 miles downwind, when the winds are greater than 50 knots at the mountaintop level. When winds are less than 50 knots at the mountaintop level, a lesser degree of turbulence may be experienced.

Some of the most dangerous features of the mountain wave are the turbulence in and below the roll cloud, the downdrafts just to the lee side of the mountain peaks, and to the lee side of the roll clouds. The *cap cloud* must always be avoided because of turbulence and concealed mountain peaks.

The following five rules have been suggested for flights over mountain ranges where waves exist:

1. The pilot should, if possible, fly around the area when wave conditions exist. If this is not feasible, he/she should fly at a level that is at least 50 percent higher than the height of the mountain range.

2. The pilot should avoid the roll clouds, since these are the areas with the most intense turbulence.

3. The pilot should avoid the strong downdrafts on the lee side of the mountain.

Table 5-1.-Wind Sheer CAT with Wind Speed 60 Knots or Greater

HORIZONTAL SHEAR (NAUT/MI)	VERTICAL SHEAR (PER 1,000 FT)	CAT INTENSITY
25k/90	9-12k	Moderate
25k/90	12-15k	Moderate, at times severe
25k/90	above 15k	Severe

4. He/she should also avoid high lenticular clouds, particularly if their edges are ragged

5. The pressure altimeter may read as much as 1,000 feet lower near the mountain peaks.

CLASSIFICATION AND INTENSITY OF TURBULENCE

The *Airman's Information Manual,* chapter 7, contains a turbulence reporting criteria table, which describes the meteorological characteristics with which the respective classes of turbulence are associated.

Terminal Aerodrome Forecast (TAF) Code, NAVMETOCCOMINST 3143.1, lists code figures for turbulence type and intensity.

The Turbulence Reporting Criteria Table, table 5-2, has been adopted as the standard. All NAVMETOCCOM units must adopt these as standards as a guide in forecasting turbulence.

The following is a guide to the classification of turbulence.

Extreme Turbulence

This rarely encountered condition is usually confined to the strongest forms of convection and wind shear, such as the following:

- Mountain waves in or near the rotor cloud, usually found at low levels, leeward of the mountain ridge when the wind normal to the mountain ridge exceeds 50 knots.

- In severe thunderstorms where the production of large hail (three-fourths inch or more) is indicated. It is more frequently encountered in organized squall lines than in isolated thunderstorms.

Severe Turbulence

Severe turbulence may also be found in the following:

- In mountain waves:

— When the wind normal to the mountain ridge exceeds 50 knots. The turbulence may extend to the tropopause, and at a distance of 150 miles leeward. A reasonable mountain wave turbulence layer is about 5,000 feet thick.

— When the wind normal to the mountain ridge is 25 to 50 knots, the turbulence may extend up to 50 miles leeward of the ridge, and from the mountain ridge up to several thousand feet above.

- In and near mature thunderstorms, and occasionally in towering cumuliform clouds.

- Near jetstreams within layers characterized by horizontal wind shears greater than 40 knots/150 run, and vertical wind shears in excess of 6 knots/1,000 feet. When such layers exist, favored locations are below and/or above the jet core, and from roughly the vertical axis of the jet core to about 50 to 100 miles toward the cold side.

Moderate Turbulence

Moderate turbulence may be found in the following:

- In mountain waves:

— When the wind normal to the mountain ridge exceeds 50 knots. Moderate turbulence may be found from the ridge line to as much as 300 miles leeward.

— When the wind normal to the ridge is 25 to 50 knots, moderate turbulence maybe found from the ridge line to as much as 150 miles leeward.

- In, near, and above thunderstorms, and in towering cumuliform clouds.

- Near jetstreams and in upper level troughs, cold lows, and fronts aloft where vertical wind shears exceed 6 knots/1,000 feet, or horizontal wind shears exceed 10 knots per 100 miles.

- At low altitudes, usually below 5,000 feet, when surface winds exceed 25 knots, or the atmosphere is unstable due to strong insolation or cold advection.

Light Turbulence

In addition to the situations where more intense classes of turbulence occur, the relatively common class of light turbulence maybe found:

- In mountainous areas, even with light winds.

- In and near cumulus clouds.

- Near the tropopause.

- At low altitudes when winds are under 15 knots, or the air is colder than the underlying surface.

Table 5-2.-Turbulence Reporting Criteria Table

Intensity	Aircraft Reaction	Reaction Inside Aircraft	Reporting Term-Definition
Light	Turbulence that momentarily causes slight, erratic changes in altitude and/or attitude (pitch, roll, yaw). Report as **Light Turbulence;** or Turbulence that causes slight, rapid, and somewhat rhythmic bumpiness without appreciable changes in altitude or attitude. Report as **Light Chop.**	Occupants may feel a slight strain against seat belts or shoulder straps. Unsecured objects may be displaced slightly. Food service may be conducted and little or no difficulty is encountered in walking.	Occasional-Less that 1/3 of the time. Intermittent-1/3 to 2/3. Continuous-More than 2/3.
Moderate	Turbulence that is similar to Light Chop but of greater intensity. It causes rapid bumps or jolts without appreciable changes in aircraft altitude or attitude. Report as **Moderate Chop.**[1]	Occupants feel definite strain against seat belts or shoulder straps. Unsecured objects are dislodged. Food service and walking are difficult.	**NOTE:** 1. Pilots should report location(s), time (UTC), intensity, whether in or near clouds, altitude, type of aircraft and, when applicable, duration or turbulence. 2. Duration may be based on time between two locations or over a single location. All locations should be readily identifiable. **EXAMPLES:** a. Over Omaha. 1232 UTC, moderate turbulence, in cloud, flight level 310, B707. b. From 50 miles south of Albuquerque to 30 miles north of Phoenix, 1210 UTC to 1250 UTC, occasional moderate chop, flight level 330, DC8.
Severe	Turbulence that causes large, abrupt changes in altitude and/or attitude. It usually causes large variations in indicated airspeed. Aircraft may be momentarily out of control. Report as **Severe Turbulence.**[1]	Occupants are forced violently against seat belts or shoulder straps. Unsecured objects are tossed about. Food service and walking are impossible.	
Extreme	Turbulence in which the aircraft is violently tossed about and is practically impossible to control. It may cause structural damage. Report as **Extreme Turbulence.**[1]		

[1]High level turbulence (normally above 15,000 feet ASL) not associated with cumuliform cloudiness, including thunderstorms, should be reported as CAT (clear air turbulence) preceded by the appropriate intensity, or light or moderate chop.

FORECASTING TURBULENCE NEAR THE SURFACE

Over land during nighttime hours there is very little turbulence near the surface. The only exception is that of high wind speeds over rough terrain. This type of turbulence decreases with increasing height.

During the daylight hours, turbulence near the surface depends on the radiation intensity, the lapse rate, and the wind speed. Turbulence intensity tends to increase with height throughout the unstable and neutral layers above the surface to the first inversion, or stable layer. Similar turbulence occurs in fresh polar outbreaks over warm waters.

Vertical gustiness increases with height more rapidly than horizontal gustiness.

Situations for violent turbulence near the surface occur shortly after a cold frontal passage, especially over rough terrain. Other examples of turbulence occur over deserts on hot days and during thunderstorms.

Peak gusts at the surface can be estimated to be essentially equal to the wind speeds at the gradient level, except in thunderstorms.

FORECASTING TURBULENCE IN CONVECTIVE CLOUDS

In the absence of any dynamic influences that might serve to drastically modify the vertical temperature and moisture distribution, on-hand rawinsonde data maybe used to evaluate the turbulence potential of convective clouds or thunderstorms for periods up to 24 hours. The following is one procedure for predicting turbulence in such clouds. This method is often referred to as the *Eastern Airlines Method:*

The technique is as follows:

- Determine the CCL.

- From the CCL, proceed along the moist adiabat to the 400-hPa level. This curve is referred to as the *updraft curve.*

- Compare the departure of the updraft curve with the free air temperature (T) curve at the 400-hPa level. For positive values, the updraft curve should be warmer than the (T) curve. That is, the updraft curve would be to the right of the (T) curve. The value of the maximum positive departure obtained in this step is referred to as ΔT.

Table 5-3 is based on several years relating AT values to commercial pilot reports of thunderstorm turbulence, and can be used to predict the degree of turbulence in air mass thunderstorms.

This method of forecasting thunderstorm turbulence is almost exclusively confined to the warmer months when frontal cyclonic activity is at a minimum. During the cooler months, frontal and cyclonic influences may cause rapid changes in the vertical distribution of temperatures and moisture, and some other methods have to be used.

FORECASTING CLEAR AIR TURBULENCE (CAT)

Clear air turbulence is one of the more common in-flight hazards encountered by high-altitude, high-performance aircraft.

Not all high-level turbulence occurs in clear air. However, a rough, bumpy ride may occur in clear air, without visual warning. This turbulence may be violent enough to disrupt tactical operations, and possibly cause serious airframe stress and/or damage.

Most cases of CAT at high altitudes can be attributed to the jetstream, or more specifically, the abrupt vertical wind shear associated with the jetstream. CAT is experienced most frequently during the winter months when the jetstream winds are the strongest.

The association of CAT with recognizable synoptic features has become better understood over the last few years. The following are general areas where CAT may occur:

- In general, in any region along the jetstream axis where wind shear appears to be strong horizontally, vertically, or both.

- In the vicinity of traveling jet maxima, particularly on the cyclonic side.

- In the jetstream below and to the south of the core.

- Near 35,000 feet in cold, deep troughs.

The instruction *Atmospheric Turbulence and Icing Criteria,* NAVMETOCCOMINST 3140.4, sets forth associated phenomena, as well as a common set of criteria for the reporting of turbulence.

Table 5.3.-Relation of Maximum Positive Departure to Thunderstorm Turbulence

ΔT (°C)	TURBULENCE
0-3	Light
4-6	Moderate
7-9	Severe
Above 9	Extreme

OBSERVATION OF SEVERE WEATHER FEATURES USING DOPPLER RADAR

Doppler radar (WSR - 88D) is proving to be a boon to the forecasting of severe weather features, such as wind shear, turbulence, and microbursts. For a discussion of these topics, refer to the *Federal Meteorological Handbook No. 11, Doppler Radar Meteorological Observations, Part D,* as well as chapter 12 of this manual.

SUMMARY

In this chapter we first discussed phenomena associated with thunderstorms; and then followed with a discussion of associated hazards aloft and at the surface. Methods used in the forecasting of thunderstorm movement and intensity were discussed. Following the discussion of thunderstorms, we covered tornadoes, tornado types, and waterspouts. A discussion on fog formation, types of fog, conditions necessary for various types of fog, and the use of the Skew T Log P Diagram in determining various fog parameters were presented. Next, we covered the processes involved in aircraft icing, intensities of icing, icing hazards near the surface, and aloft. Operational aspects of aircraft icing were then discussed, followed by types of icing and icing intensity forecasts. The last topics presented were turbulence characteristics, classification and intensities, the forecasting of turbulence near the surface, in-cloud, and CAT, as well as a discussion of the benefits of Doppler radar in forecasting turbulence.