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Weather Journeyman

Volume 2. Forecast Surface Weather Elements and Flight Weather Elements

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Material in this volume is reviewed annually for technical accuracy, adequacy, and currency. For SKT purposes the examinee should check the *Weighted Airman Promotion System Catalog* to determine the correct references to study.

IN VOLUME 1 of this course, you learned about climatology, the regional area forecast program, and forecast reviews. In this volume, you'll learn how to forecast weather elements. Forecasting weather elements, whether occurring at the surface or at flight level, will be part of your mission as a weather journeyman at an operational weather squadron (OWS). As a forecaster you will be responsible for writing forecasts for your area of responsibility; you'll also be issuing forecasts for other locations. Your forecasts will contain information on what effect or impact the weather elements will have on these various locations. The customer, who knows how the weather is going to impact their mission, has an advantage in planning their mission strategies. That's what your job is all about—to help customers exploit the weather for battle!

This volume is divided into 5 units. In unit 1 you'll learn about forecasting the major surface weather elements: clouds, precipitation, visibility restrictions, wind, temperature, and pressure. Unit 2 covers forecasting flight weather elements, such as turbulence, icing, and low-level wind shear. In unit 3, you'll explore convective severe weather. Unit 4 will provide a detailed look at non-convective severe weather. The fifth and final unit will explain how to interperet data from upper air soundings and how to use this data to improve your forecast.

To impart an analogy, you're about to read the "meat and potatoes" of our job. Forecasting weather elements is the essence of why weather journeymen are employed in the USAF. Take heart in this fact and make a concerted effort to learn the material in this volume because mission completion depends on it. Good luck and enjoy!

Code numbers on figures are for preparing agency identification only.

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This volume is valued at 33 hours and 11 points.

NOTE:

In this volume, the subject matter is divided into self-contained units. A unit menu begins each unit, identifying the lesson headings and numbers. After reading the unit menu page and unit introduction, study the section, answer the self-test questions, and compare your answers with those given at the end of the unit. Then do the unit review exercises.

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THE TERMINAL FORECAST is often mentioned as the end-product of all the work and research that you, as a weather journeyman do. The terminal aerodrome forecast, commonly referred to as the TAF, is your product to your customers, the "warfighter." As a weather journeyman at an operational weather squadron, there are few tasks more important than issuing TAFs for a base or other location. The TAF contains forecast information about clouds, visibility, weather, obstructions to vision, wind, temperature, pressure, turbulence, and icing. The question then arises: How would you, predict these parameters? How are the specific details that are forecast found?

Unfortunately, there are no fixed rules for forecasting any of these elements. Each time a forecast is made, you must weigh all the parameters, consider all the theories, evaluate all the atmospheric model data, and then come up with a value for each element in the forecast. You'll need to remember that applied or general meteorology takes the theories of dynamic and physical meteorology and applies them to situations as they occur in the atmosphere.

1–1. Cloud Forecasts

When you look at those puffy white cumulus clouds on a summer afternoon did you ever wonder why and how clouds form? Why are clouds present on some days and not others? How does water vapor transform from an invisible gas to frost on the grass? This section answers all of these questions and more. The answers to these questions can be explained by physics.

The word physics tends to scare many people, but physics is simply the study of physical properties and the interactions between matter and energy. There are instances when physics can get very complex, especially in instances where mathematical equations are used. However, the material presented in this section concentrates on atmospheric principles governed by the laws of physics, rather than mathematical formulas.

The following lessons discuss the three states or physical forms that matter can change to, especially water. We'll see how water droplets form and how they form clouds. You'll also explore the physical processes that assist the formation of clouds.

To begin, let's first discuss the three states of matter or why water can exist as a vapor, ice, or liquid, and how these changes in state occur.

201. States of matter

In the study of meteorology, there are three states in which matter may exist—solid, liquid, and gas. While all matter may exist in any of these states, the matter that concerns meteorologists most is water (H_2O). In the solid state, you know it as ice; in the liquid state, as water; and in the gaseous state, as a vapor. Each of these states of H_2O occurs naturally in our atmosphere, which makes understanding the changing of states so important.

Changes of state

A change of state is the process of matter changing from one state to the other. For example, water becoming ice or a gas. In meteorology, there are terms used for the various changes of state.

Vaporization

Vaporization is the change of a liquid to a gas. You may know this by the more common term—evaporation. When water molecules escape to the atmosphere as water vapor, they are said to evaporate or vaporize.

Condensation

Condensation is the reverse of vaporization—the process by which water vapor becomes a liquid. Clouds and rain are the result of water vapors condensing in the atmosphere.

Freezing

Freezing is the change of state from a liquid to a solid. You'll find that water in the atmosphere does not always freeze at 0°C. Water droplets in clouds may remain in the liquid state at temperatures as cold as –20°C. For water to freeze, it is necessary that some impurities be present in the water. Condensed water in the atmosphere lacks the mineral impurities (freezing nuclei) necessary for ice crystals to form.

Fusion

Fusion is the change of a solid to a liquid or, in other words, the melting of ice.

Sublimation

Sublimation occurs when a solid changes to a gas without an intermediate liquid stage. This is where ice changes directly into water vapor without becoming a liquid and under certain conditions is why snow disappears when temperatures are below freezing.

Deposition

Deposition is the process by which water vapor changes directly into a solid. An example would be frost forming; water vapor in the atmosphere is "deposited" on freezing surfaces.

Exchange of heat due to changes of state

When the three states of matter (solid, liquid, gas) change from one to another, energy (heat) is exchanged with the surrounding environment. This heat is either taken from or added to the environment surrounding the matter undergoing change.

The reason for this exchange of heat is that molecules must bond closer together or farther apart when matter changes states. For example, for liquid water to become a solid the molecules of water must bond closer together; for water vapors to form, they must move far apart. The heat that is exchanged because of these changes of state is called *latent heat*. Figure 1–1 shows the changes of state processes.

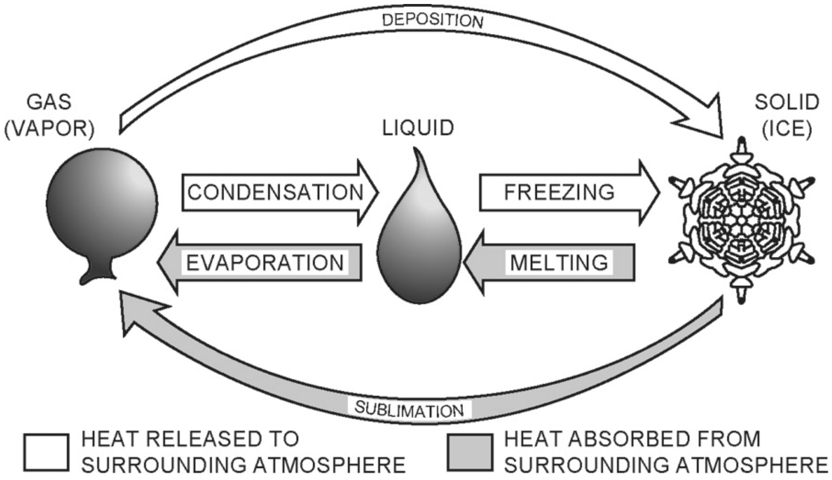

Figure 1–1. Changes of state processes.

Changes in state that cause heat absorption

The changes in state that requires absorption of heat from the environment and a transfer of that heat to the mass are the changes from a solid to a gas and all points between. These are *vaporization, fusion*, and *sublimation*.

When any of these processes occur in the atmosphere, heat is removed from the atmosphere, thus causing cooling. For example, on a hot summer day, air temperatures are several degrees cooler near a large body of water. This is due to evaporation or vaporization.

Changes in state that release heat

The changes of state that release heat to the environment are the changes from a gas to a solid, and all intermediate changes. These include *condensation, freezing,* and *deposition*.

When any of these processes occur, heat is added to the atmosphere, causing a rise in the temperature. For example, when saturated air is cooled adiabatically, the moisture in the air condenses and releases heat to the atmosphere. This is why the moist adiabatic lapse rate is less than the dry adiabatic lapse rate. This is also why, if condensation occurs, such as in the formation of fog, the temperature rises slightly or does not fall as quickly.

Latent heat exchange plays a major role in the formation of clouds and precipitation. We'll discuss the effects of latent heat a little later in this volume.

202. Cloud microphysics

Clouds form when the atmosphere is saturated. Saturation occurs when the air holds the maximum amount of water vapor possible. At saturation, the temperature is equal to the dew-point temperature ($T = T_d$); the vapor pressure is equal to the saturation vapor pressure ($e = e_s$); and the relative humidity is 100 percent. Saturation is a state of equilibrium as it was just explained. Water is constantly evaporating and condensing. The amount of water vapor molecules in the air is constant, but the actual molecules are constantly changing their state (or phase).

Cloud condensation nuclei

The cloud condensation nuclei (CCN) are necessary for water vapor to convert to liquid. These nuclei are composed of sea salt, volcanic sulfates, forest fire smoke, clays, and other fine terrestrial dust particles. These are particles on which water vapor condenses to form water droplets. The most effective nuclei are hygroscopic (water attracting) and water soluble (can be dissolved by water). If the CCN are water soluble, the resulting water droplet is not pure water but a solution. The concentration of the solution (mass of CCN to volume of water in the droplet) depends upon the size of the CCN and the amount of vapor that has condensed onto it (i.e., the volume of the droplet). Cloud droplets are initiated by the condensation of water vapor on the CCN. These droplets continue to grow as water vapor condenses onto the CCN.

Droplet initial growth

The air immediately surrounding all the droplets is said to be saturated ($e = e_s$). Initially, whether a droplet grows or not depends on the saturation vapor pressure (e_s) of the droplet and its surrounding environment. In figure 1–2, the example on the left shows initial growth of a droplet due to the $e_{s\ env} > e_{s\ droplet}$ and the droplet grows due to condensation. While the example on the right shows that the $e_{env} < e_{s\ droplet,}$ therefore, the droplet shrinks through evaporation.

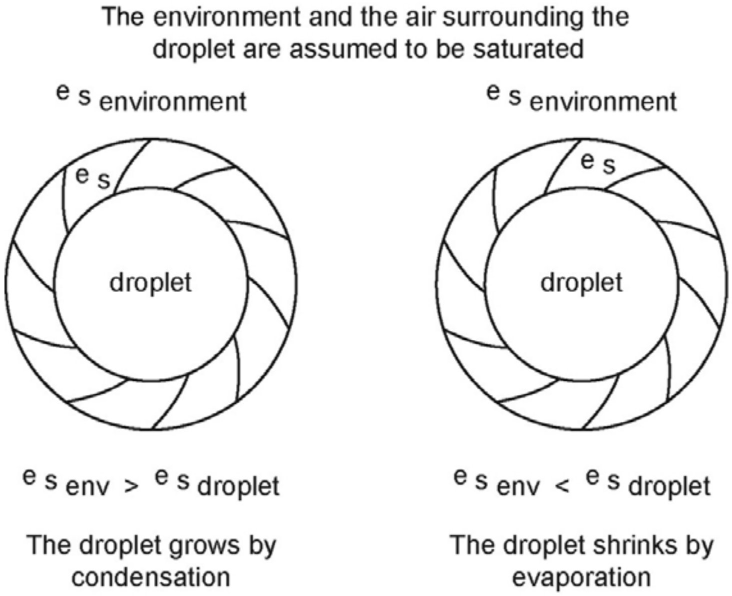

Figure 1–2. Comparisons between e_s of a droplet and its environment.

Differences in saturation vapor pressures

After their initial growth, droplets have different saturation vapor pressures based on their concentration and size. Droplets grow due to solute and curvature effects. Before we discuss these effects, it is important to note that the solute and curvature effects occur between other droplets and *not* the environment.

Solute effect

When droplets consisting of higher concentrations of solution, such as CCN, have a lower saturation vapor pressure than the more pure droplets the solute effect occurs and the formula is (e_s pure > e_s solute). For example, let's compare the two droplets in figure 1–3. Droplet "b" is less pure than droplet "a", therefore $e_s a > e_s b$. Droplet "a" shrinks due to evaporation at the expense of droplet "b" growing due to condensation. Flow between the two factors is from higher to lower pressure.

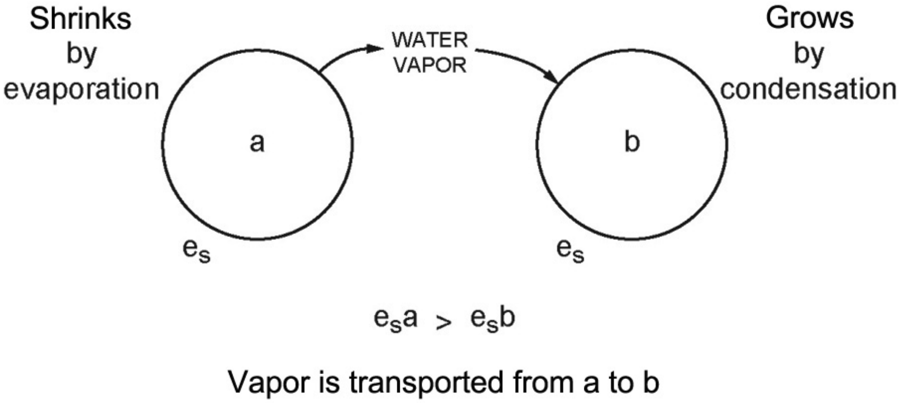

Figure 1–3. Example of solute effect between two droplets.

Curvature effect

When small droplets, which are tightly curved, have a higher saturation vapor pressure than larger droplets, which are less curved, it is called the curvature effect. The formula is ($e_{s\ small} > e_{s\ big}$). For example, if we have two droplets, one small and one large, the larger droplet grows at the expense of the smaller droplet. As we stated earlier, flow between the two factors is from higher to lower pressure.

Collision and coalescence

Once droplets grow large enough to fall, they can collide with other falling droplets and merge to form even larger droplets. The process where the droplets collide and merge is called coalescence. This is the most efficient method of droplet growth (*much* more efficient than condensation). While this process is occurring; the droplets still grow by condensation.

Factors that affect efficiency

Different conditions affect the efficiency of the collision-coalescence process. Let's examine three factors that affect the efficiency of this process.

Residence time in the cloud

The longer a droplet remains in the cloud, the greater chance it has to interact or coalescence with other droplets. Thick clouds, such as nimbostratus clouds, provide a greater distance for the droplet to fall, thereby increasing the coalescence time. Updrafts associated with cumulus and cumulonimbus clouds account for transporting droplets to upper portions of the cloud. These updrafts also increase the distance and coalescence time of the droplet (fig. 1–4).

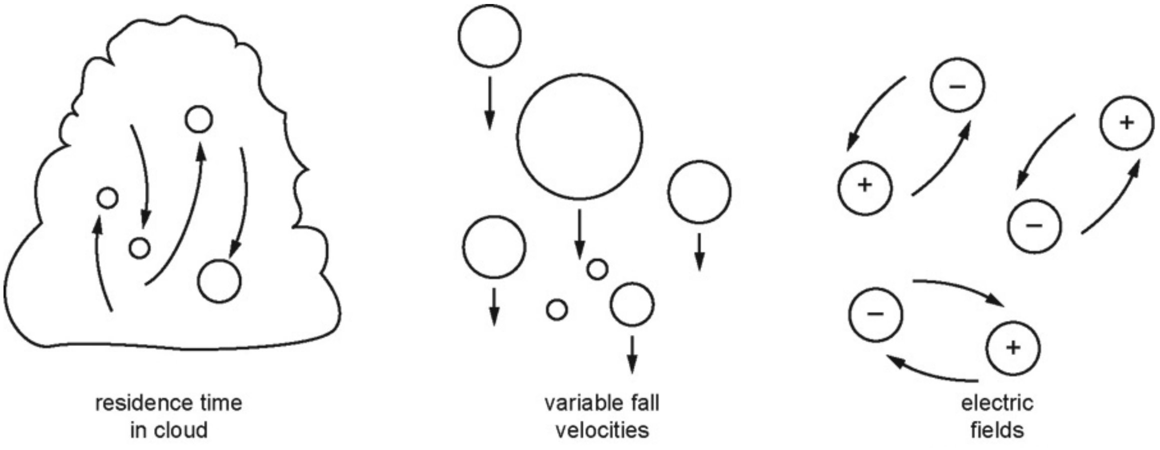

Figure 1–4. Factors affecting collision and coalescence efficiency.

Variable droplet fall velocities

If droplets are falling at different rates (velocities), then the faster falling droplets catch up, collide, and coalescence with the slower falling droplets. Larger droplets fall at a faster rate than smaller ones. A wide spectrum of droplet sizes is indicative of the variable fall velocities.

Electrical fields

Droplets exposed to an electric field develop a net charge, either positive or negative. Droplets with opposite charges attract each other. The electrical attraction causes more collisions between droplets to occur.

Initiation and growth of ice crystals

Ice nuclei (IN), which are similar to CCN, are necessary for liquid or vapor to form ice crystals. Water droplets at temperatures below 0°C are called *supercooled*. These droplets exist in a liquid state. If these droplets encounter an ice crystal, they automatically freeze onto the ice crystal. Once a crystal is formed, it grows by deposition (conversion of water vapor to ice). This is because the saturation vapor pressure over ice is less than that over water ($e_{s\ ice} < e_{s\ water}$).

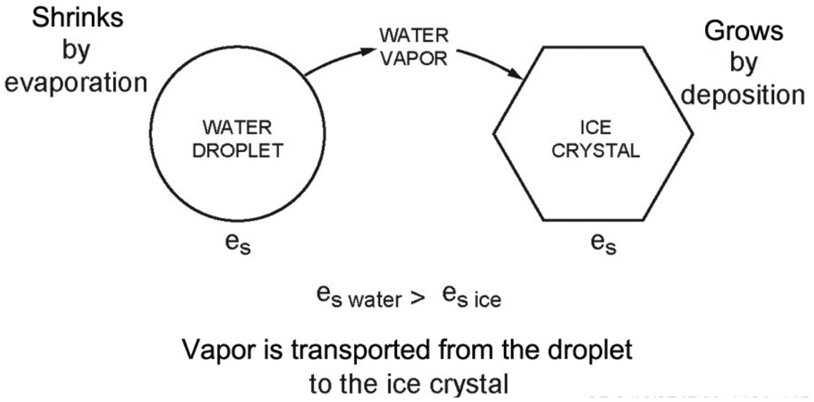

Figure 1–5. Interaction of cloud droplets and ice crystals.

Therefore, the ice crystals grow at the expense of water droplets. Figure 1–5 shows an example of a supercooled water droplet on the left and an ice crystal on the right. Since the e_s over ice is less than that of a water droplet, the ice crystal grows by deposition.

Once the crystal grows large enough, it undergoes a process similar to collision-coalescence. The crystal can collide with other cloud droplets. If the cloud is colder than 0ºC, the droplets freeze on contact. The crystal can also collide with other ice crystals. It is the collision of ice crystals that cause the formation of large snowflakes.

203. Cloud dynamics

In the previous lesson we covered the very small-scale elements that are conducive to cloud formation. In this lesson, we investigate the larger, more general atmospheric dynamics that must also be available for clouds to form. We conclude by probing the dynamics that lead to cloud dissipation.

Generally, cloud types and the number of clouds are determined by the amount of atmospheric moisture available, temperature, stability, and lifting mechanisms. Let’s explore these general cloud dynamics.

Atmospheric moisture

First and foremost, there must be enough available atmospheric moisture present for clouds to form. No amount of lift or cooling can produce clouds if sufficient moisture is not present. Later in this unit we discuss atmospheric moisture and its measurement in greater detail.

Temperature and stability

Cooling processes such as adiabatic cooling and radiational cooling are the principle condensation mechanisms. Adiabatic cooling is the most effective means of cooling water vapor until it condenses.

Lifting mechanisms

Lifting mechanisms include orographic lift, frontal lift, and low-level convergence.

Orographic lift

Horizontal motion is converted to vertical motion proportional to the slope of the terrain. Even relatively flat terrain may have slopes of 1 mile vertical to 200 miles horizontal. Craftsmen must have a thorough grasp of geographic details over the forecast region to assist in the cloud forecast. This phenomenon occurs quite often on the windward side of mountains.

Frontal lift

The amount of frontal lift depends on the frontal slope. The frontal slope is significant because it represents the potential lift of a front. A steep slope suggests the frontal lift could be strong if the wind flow forces air to ascend the frontal surface. Conversely, a shallow slope suggests the frontal lift would be weak if the wind flow forces air to ascend the frontal surface.

Low-level convergence

Low-level convergence is a measure of the rate of the net addition of mass into a volume at a given point. This convergence can be directional, speed, or a combination of both. The mass of air converging in the low levels just above the surface must go somewhere. Since it cannot go into the ground, it rises, giving lift to the air.

Directional convergence

Directional convergence is the coming together of wind flow, which results in mass being added to an area.

Speed convergence

Speed convergence is caused by winds rapidly decreasing in speed downstream. The higher speed winds push mass into an area faster than it can be removed by the slower speed winds, thus increasing the mass.

Cloud dissipation

Just as there were a number of reasons for cloud formation, cloud dissipation also has more than one reason for occurrence. Evaporation of water droplets and the sublimation of ice crystals (conversion from ice directly to vapor) are ways that clouds can dissipate. These dissipation dynamics are caused by two mechanisms—decreasing the moisture and increasing the air temperature.

Decreasing the moisture

Decreasing the moisture of the cloud can be accomplished in two ways. The first includes the cloud “raining itself out.” The precipitation evaporates while it falls or after it reaches the surface.

Moisture can also be decreased by a process known as dry-air entrainment as shown in figure 1–6. Dry-air entrainment is the process by which the outer edge of the cloud mixes with dry air outside of the cloud. This outer edge of cloud then evaporates or sublimates with the drier air.

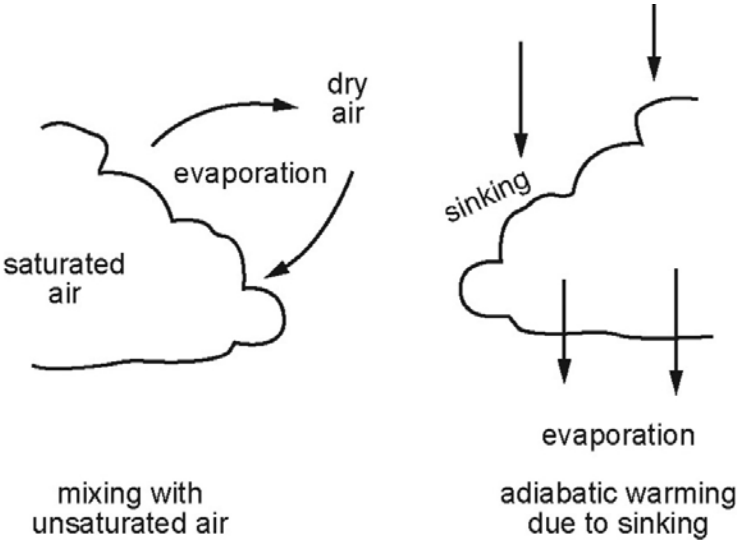

Figure 1–6. Cloud dissipation processes.

Increasing the air temperature

Increasing the air temperature within a cloud can be accomplished two ways. The first includes adiabatic warming associated with downward vertical motions (sinking air). These motions occur with either subsidence or as a leeside mountain effect where moist air is compressed and warmed. The cloud either evaporates or sublimates.

Secondly, an increase in air temperature can be achieved by warming from below. For example, higher air temperature is the primary mechanism for the dissipation of radiational fog. Solar heating is what increases the air temperature of fog (a cloud on the ground), causing it to dissipate.

Now that you have a good understanding on the dynamics behind cloud formation, it’s time to cover one of the most difficult tasks facing weather forecasters—cloud forecasting.

204. General tools for forecasting clouds

Cloud cover is one of the most significant phenomena to affect aviators and military strategic planners. Even in the age of all-weather aircraft and precision-guided munitions, the fact remains—if the pilot can’t see the target, then it is difficult to deliver the munitions on the intended target. The Gulf War in 1991 and NATO’s Operation Allied Force in the Balkans in 1999 proved that significant cloud cover reduces tactical targeting opportunities and degrades bombing accuracy.

Before we begin to discuss specific rules and techniques to forecast clouds, let's look at four general tools that help you in your quest for a cloud forecast that's on target. These tools help you, the novice forecaster, to start formulating your cloud forecast and avoid the "I don't know how to start" syndrome.

Climatology

Climatology is the science that deals with climates and their phenomena. It is a time-proven method that works well for forecasting clouds and ceilings and it should be used when beginning the forecast process. Derived from decades of observational data, climatology helps get you "in the ballpark" with your forecast and avoid gross errors in the forecast of cloud and climate parameters. For example, climatology for your location indicates that the hottest temperature ever for the month of August was 110°F. Forecasting a high temperature to be greater than 110°F would be unlikely unless you had strong atmospheric indicators to the contrary.

The use of climatology can guide you to the most sensible range of values for a particular forecast parameter. There are several sources for climatological data. The 14th Weather Squadron provides numerous climatology products to assist you, such as the products we discussed in the previous volume of this course. Remember to bookmark the 14th Weather Squadron's website as new climatology products continue to become available. The following is a brief description of some of these products that are available.

Wind stratified conditional climatology

Wind stratified conditional climatology (WSCC) provides you with information on ceiling and visibility categories in a tabular format. WSCC tables are more commonly referred to as "CC tables" in the weather community. These tables are dependent upon the initial conditions, from which they provide you with valuable prognostic information. They indicate either persistence or change in the characteristics of the ceiling and visibility based on the initial ceiling, visibility, wind direction, and time of the day.

Modeled curves

Modeled Curves (MODCURVES) is a program that uses climatology as a guide to provide the temperature baseline for the time of year and time of day. Adjust the display to reflect current or expected weather conditions that may impact cloud forecasts (i.e., temperatures, winds, etc.).

Surface Observation Climatic Summaries

Surface Observation Climatic Summaries (SOCS) is a multi-part climatic brief. At least 5 years of observational data are required to create a SOCS and it can update whenever 10 additional years of data are added to the database or upon request. The SOCS summarizes hourly observations including the "summary of the day" data for a given weather station.

SOCS summarizes observed data in eight categories: atmospheric phenomena; precipitation, snowfall and snow depth; surface wind; ceiling, visibility, and sky cover; temperature and relative humidity; pressure; crosswind summaries; and degree days.

Model and centralized guidance

Today's computers have the capability to generate a great amount of data from several atmospheric models. The Air Force Weather Agency (AFWA) and the National Weather Service (NWS) produce centralized charts which can help you forecast clouds. The NWS weather depiction chart like in figure 1–7 allows the forecaster to see cloud conditions upstream from their location.

AFWA produces several graphic products based on the Weather Research and Forecast (WRF) model that can assist you in forecasting clouds (figs. 1–8 and 1–9).

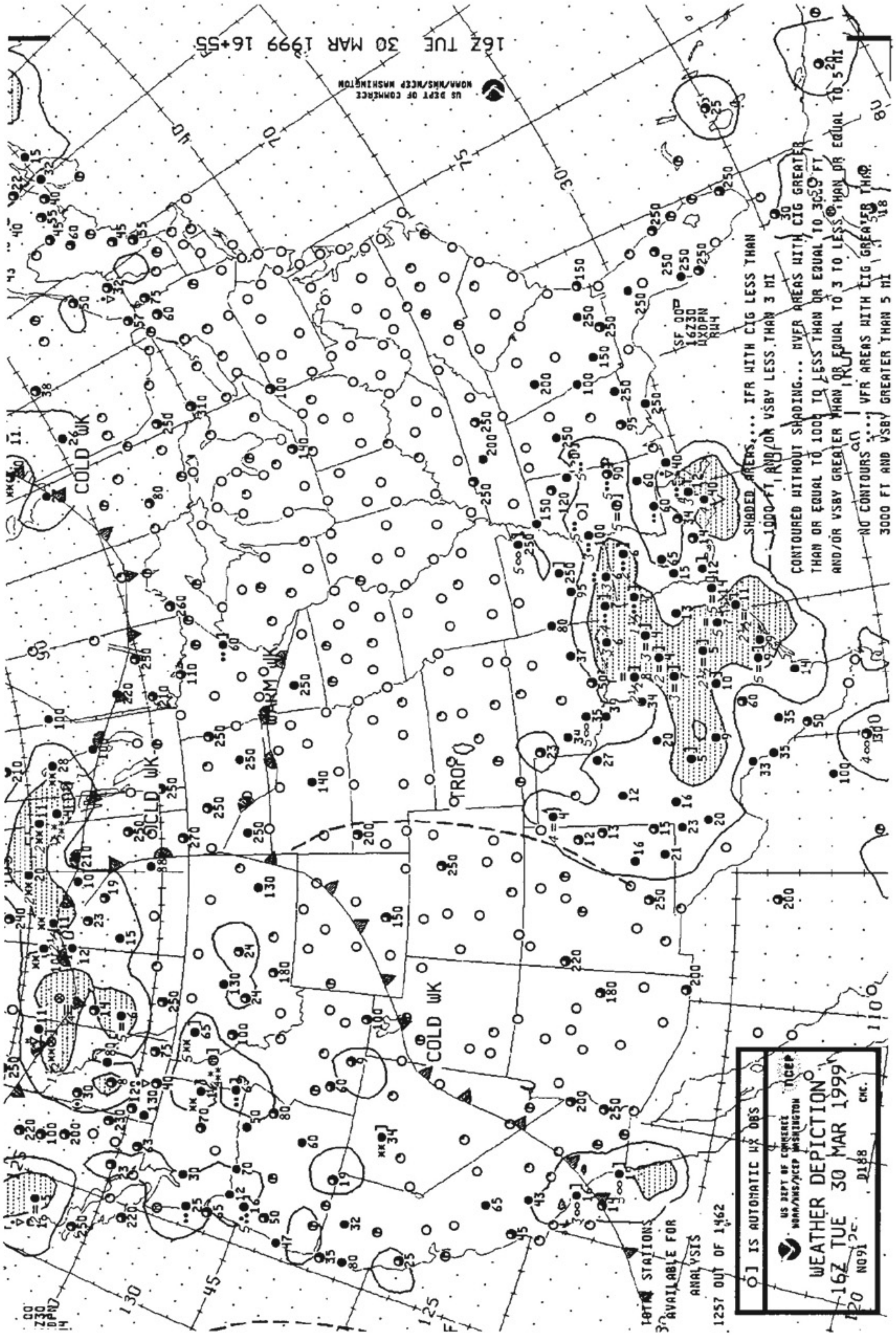

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Figure 1–7. Weather depiction.

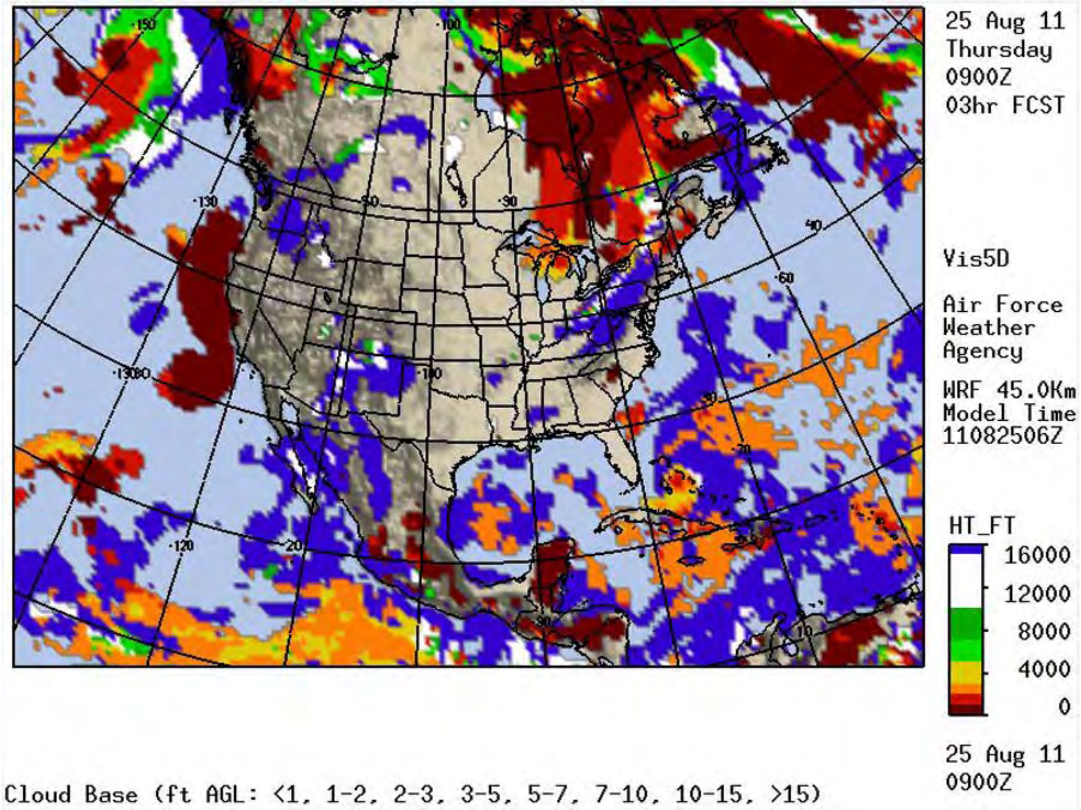


Figure 1–8. WRF cloud base forecast forecast.

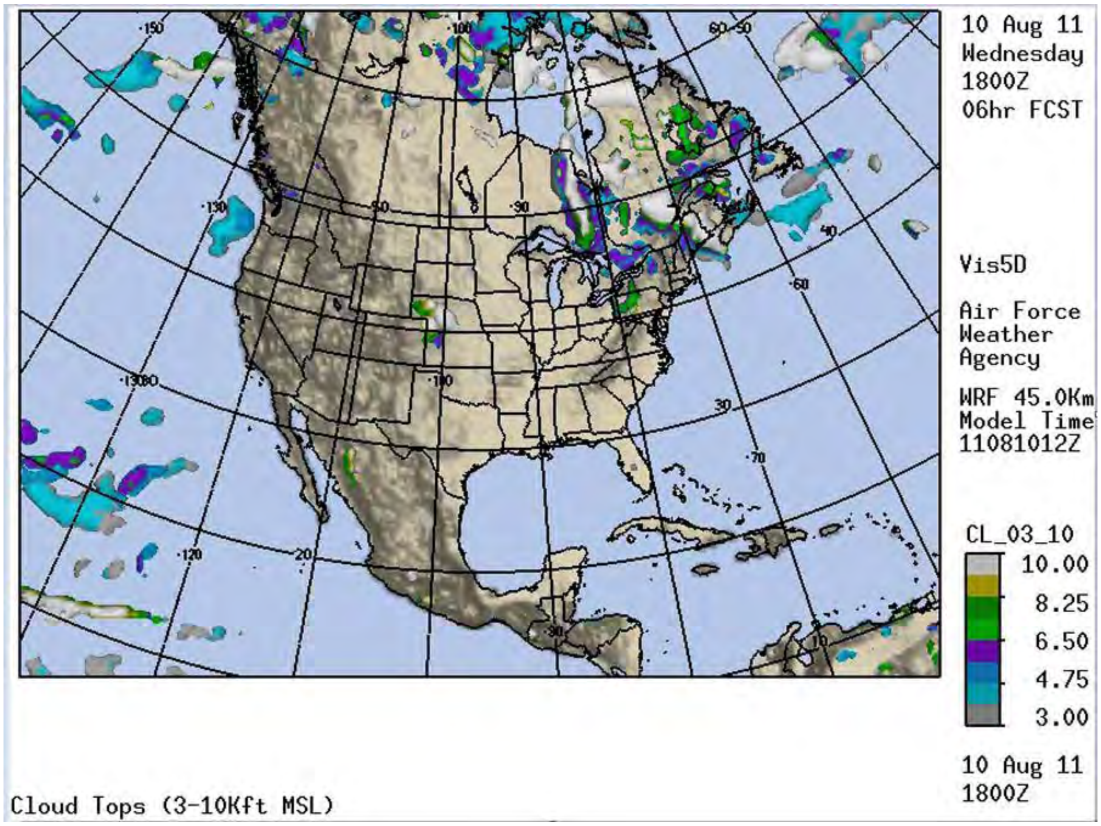


Figure 1–9. WRF cloud top forecast (3–10K MSL).

One of the phenomena that numerical products provide forecast guidance on is cloud amount and relative height range.

Model output statistics

Model output statistics (MOS) guidance is an excellent tool for cloud forecasting. MOS derives forecasting relationships by correlating past model output with station climatology. Climatology is the key ingredient. If the weather is abnormal for a particular time of year, the MOS data is biased towards climatology (average conditions), and may not be as accurate during these times. As always, it’s imperative to verify the model before using MOS. The table below is a brief explanation of the MOS header formats.

Header D	escription
CIG	Ceiling height forecast for model output statistics.
CLD	Opaque cloud cover forecast for specified time (overcast, broken, scattered, few, clear).
T06/T12	Thunderstorm/conditional severe thunderstorm probability for 6-hour and 12-hour period.
Q06/Q12	Precipitation amount forecast for 6- and 12-hour periods.
OBV	Obstruction to vision forecast for a specified time (H - Haze, F - Fog, BR- Mist and N - No haze or fog).

Below is the ceiling height code for model output statistics.

Ceiling Height Code	Cloud Height
1	<200 ft
2	200 to 400 ft
3	500 to 900 ft
4	1,000 to 1,900 ft
5	2,000 to 3,000 ft
6	3,100 to 6,500 ft
7	6,600 to 12,000 ft
8	>12,000 ft or unlimited ceiling

Forecast relative humidity values

Information for forecasting relative humidity values can be in Forecast Output United States (FOUS) bulletins located on the Joint Air Force and Army Weather Information Network (JAAWIN) website. The forecast relative humidity (RH) values in the Global Forecast System (GFS) and North American Model (NAM) numerical bulletins specify forecast relative humidity percents for layers of the atmosphere above a station. Relative humidity for three different layers is specified and is identified by R1, R2, and R3 as explained below in the nested grid model (NGM) numerical output bulletin.

Identification Number	Explanation
R1	Is the relative humidity of the surface to 1,000 foot layer centered near 500 feet above ground level (AGL).
R2	Is the relative humidity of the 1,000 to 17,000 foot layer centered near 9,000 feet AGL.
R3	Is the relative humidity of the 17,000 to 39,000 foot layer centered near 28,000 feet AGL.

Use this information (after initializing and verifying the model) to help determine cloud amounts and levels through the 48-hour forecast point.

NOTE: This technique doesn't necessarily help to determine the cloud base; it only indicates that clouds may exist in that layer.

You need to remember that these percentages are layer averages. Shallow cloud decks may be present, but aren't identified because shallow layers of relative humidity values become "averaged out" over the entire layer. This is especially true with the R2 and R3 layers.

Weather Research and Forecast based meteograms

The WRF-based meteogram is a time cross-section forecast trend of eight meteorological parameters including clouds (fig. 1–10). The initial time is on the left side of the diagram, and forecast times extend to the right until the end of the forecast period.

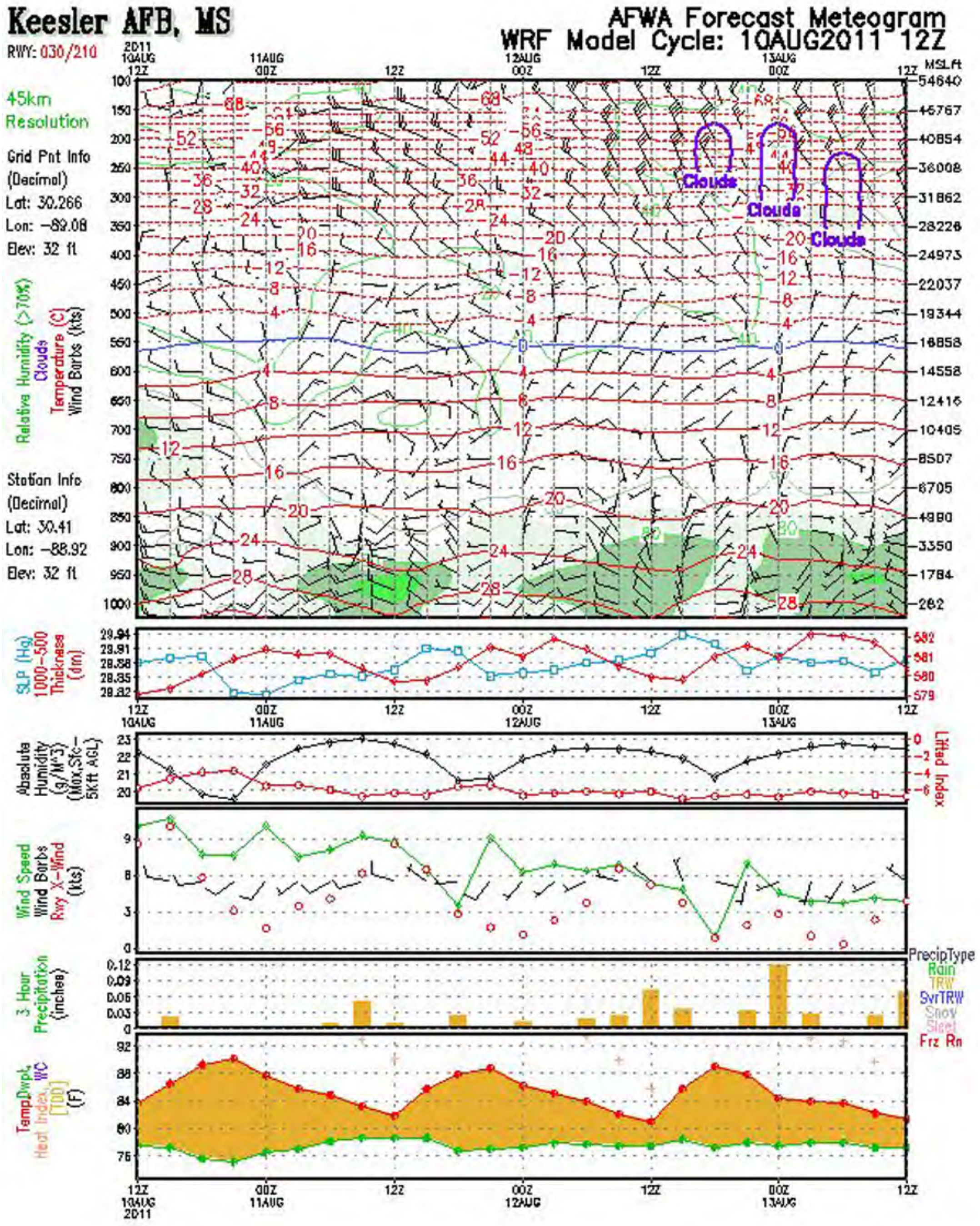


Figure 1–10. AFWA meteogram.

The top time cross-section of a meteogram displays cloud forecasts which are outlined in purple contours, dew point depressions, and relative humidity greater than 70 percent. The WRF algorithms that generate the cloud forecast work best in convective situations and don't do well when predicting cirrus clouds. Dew point depressions less than 6° that are higher than 24,000 feet are displayed and can be used to forecast cirrus clouds.

Extrapolation

This technique refers to the forecasting of a weather feature based solely on its recent past movement. A detailed study of short-range extrapolation with examples is covered in a later section of this volume.

To use extrapolation techniques in short-range forecasting (0 to 6 hours), determine the positions of fronts and pressure systems, their direction and speed of movement, precipitation and cloud patterns that might affect the local terminal, and the upper-level flow that affects the movement of these weather patterns.

To forecast clouds by extrapolation, simply advect them downstream. For an analysis of clouds by heights or type, using a satellite or a nephanalysis aids tremendously, especially if the previous forecast (that can be used for continuity) was annotated.

Weather radar

Doppler weather radar can detect cloud layers by sensing large ice crystals present in middle and high-level clouds, and refractive index gradients associated with all clouds. Typical reflectivities are between –12 to +15 decibels (dBZ), but may range as high as +20 dBZ. Using the following products may help in identifying and advecting clouds.

NOTE: Care should be exercised because in some instances radar data should not be used without other supporting products.

Velocity Azimuth Display Wind Profile

When the VWP displays "still not clear" look for invading upper-level wind barbs that signify clouds are progressively advancing towards the radar data acquisition (RDA) unit.

Reflectivity

On the base reflectivity (R) product you can determine the height, thickness, and location of clouds. To determine these parameters use the following steps:

1. Determine and use the best elevation that depicts the cloud layer.
2. Place the cursor on the edge of the echo closest to the RDA unit and note the readout of azimuth, range, and elevation in mean sea level (MSL) of the base of the layer.
3. Determine and use the highest elevation that shows the cloud layer.
4. Place the cursor on the edge of the echo farthest from the RDA and note the readout of the azimuth, range, and elevation in MSL of the top of the layer.

You may have to repeat the preceding technique if the cloud base or top is ragged and not uniform in height. By repeating this technique, you obtain an average height and thickness of the cloud.

Reflectivity cross section product

The reflectivity cross section (RCS) product helps to determine the top of a cloud layer and its depth, depending on the distance from the radar and the viewing angle. Keep in mind, the resolution of the base reflectivity product is better in determining the top of a cloud layer and its depth. The RCS product integrates returns from the surface up to 70,000 feet and tends to exaggerate the cloud layers.

Echo tops

The echo tops (ET) product can provide an indication of the top of a cloud layer using the threshold value of 18 dBZ. Always use the reflectivity product in conjunction with ETs to determine the existence and extent of the cloud layers.

The forecasting tools depicted in this lesson are applicable to many different cloud types. In the next lesson, we explore techniques specific to convective clouds.

205. Convective clouds

The occurrence of convective or cumuliform clouds is forecast in much the same way as stratus clouds. The main difference is the added consideration of atmospheric stability. Cumulus clouds result from vertical currents in the air, and these vertical currents occur in an unstable or conditionally unstable atmosphere. If temperature and moisture content indicate possible cloudiness and the existing lapse rate exceeds the moist adiabatic lapse rate, the cloud that develops is a cumulus.

The amount of cloudiness is determined from the relative humidity (RH), unless an overcast condition is indicated. A solid overcast of cumulus clouds is highly unlikely. Therefore, the cloud forecast should include stratocumulus or nimbostratus clouds along with cumulus clouds when overcast conditions are expected.

One reason for this condition is that updrafts in which cumulus clouds develop are not normally widespread enough to produce a large mass of convective clouds, but rather the updrafts are interspersed with adjoining downdrafts in which no clouds or stratiform clouds are observed.

Bases of ordinary cumulus clouds are usually found existing between the lifting condensation level (LCL) and the convective condensation level (CCL). They tend to favor the CCL where thermal currents are the greater factor in the cloud development and approaching the lower LCL when mechanical lift is the more predominant factor.

The LCL is the level in the atmosphere where condensation occurs with a parcel of air due to adiabatic cooling. The LCL is already calculated and displayed on Skew-Ts available from the weather distribution system or the JAAWIN website.

Bases of altocumulus clouds are usually forecast to occur where the RH falls within the critical values for formation and where the atmosphere becomes conditionally unstable. It is possible to have a moist layer of stratus type clouds topped with cumulus type clouds, the bases of which are adjacent to the tops of the lower stratus deck. This occurs when, within a moist atmosphere, the temperature lapse rate changes from stable to unstable.

The heights of the tops of convective clouds are difficult to forecast because so much depends on the amount of moisture contained within the layer where the clouds are formed and on the degree of instability throughout the layer in which the clouds are growing. Cumulus tops are not confined to the level where the RH drops below 65 percent. Strong instability can, and does, send the tops of cumulus clouds through dry layers at higher levels; a characteristic common in thunderstorms.

The tops of cumulus clouds are usually limited to that point where the moist adiabat, passing through the cloud-base temperature, crosses the free-air temperature value at a higher level. However, in a situation of large amounts of moisture and in strong convective currents, cumulus tops may penetrate a stable layer and continue to build above if the stable layer is not too thick and is topped by another conditionally unstable layer. Also, not all cumulus tops build to a point where the cloud-base moist adiabat recrosses the free-air curve. Cumulus tops, through a layer of rather dry air whose lapse rate is more than the moist adiabatic, are likely to be irregular, with scattered or isolated tops reaching elevations far more than most of the tops. However, the tops of convective clouds usually do not exceed the height of the tropopause

Another technique to use to forecast bases of convective clouds is to use the current dew point depression and the table below. (The table is in a two column format. Read it from the left hand column from top to bottom and then the right hand column from top to bottom.) This table is not suitable for use at locations situated in mountainous or hilly terrain and should only be used when clouds are formed by active surface convection in the vicinity. Use this table with caution when the surface temperature is below freezing due to possible inaccurate dew points at low temperatures.

DPD (°C)	Estimated Cumulus Height (ft)	DPD (°C)	Estimated Cumulus Height (ft)
0.5	200	7.0	2,800
1.0	400	7.5	3,000
1.5	600	8.0	3,200
2.0	800	8.5	3,400
2.5	1,000	9.0	3,600
3.0	1,200	9.5	3,800
3.5	1,400	10.0	4,000
4.0	1,600	10.5	4,200
4.5	1,800	11.0	4,400
5.0	2,000	11.5	4,600
5.5	2,200	12.0	4,800
6.0	2,400	12.5	5,000
6.5	2,600		

During days when the air is unstable and the moisture is plentiful, you’ll get a chance to put the knowledge learned in this lesson to good use. Now let’s look at the opposite side of the spectrum, which is forecasting clouds when the air is stable—stratus clouds.

206. Stratus clouds

Stratus clouds are associated with a stable atmosphere in which vertical currents are, at most, very light. You can forecast stratus clouds if you use the dew point spread or RH considerations discussed earlier; you may also use synoptic association to a great extent. The RH or dew point spread criterion applies to forecasting stratus clouds. Bases of stratus clouds are usually forecast near the level where the RH reaches or exceeds 65 percent, and the tops are bounded by the 65 percent RH value as the moisture decreases. Between these levels, the amount of cloud cover depends on the RH at the various levels.

An exception to this rule is with high RHs in the layer from the surface of the earth up to several thousand feet. Here, the bases of the clouds are not touching the surface, but occur at some level above the surface. This level tends to be fairly constant for a given synoptic pattern at any given air base and, generally, is something less than 1,000 feet.

Synoptic association plays a big part in forecasting stratus-type clouds. Stratus clouds occur where the atmosphere is stable and the moisture content is high enough to produce the cloud. Onshore winds or large areas of slight convergence (such as those observed over the eastern and central United States when under extensive southerly flow of warmer air over a cooler surface) usually produce stratus clouds when the contours are straight or slightly cyclonically curved.

No stratus or only scattered patches occur when the curvature is anticyclonic. If considerable diurnal heating occurs, cumulus clouds are possible. Also, stratus clouds are common along warm fronts where the air is stable and lift is being produced mechanically.

Wind speed

As mentioned previously, wind speed is a determining factor for the formation of stratus clouds or fog. It's important to note that there is no one particular wind speed that can be used to determine the formation of stratus clouds. Like many other situations in weather, the wind speed which determines stratus formation is dependent upon local conditions including topography. Typically, stratus clouds form due to nocturnal cooling with geostrophic wind speeds in excess 15 to 20 knots at an inland location. At a coastal location wind speeds exceeding 10 to 15 knots are needed and winds greater than 30 knots are needed for stratus cloud formation in a deep valley.

Dissipation of stratus clouds

An empirical rule states that stratus clouds dissipate when the RH required for its formation decreases below 65 percent. The two methods that dissipate stratus clouds are heating of the atmosphere or advection which results in mixing of a dryer air mass.

In situations where stratus clouds form in a stagnant air mass or advects in from over a body of water the temperature at which stratus dissipates or "burns off" is also the temperature at which it begins to form again. In other words, if a stratus deck cloud dissipates at a temperature of 73°F/23°C in the morning, you can look for the stratus to reform when the temperature decreases below 73°F/23°C in the evening.

You can calculate when stratus clouds will dissipate by utilizing an upper air sounding that is representative of the air mass over your location. Figure 1–11 illustrates this process. By finding the average mixing ratio through the stratus cloud and finding where this average crosses the temperature curve, you can find out the thickness of the stratus layer. The thicker the stratus the higher the temperature needed to completely dissipate it. By following the dry adiabat, which intersects the temperature curve at the top of the stratus layer, and following it down to the surface you can determine the temperature at which the stratus completely dissipates.

Fog, which is covered later, has basically the same formation process as stratus clouds. The major difference is the level of the inversion and the wind speed.

Figure 1–12 illustrates the effect of the wind on the formation of stratus. If the wind is too light, fog forms and if it is too strong, the stratus cloud is probably blown away.

Stratus clouds associated with the low-level jet stream

The establishment of a low-level jet (LLJ) stream over the Midwest can be a mechanism by which stratus clouds form over the Midwest. In the fall and winter months, moisture advection originating over the western Gulf of Mexico and Texas can advect warm, moist air northward over the Midwestern United States to the Great Lakes as shown in figure 1–13. This warm moist air advection over cooler land frequently results in low stratus clouds, dense fog, drizzle, or freezing precipitation. The LLJ stream usually is found ahead of a low-pressure system front advancing toward the Mississippi River Valley with a high-pressure system over the eastern seaboard. The advection of this stratus cloud northward can cover 800 to 1,000 miles in 24 hours, which is considered a very fast moving formation. Monitoring the low-level winds, 925-millibar (mb), 850mb, weather depiction products and satellite imagery help in identifying this situation.

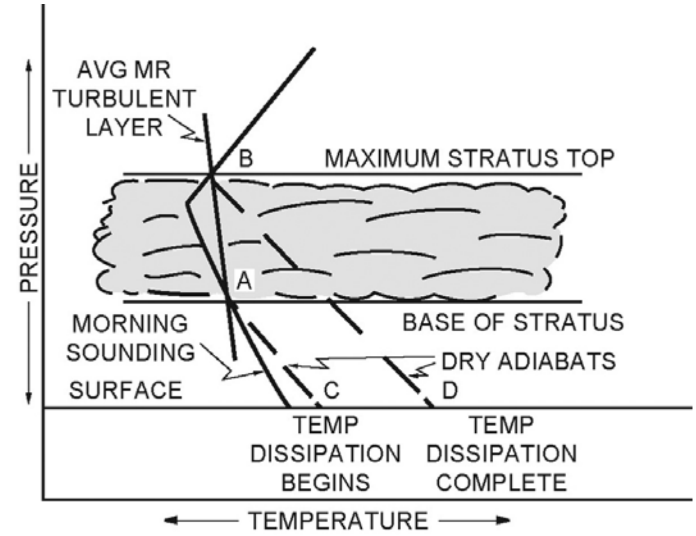


Figure 1–11. Dissipation of stratus clouds using mixing ratio and temperature.

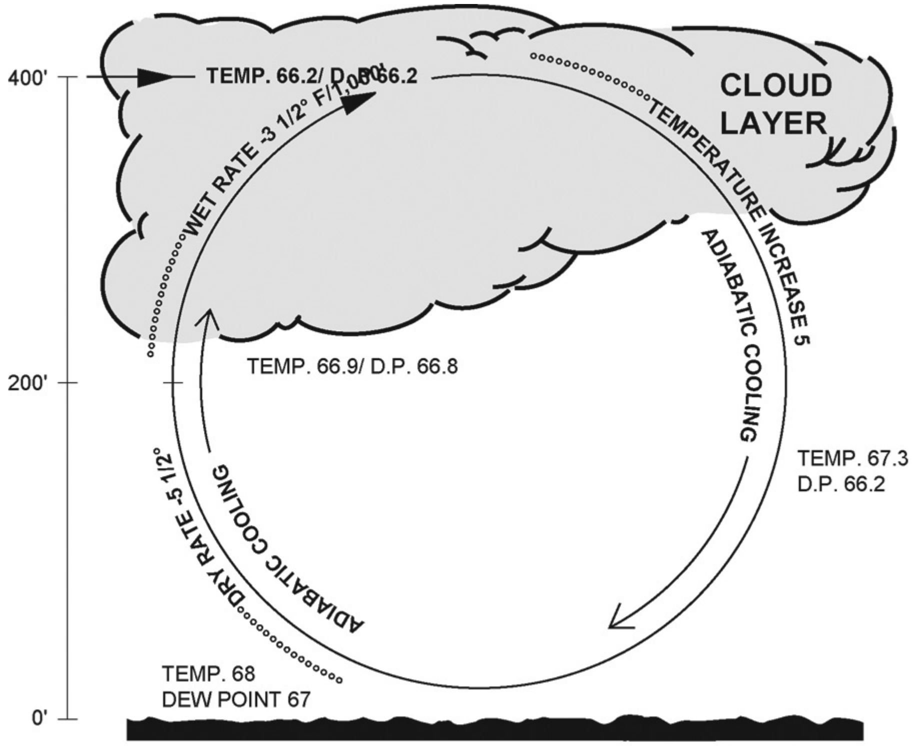


Figure 1–12. The effect of wind on developing stratus clouds.

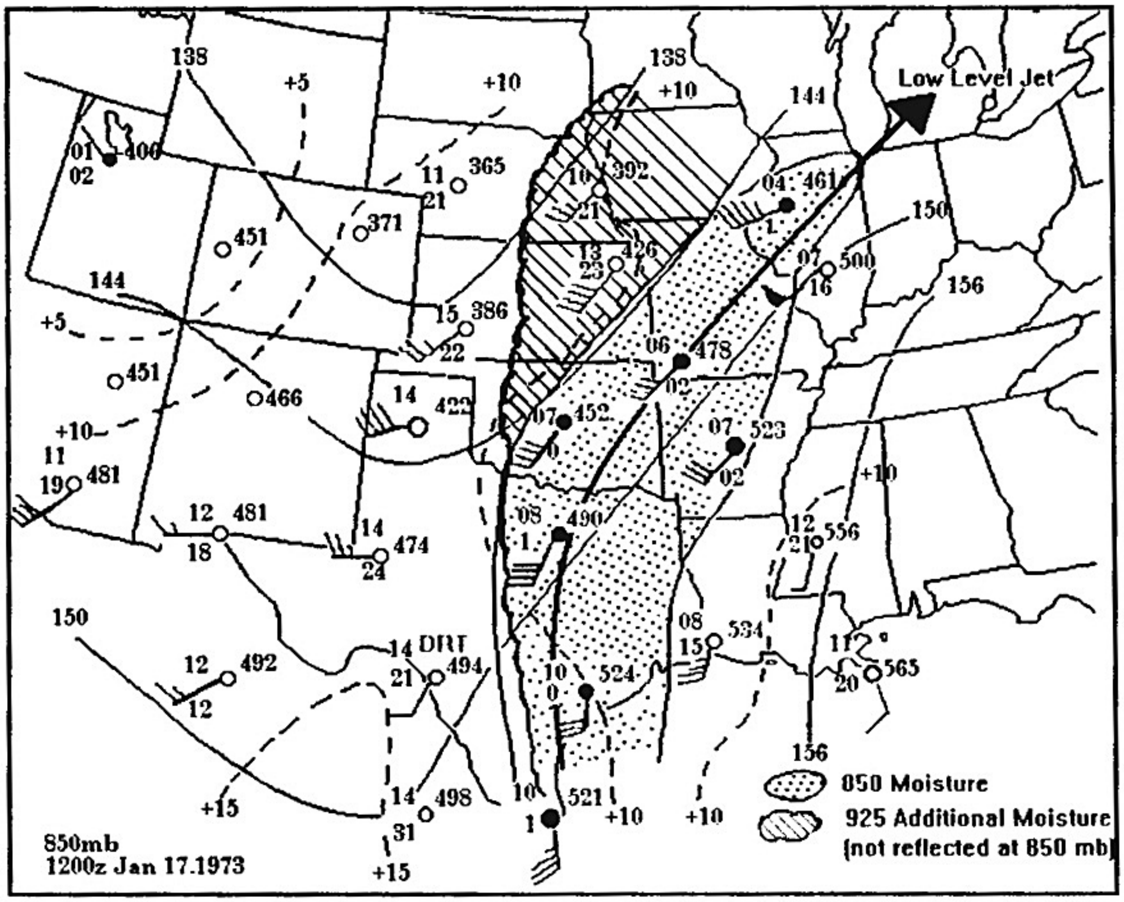


Figure 1–13. Advection of Gulf stratus clouds with an LLJ stream.

Stationary fronts over the Gulf States

This next regime that can cause stratus cloud formation occurs most often in the fall and winter months. A stationary front orients southwest to northeast or west to east across the Gulf States and a low-pressure system forms on the front (fig. 1–14). The development of overrunning moisture and warm air east of the low can cause low stratus clouds and precipitation to form. This will cause the stratus cloud to move slowly north through the Midwest.

Monitoring meteorological satellite imagery and nephanalysis and weather depiction products are effective metwatch techniques in these situations. The geostationary operational environmental satellite (GOES) low cloud (LC) curve is useful for identifying stratus clouds on infrared imagery.

Another metwatch technique is to monitor reporting stations for increasing surface dew points and decreasing dew point depressions over the central portion of the US.

The leading edge of the low-level moisture advection moves at the relative speed of the LLJ stream wind speed. You can also use upper-air soundings to obtain forecast information concerning stratus cloud layers.

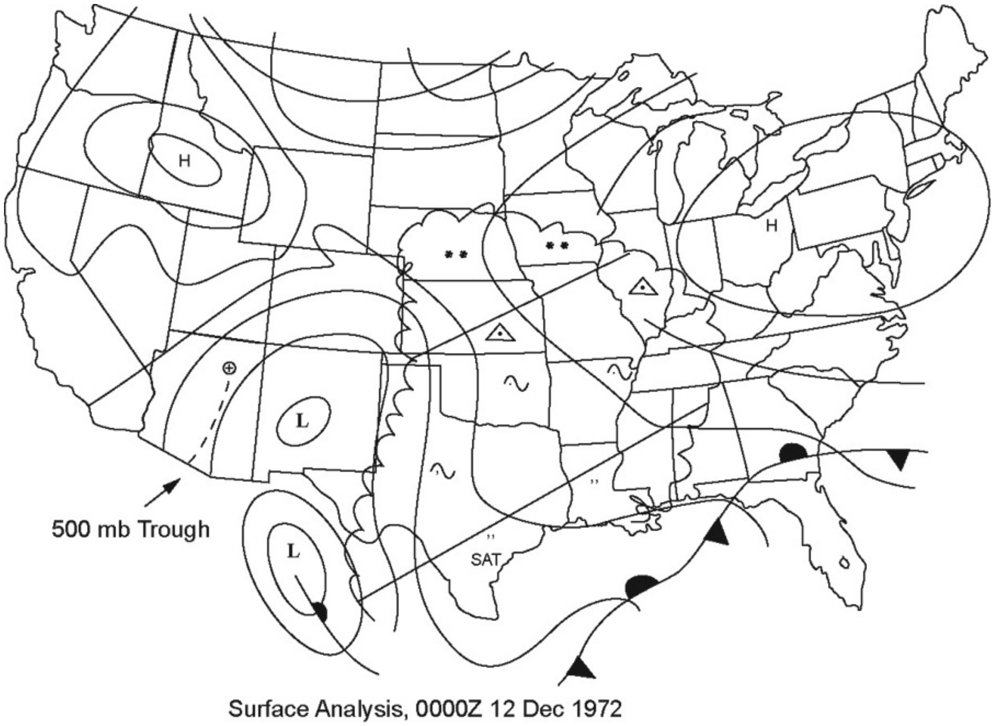


Figure 1–14. Stationary front over the Gulf coast.

207. The relationship between clouds and upper-air features

To provide a more accurate forecast, weather journeymen have derived many rules for associating cloud cover with upper air and other synoptic features. The following 11 rules depict the relationship between clouds and upper-air features.

1. When the contours at 700mb are perpendicular to the surface cold front, the band of weather associated with the front is narrow. There may be a band of clouds associated with a squall line ahead of the inactive (katafront) front.
2. When the contours at 700mb parallel a cold front, the clouds and precipitation extend behind the front as far as the wind remains parallel to the active (anafront) front.
3. Few clouds and very little weather are associated with the front if the 1,000mb to 500mb thickness lines or the 700mb isotherms are nearly perpendicular to the cold front.
4. Clouds and weather are most strongly associated with the front when the 1,000mb to 500mb thickness lines or the 700mb isotherms are parallel to the cold front.
5. Cloudiness and precipitation may be found (given sufficient vertical motion and moisture) under cyclonically curved contours aloft no matter the presence or absence of surface features.
6. In a cold air mass, the instability showers and cumuliform clouds occur *only* where the air is moving in a cyclonically curved path.
7. Warm front cloudiness and precipitation occurs where the 700mb wind flow is across the warm front from the warm side to the cold side and turning cyclonically or moving in a straight line.
8. The 700mb ridge line may be considered as the forward limit of pre-warm front middle and low cloudiness. The 500mb ridge line may be considered the forward limit of the cirrus cloud shield. The sharper the anticyclonic turning of the ridge line, the more accurate this rule.
9. In a warm air mass moving with a component from the south, cloudiness and precipitation is greatest under a current that is turning cyclonically or even moving in a straight line.

10. Clear skies occur when a current of air is moving from the north in a straight line or curving anticyclonically, and in a southward component when it is moving anticyclonically.
11. Cloudiness and precipitation are indicated in areas of positive vorticity advection (PVA) and clear skies under areas of strong negative vorticity advection (NVA).

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

201. States of matter

1. What are the three states of matter?

2. Match the changes of state in column B with the descriptions in column A. Items in column B are used only once.

Column A	*Column B*
____ (1) A solid becomes a liquid.	a. Vaporization.
____ (2) A gas becomes a solid.	b. Condensation.
____ (3) A liquid becomes a gas.	c. Freezing.
____ (4) A solid becomes a gas.	d. Fusion.
____ (5) A liquid becomes a solid.	e. Sublimation.
____ (6) A gas becomes a liquid.	f. Deposition.

For each of the following changes of state, determine whether H_2O would take heat from the atmosphere or release heat to the atmosphere.

3. Vaporization.

4. Condensation.

5. Freezing.

6. Fusion.

7. Sublimation.

8. Deposition.

202. Cloud microphysics

1. What two effects cause differences in saturation vapor pressure (e_s) in terms of CCN growth? Explain these effects assuming that the environment and air surrounding the droplet are saturated.

2. Would you expect greater collision-coalescence of a droplet to occur with stratus clouds or cumulonimbus clouds? Why?

3. A large array of droplet sizes would be indicative of what?

203. Cloud dynamics

1. What is the most effective means of cooling water vapor until it condenses?

2. A cold front has a slope of $\frac{1}{50}$ (1 mile rise over 50 mile run). A second cold front has a slope of $\frac{1}{100}$. Which front has the potential to produce more clouds? (Assume each has the same stability and atmospheric moisture present.) Why?

3. Our station has winds of 270º at 25 knots and winds 20 miles to the east of your station are 270º at 10 knots. Based solely on the wind flow, where might clouds develop and why?

4. Your station has a north/south-oriented mountain range to your east. The winds at all levels are from the west. Satellite imagery shows clouds on the west side of the mountains but no clouds are apparent to the east side of the mountains. What might be a logical explanation for this situation?

204. General tools for forecasting clouds

1. What are four general tools for forecasting clouds?

2. What is a general tool for forecasting clouds that is derived from decades of observational data and is a time-proven method?

3. What general tool for forecasting clouds is generated from model data and provides guidance on relative cloud amount and height range out to 48 hours?

4. What general tool for cloud forecasting is being used when you simply advect clouds from their current location downstream?

5. Which general tool for forecasting clouds should NOT be used without other supporting products?

205. Convective clouds

1. What is the main difference and added consideration when forecasting cumuliform clouds versus stratiform clouds?

2. At what level do clouds form when mechanical lift is the predominant factor?

3. What is the lifting condensation level?

4. What causes thunderstorm tops to penetrate into dry layers at higher levels?

5. What is a method to forecast bases of clouds caused by active surface convection using a surface weather parameter?

206. Stratus clouds

1. With what type of atmosphere are stratus clouds associated?

2. Where do the bases of the stratus clouds form?

3. Where are the tops of the stratus in relation to the RH?

4. Given the same synoptic situation, at what height would stratus clouds form?

5. How does wind affect the formation of stratus clouds?

207. The relationship between clouds and upper-air features

1. In the following statement, what is incorrect? Cyclonically curved contours create clouds and precipitation only if there is sufficient moisture associated with a surface system.

2. In a cold air mass, when do showers and cumuliform clouds occur?

3. As a warm front approaches your station, when can you expect the cirrus cloud to reach you?

1–2. Forecasting Precipitation

Precipitation forecasts are, for the most part, a continuation of cloud forecasts. The considerations that must be made to prepare cloud forecasts are the groundwork for the precipitation forecast. Naturally, the most important requirement for precipitation is moisture in the atmosphere, although this alone does not imply precipitation. Many things influence the production of precipitation and some of these factors are not understood. Some mysteries of precipitation, as well as forecasting techniques, are covered in the following lessons.

Many precipitation forecasts concern themselves with the movement of precipitation areas that are already in existence. Accurate timing of the system movement is necessary for forecasting the time they will cover a specific location or an area of concern. For short periods, simple extrapolation is the best. You find that the extrapolation of precipitation areas, though they are more than three to six hours from your area of concern, is a good way to keep informed of their movement. Extrapolation proves valuable in verifying the periodic positions that you have shown on the prognostic product. Deviations from the predetermined positions may then be used to help you modify future forecast positions.

Doppler weather radar and enhanced infrared meteorological satellite imagery can be used to track existing areas of precipitation and extrapolate the future movement of precipitation. New Doppler algorithms can even predict in which direction convective activity will move and what time it is likely to occur over a particular location. Doppler technology adds the ability to “see” the wind field and gain even further insight into wind shear and severe storm potential. Exercise caution when using the radar; beware of over-interpretation of the products. Many beginning weather journeymen have been guilty of issuing unwarranted warnings of severe weather because they hit the “panic button” when they observed a strong return on a radar product. Always use other tools to supplement radar. Too many erroneous warnings detract from the value of good advisories and warnings.

208. Using temperature and dew-point considerations to predict precipitation from stratiform clouds

In using the dew-point spreads to show possible precipitation, we find that the criteria used to determine overcast skies, $T - Td \leq 2°C$ (discussed earlier) are valid for precipitation considerations. The 850mb and 700mb products may be analyzed for T - Td spreads that can be used to show areas of clouds and possible precipitation within areas of less than 2°C spread. Scan the Skew-T diagram or locally generated vertical cross sections for moisture content (mixing ratio and RH) and stability.

RH indicates the nearness to saturation that helps in determining whether precipitation occurs. The mixing ratio is used in estimating the amount of precipitation likely to fall, should it occur. The stability (vertical motion) helps you in determining whether cloud depth is increasing or decreasing. Neutral or downward vertical motion (negative vertical velocity, Q divergence, NVA) indicates cloud dissipation and, therefore, no precipitation; slight upward vertical motion could indicate a change from drizzle to rain, or perhaps heavier rain; moderate to strong vertical motion would indicate a change to convective or showery precipitation and possible thunderstorms.

Another aid in determining the amount of precipitation is the perceptible water content, or quantitative precipitation forecast (QPF), of the atmosphere. This data may be used subjectively in determining the amount and intensity of precipitation.

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

A study showed that in 87 percent of the cases in which drizzle occurred, it fell from clouds whose temperature at cloud top was warmer than –5°C. The frequency of rain or snow increased markedly when the cloud top temperature was colder than –12°C, especially when the thickness of the cloud was greater than 8,000 feet.

Approaches to forecasting liquid, freezing, or frozen precipitation are varied. Short-range forecasting (0–4 hours) techniques seem to center on continuity and knowledge of local weather characteristics, an “educated extrapolation.” Observed changes in temperature, current weather, radar returns, and other indicators are the tools that you use most often.

Mid-range (4 to 18 hours) forecasting techniques are perhaps the most numerous and the most diverse. As the forecasting value of your chosen short-range techniques lessen, you likely join most other weather journeymen and increase the emphases you put on thermodynamic considerations (thickness, temperature lapse rate, stability, etc.), modeled output statistic (MOS), and storm tracking. Of these, thickness is probably the most used tool because of its direct relationship to the mean temperature through a given layer. By combining thickness values with other locally validated predictors, rules-of-thumb, and your knowledge of local effects, you can ensure your best possible forecast.

Longer-range (beyond 18 hours) forecasting techniques include many of those used in mid-range forecasting, but focus on synoptic reasoning with primary emphasis on the deepening of low-pressure and/or occluding of frontal systems and their associated vertical motions (vorticity, omega, Q-convergence, etc.).

209. Synoptic considerations in predicting precipitation

During winter the location of the snow or rain zone usually depends on relatively small-scale synoptic considerations such as the exact track of the surface disturbance, the wind direction at a coastal station, the position of the warm front, and the orientation of a ridge northeast of a low. However, the larger synoptic features do influence the approximate position of the snow or rain zone. An awareness of this alerts the weather journeyman and is a useful tool in forecasts extending beyond 24 hours.

In the larger sense, the snow or rain zone is tied to the position of the polar front. The polar front location is, in turn, closely related to the position of the belt of strong winds in the middle and upper troposphere. When the wastrels are pushed southward, the storm track is similarly affected and the snow or rain zone may be as far south as the southern US. As the wastrels shift north of their normal position, the storm tracks develop across Canada. With the northward shift, the US has above normal temperatures, and the snow or rain area may exist only along, or north of, the Canadian border.

With a flat, fast westerly flow aloft, the snow or rain zone extends in a narrow west-to-east belt often well ahead of the surface perturbation that undergoes little latitudinal displacement as the perturbation moves across the country. Usually there is little rain, but snow can be found immediately north of the warm front. Most stations where precipitation occurs do not undergo a change from one type of precipitation to another, since there is relatively little advection of warm or cold air with rapid zonal motion.

The snow or rain patterns just discussed have been associated with an active low of the classical type. The rate of precipitation accumulation with this type may be rapid and the transition period from rain, freezing rain, and snow is short, usually a few hours or less. Another snow or rain problem occurs with the quasistationary front in the southern US under a broad west or southwest flow aloft and a weak surface low. The precipitation then tends to become elongated in the direction of the upper-level current. Precipitation rates may be slower, but it stretches over a longer period.

Often, a broad area of sleet and freezing rain exists between belts of snow and rain. Serious icing conditions can develop over an extensive region over several hours or more. This pattern changes as either an upper trough approaches from the west and initiates cyclogenesis on the front, or as the flow aloft veers (turns clockwise with height) and the precipitation dies out.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

208. Using temperature and dew-point considerations to predict precipitation from stratiform clouds

1. For each case below, indicate whether rain should be observed and, if so, indicate whether it should be continuous or intermittent (if possible).

   a. Dew point spread of 4°C at 850 and 700mb.

   b. Dew point spread of 2°C at 850 and 700mb.

   c. Cloud top temperature –10°C in an unstable cloud.

   d. Cloud top temperature –17°C in an unstable cloud.

   e. Cloud top temperature –15°C in a stable cloud.

209. Synoptic considerations in predicting precipitation

1. The snow or rain zone is frequently determined by what parameter scale?

2. Generally, what front is connected with the snow or rain zone?

3. With a flat, fast westerly flow aloft, in which direction does the snow or rain zone spread?

1–3. Visibility Restrictions

The *Glossary of Meteorology,* published by the American Meteorological Society defines visibility as: “The greatest distance in a given direction at which it is just possible to see and identify with the unaided eye, in the daytime, a prominent dark object against the sky at the horizon, and at night, a known, preferably unfocused, moderately intense light source.”

Techniques for forecasting surface visibility have probably been the subject of as many or more local studies than any other single weather problem. Thus, if a study is available at the terminal, this is the place to start the visibility forecast.

The forecasting of visibility is primarily a problem of predicting restrictions to visibility. Any number of parameters, varying from particles that block the visibility to those that diffuse the light, can contribute to restricting or limiting the visibility. Frequently restrictions to visibility are caused by obstructions to visibility which can be divided into dry obstructions (lithometeors) and wet obstructions (hydrometers). Examples of lithometeors are haze, smoke, and blowing dust and sand. Examples of hydrometeors are precipitation, fog, mist, drifting snow, blowing snow, and blowing spray. This section deals with the effects of the various visibility restrictions and the conditions that favor their existence.

210. Formation and dissipation of fog

The most prevalent restriction to visibility is fog. A large variety of causes can produce fog, which presents a major problem to the station weather journeyman. Thus, it is important that the weather journeyman have a solid understanding of the basic mechanisms that produce fog.

Formation

Air can be brought to saturation by two processes—air mass cooling or the addition of water. Sometimes one or both of the processes contribute either simultaneously or one after another.

Air mass cooling

Air mass cooling may be caused by one or more of the following ways:

- Advection of warmer air over a colder surface.
- Adiabatic cooling of air by orographic, frontal, or turbulent lifting.
- Cooling of the earth’s surface by outgoing nocturnal radiation, with resulting cooling of the lower layer of air.
- If two nearly saturated air masses, one cold and one warm, are mixed, it is possible for the resulting air mass to be saturated. This process can theoretically produce fog in the mixing zone along a front.

Adding moisture

Moisture may be added by one or more of the following processes:

- Evaporation from a wet surface, either land or sea.
- Evaporation from falling precipitation under certain conditions.
- Moisture from combustion of hydrocarbon fuel, such as gasoline.

Dissipation

The fog and stratus cloud can be dissipated by either of two ways—heating the air or removing the moisture.

Heating

Any one of the following occurances will cause an increase in air temperature:

- Advection over a warmer surface.
- Turbulent mixing of the fog layer with adjacent warmer air aloft.
- Contact warming of the ground layer by incoming solar radiation during the day.
- Adiabatic warming of the air by subsidence, down slope motion, or the turbulent transfer of heat downward.

Removing moisture

A decrease in the moisture content can be caused in the lower layers by turbulent transfer of moisture upward, by turbulent mixing of the fog layer with adjacent drier air, or by condensing out water vapor as rain, dew, or frost.

211. Types of fog and forecasting them

Fog may form whenever the air at or near the surface of the earth becomes saturated. This saturation can result from precipitation (cooling and moisture added by evaporation), but is more often due to nocturnal radiation or air mass advection, or even more likely, a combination of the two. There are many types of fog, which are named after their formation process (what brought the air to saturation) or location, we’ll discuss eight of them: radiation, continental high inversion, sea, combined-process, steam, upslope, frontal, and ice.

Radiation fog

This is the most common type of fog encountered in the Northern Hemisphere. Radiation fog occurs when the temperature of stagnant air in contact with the ground is cooled to its dew point by nocturnal radiation. This type fog may be classified as *ground fog* or *fog* depending on its observed or forecast depth. Radiational cooling is the reason this fog forms. Classification as ground fog or fog is simply an observed condition based on whether it is less than, equal to, or greater than 20 feet in depth. Radiational cooling is, of course, a factor in all fogs that form over land at night. In ground fog, radiational cooling is the most important factor in causing the fog.

The layer of air next to the ground is cooled by conduction and spreads upward by a very slight turbulent mixing. This cooling effect usually reaches up only a few feet in one night under calm and ordinary terrain conditions.

The degree to which an air mass cools in one night by nocturnal radiation depends on the initial temperature and on the rate and duration of net long-wave radiative transfer. This, in turn, depends on the moisture content and the amount of clouds.

Short days, like those that occur in late fall, winter, and early spring, also help to hold down the maximum daytime temperature. However, under a low subsidence inversion in fall, the maximum may be high without reducing the probability of radiation fog. The temperature fall caused by radiational cooling at night depends largely on the clearness of the skies. Just as a cloud deck retards the radiational cooling of the earth, so is moist air more restrictive to this process of decreasing the radiational cooling process than dry air. Thus keeping temperatures warmer overnight and lessening the chance of radiational fog.

It is easy to see that the moisture content of the air over a station, in addition to the other conditions that are considered prerequisites, plays an important part in the formation of fog. The long nights of late fall, winter, and early spring are most conducive to the formation of ground fog.

One study of radiation fog in Europe suggests during the fall and winter months that if radiation fog persists and doesn’t burn off by a certain time that it will persist all day and into the following night (fig. 1–15).

Month	Time
September	09Z
October	10Z
November	11Z
December	12Z
January	12Z
February	01Z

Figure 1–15. Critical dissipation times for fog in Europe.

There are two factors that, although they do not in themselves cause cooling of the air, are important in the formation of radiation fog. The first is cold air drainage. In irregular terrain, the cold air tends to “seek its own level” and pool in the lowest depression to which the fog may be confined as a result. This is the same mechanism that causes the drainage winds that we’ll examine later. The second factor of importance is that, although the ground becomes very cold during the night, only the air in the lowest three or four feet is cooled by the slight mixing in calm wind conditions.

When no wind is present, condensation occurs first at the earth’s surface as dew, and then either no fog at all or a very shallow layer of ground fog forms. A light wind (2 - 7 knots) persisting well into the night increases, through turbulent mixing, the thickness of the cold layer of air (and decreases the rate of air temperature fall at the surface). This, in turn, decreases the likelihood of dew and may ultimately allow the fog formation to spread through a much deeper layer. A still stronger breeze (8 - 12 knots) may produce enough mixing so that none of the air at the surface reaches saturation. Instead, a stratus deck may form aloft. The presence of such a strong wind is usually associated with a significant advective change, which by definition removes any resulting fog and stratus clouds from the pure radiation class.

The water vapor content of the air is indicated for short-period forecasting by the dew-point temperature. When the actual temperature falls to the dew-point temperature during the night, it may be assumed that saturation occurs. Thus, in making a fog forecast, knowledge of the dew point and the minimum expected nighttime temperature would be valuable information. The dew point is not necessarily constant during the night. It increases if moist air is advected into an area, and decreases if drier air is advected into the area. Even when no advection is present, the dew point usually falls a little at night because of the condensing out of some water vapor as dew or hoar frost. The first formation of dew and the first wisps of ground fog are often not revealed by the temperature readings, for in calm air and over level ground, temperatures and dew points are apt to be considerably lower at the surface than at the observation point (i.e., very steep vertical gradients may exist in the lowest few feet).

The supply of moisture in the lower layer of air can be increased by evaporation from a wet surface. An air mass that is too dry for saturation from radiational cooling may be altered sufficiently by its passage over a lake or river during the day. It then causes radiation fog to occur during the night. If the ground has been saturated by recent rains, evaporation frequently raises the dew point to such a value that heavy ground fog readily forms (this is a common cause of ground fog in the central US). The dew point can also be raised sufficiently by evaporation from falling rain so that subsequent radiation causes fog formations. This usually happens in post-frontal situations.

There is one synoptic situation that is most conducive to the formation of fog. It is when a stable air mass is moist in the lower layers but dry in the upper layers. This type air mass is under a cloud cover during the day but under clear skies at night. Also, when winds are light, nights are long, and the surface is wet it usually indicates a stationary, subsiding, high-pressure area conducive for fog development. In late fall and winter in humid climates of the middle latitudes, the ground is nearly saturated with water, in which only a small amount of rainfall is needed to have standing pools of water. Therefore, the long nights and short days during this season make it most favorable for the formation of radiation fog.

Continental high-inversion fogs

This type of radiation fog is associated with an inversion base from a few hundred to a few thousand feet above the earth. In hilly or mountainous areas, a layer may appear as a stratus deck over the valleys and as surface fog on the slopes of the surrounding hills and mountains. High-inversion fog occurs in the winter in moist air underlying a subsiding anticyclone. The stratus deck forms initially at the base of the subsidence inversion and builds downward into the air beneath. Since the subsiding air above the inversion is relatively clear and dry, nocturnal radiation from the top of the cloud deck results in an intensification of the inversion and a thickening of the stratus layer.

A persistent type of continental high-inversion fog occurs in valleys with access to maritime polar air. The moist maritime air may become trapped in the valleys beneath a subsiding stagnant high-pressure cell, often for periods of two weeks or more. Because of the nocturnal radiation, the trapped air becomes colder. Normally, the stratus cloud forms for only a few hours the first night and dissipates with heating the following day. Each succeeding night the cloud deck thickens, and the next day the stratus cloud persists longer. Eventually, if there is no frontal passage to clear out the stagnant air mass, the stratus cloud builds down to the ground and persists until a front of sufficient strength to change the air mass passes through.

The presence of a stationary high-pressure system over an area subject to continental high-inversion of fog is the first indication of possible formation. A careful study of the synoptic surface and the 700mb products shows the tendency of the high to become stationary and the appearance of the first signs of stratus clouds or fog. The strength of the subsidence inversion and moisture content of the lower and upper-air masses can be determined from the appropriate RAOB. A well-established stagnant high-pressure system, with a strong subsidence inversion, separating a very humid air mass below from a dry air mass above, is the situation most likely to produce a persistent high-inversion fog. After such a system becomes established, it is important to watch for signs of a weakening or moving of the high-pressure system and the approach of a surface front.

Sea fog

Sea fog is an advection fog that forms in warm moist air that is cooled to saturation as it moves across cold water. The cold water occurs as a well-defined current (such as the Labrador current) or as the result of a gradual latitudinal cooling from south to north. The dew point and temperature undergo a continuous change as the air moves across increasingly colder water. The surface air temperature falls steadily and tends to approach the water temperature. The dew point also tends to approach the water temperature but at a slower rate. Thus, if the dew point is initially higher than the coldest water temperature to be crossed and if the cooling process continues long enough, the temperature of the air falls to the dew point, thus causing fog.

If, on the other hand, the initial dew point is less than the coldest water temperature, the formation of fog is unlikely. Generally, in northward moving air, or in air that has previously traversed a warm ocean current, the dew point is initially higher than the cold water temperature in the north, and provided there has been sufficient translation of the air mass, fog occurs.

The following table lists different guidelines for forecasting sea fog during the cooler months over the Gulf of Mexico.

Type of Sea Fog	Ceilings (Hundreds of ft)	Visibility (Miles)	Occurrence	Frequency (Percent)
Warm Advection *(cooling)*	< 5 5–10 > 10	< 2 2 < 6 ≥ 6	Occasional Frequent Frequent	50
Cold Advection *(evaporation, steam)*	< 5 5–10 > 10	< 2 2–3 > 3	Occasional Frequent Occasional	25
Frontal *(along and 50–70 NM north of* *warm or stationary front)*	< 5 5–10 > 10	< 2 2–4 > 4 ≤ 6	Frequent Occasional Occasional	20
Radiational *(light wind—clear skies)*	≤ 2 > 2–5	≤ 1/2 > 1/2 ≤ 2	Frequent Occasional	5

Forecasting the dissipation of sea fog over cold water depends on a change in air mass, or a change in the wind direction that carries the fog over a warmer surface. A movement of sea fog onshore to warmer land leads to rapid dissipation. With heating from below, the fog first lifts, forming a stratus deck. With further heating, this overcast breaks into a stratocumulus layer and eventually into convective clouds or evaporates entirely. An increase in the wind speed can raise a surface fog into a stratus deck, at least temporarily. Over very cold water, dense sea fog may persist even with high winds.

Combined-process fog

Combined-process fog is produced by a combination of advection, radiation, and/or the natural (precipitation) or manmade (hydrocarbon fuels) addition of water vapor. Warm air is often advected inland from over a moisture source. When the air is over land, it experiences radiational cooling and the formation of fog. This fog is often seen in the Gulf States and on the east coast of the United States because of the southerly wind flow from the Gulf of Mexico and southeasterly flow from the Atlantic Ocean.

Steam fog

Steam fog, or Arctic Sea smoke, is often regarded more as a curiosity than as a serious impairment to visibility. However, under certain circumstances, it may become quite dense and persistent. It then presents a serious hazard to both shipping and aircraft operations. Steam fog develops when very cold air blows over open water and induces a rapid transfer of heat and moisture from water to surface air. The heating from below produces an unstable lapse rate, and the associated small-scale convective eddies carry the warm moist surface strata aloft, where they mix with the cold dry air. The first result of this mixing is to produce condensation as fog, but continued mixing of the dry air causes the fog to dissolve.

As a rule, steam fog appears as wisps of smoke emanating from the sea surface. However, if a strong inversion is present, the upward mixing is confined to a relatively shallow layer within which the fog collects and assumes a more uniform density. Under these conditions, the visibility may quite often be reduced to 300 meters or less.

Upslope fog

Upslope fog occurs when air is lifted over sloping terrain and may be regarded as either a cloud or fog, depending on the point of reference of the observer. When the air flows up the slope, it is cooled adiabatically and condensation occurs if saturation is reached. Since saturation is normally reached at higher altitudes, upslope fog generally forms at the higher elevations and builds downward. It can be sustained in higher wind speeds than other types of fog because of the increased adiabatic cooling accompanying the increased speed.

Frontal fog

Frontal fog is usually divided into two primary types, pre-frontal and post-frontal. Pre-frontal, or warm frontal, fog is the predominant type. It occurs over widespread areas of warm fronts and is caused by the evaporation of warm precipitation into colder air.

Post-frontal, or cold frontal, fog is less likely to occur than warm frontal fog. The primary cause of this fog is also evaporation of water into the colder air. However, it comes either from rain falling into the colder air or from the surface that has pools of water standing from previous rainfall. Dissipation of fog, in both cases, occurs after frontal passage and increased heating or mixing from the winds.

Ice fog

Ice fog is largely a manmade addition to the Arctic scene, and perhaps has been the object of more intensive study than the more familiar water fogs. Ice fog of significant density is found only near human habitation, where large quantities of water vapor are added to the air through the burning of hydrocarbon fuels. Steam vents, motor vehicles, and jet exhausts are the major sources of moisture that can produce a sharp reduction in visibility in restricted areas.

Meteorological conditions that favor the formation of ice fog are low temperatures (usually below –20°F but more likely below –30°F) and a low-level inversion that traps and concentrates the moisture in a shallow layer.

Graphical method for forecasting fog

The graphical method is valid for short (0 to 4 hours) periods and all times of the day. The previous 3- and 6-hour temperature and dew point and graph paper are the only tools needed. Use the temperature scale in effect for the period being plotted on the graph. For example, Figure 1–16 shows a Y-axis scale gradation from +4 to –2 because the 6-hour temperature was 4°C and dew point was –2°C respectively; then use a X-axis gradation from N6 to 6. Apply the following when using the graph:

1. Plot the current temperature (T) and dew point (T_d) on the vertical line labeled “N.”
2. Plot the 3-hour old temperature and dew point on the vertical line labeled “N–3”, the 6-hour old “N–6” line.
3. Connect the plotted temperature values with a line, extending the line to the right edge of the graph. Similarly, connect the plotted dew-point values and extend this line to the edge of the graph.
4. If the lines do not intersect, stop. Do not forecast fog for the following 4 hours. If the lines do intersect, from the point of intersection you can find the forecast time by proceeding vertically downward to the time scale. Add “N” to the forecast time to arrive at an onset time for the fog. For example, in figure 1–16, if the current time was 0900 UTC, then forecast fog at 1200 UTC (0900 + 3 hours).

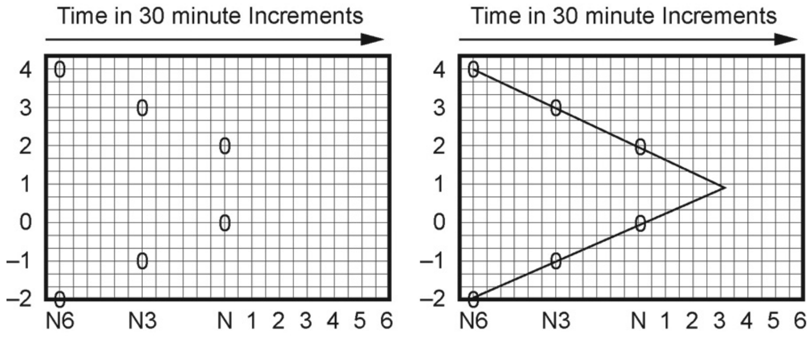


Figure 1–16. Graphical method for forecasting fog.

Meteograms and weather research and forecast graphic products used for forecasting fog

The AFWA-provided WRF-based meteogram is another tool that can be used to forecast visibility in an indirect way. The meteogram provides a forecast for temperature and dew point for a specified time period based on the resolution you select. You can generate an interactive meteogram for any location you chose. The meteogram shows a graphical representation of temperature and dew point and you can identify when the temperature dew point spread decreases (increasing moisture) or increases (decreasing moisture). The graphical trend of the temperature and dew point forecasts can provide a valuable tool indicating whether fog may develop, or if present, may dissipate. The meteogram also provides information on wind speed and cloud cover which affects the development of radiational fog.

Another great tool used for fog forecasting is the WRF-based surface visibility graphic products as shown in figure 1–17. Forecast surface visibilities are displayed in color for 0 to ≥8 miles.

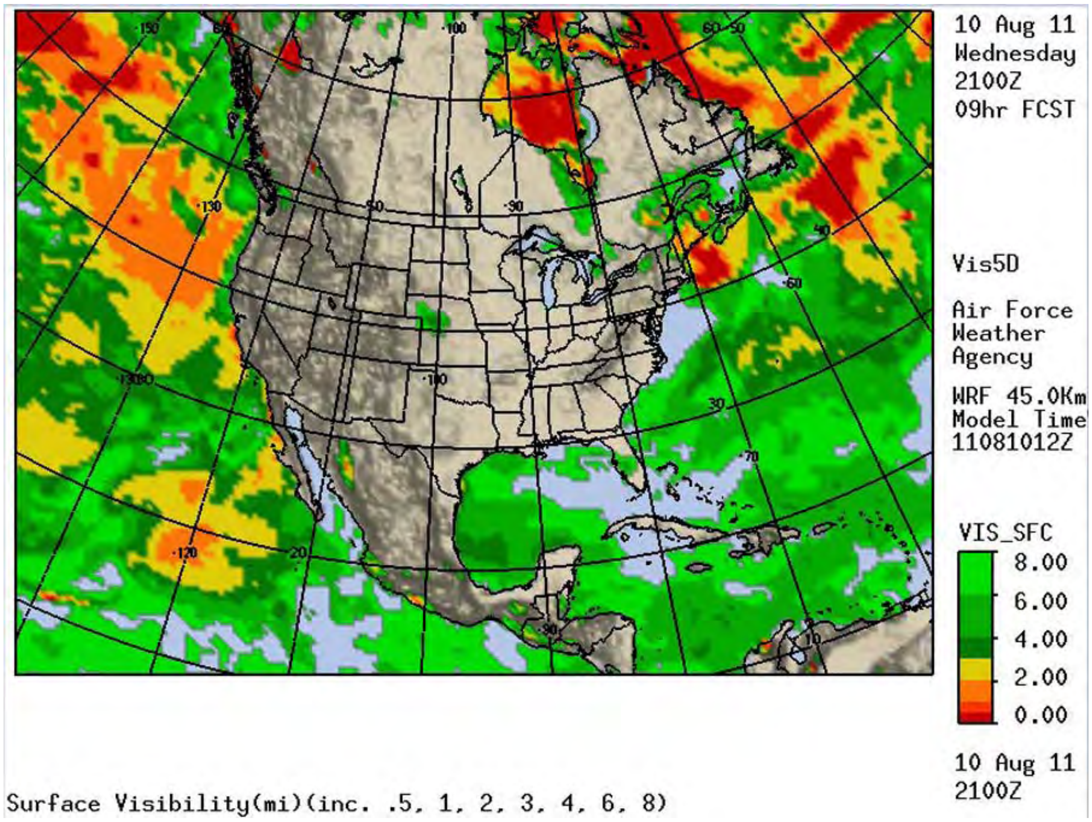


Figure 1–17. WRF-based visibility forecast graphic.

As mentioned earlier, fog is a hydrometeor. In the next lesson, you'll explore restrictions to visibility caused by lithometeors.

212. Visibility restrictions caused by haze, smoke, windblown particles, and precipitation

Other restrictions to visibility may be caused by haze, smoke, windblown particles, and precipitation. Here, we look at each of these restrictions.

Haze

Haze is an ever-increasing problem for the weather journeyman. It is a direct result of industrialization, or an accumulation of sea salt, or both. Since haze is an accumulation of very small particles in the atmosphere, it does not result in the actual blocking of light, but causes the blue light rays to be diffused or refracted.

The phenomenon (light being diffused or refracted) also results in different visual ranges within a uniformly dense layer of haze, depending on whether the observer is looking into the sun or away from it. This should be considered when you brief a pilot on visibility restrictions caused by haze. Little has been written on forecasting visibility restrictions in haze, but the weather journeyman can make a reasonable forecast of haze by keeping in mind a few facts:

1. Haze, like fog, requires a stable atmosphere. Large, summertime, stagnant, high-pressure systems with their extensive areas of subsidence, frequently produce broad areas of haze.
2. Haze develops in layers that can restrict visibility to less than one mile, but usually not below three to six miles.
3. Industrial areas and coastal areas are, as a rule, most conducive to haze formation.

The height of the top and bases of haze layers aloft are important because they mark the bases of inversions. Pilot reports can provide you with the heights of bases and tops.

Forecast haze to dissipate when the atmosphere is forecast to become unstable. This can occur with heating, advection, or turbulent mixing (high-speed surface winds). Another consideration in making visibility forecasts is the pollution of the atmosphere. If the air in question has recently moved into your area, the probability of haze is less than it would be if the air has been over a local industrial area for an appreciable time.

Smoke

Smoke is more localized than any other restriction. Smoke can become an obstruction to visibility, as either a general area or a layer of smoke over the station, or it can be limited to a thin band passing over part of the air base. In the general area or layer, the layer of smoke forms because of extreme stability and calm or very light winds. If a source of smoke is located near a station and the station is in a shallow basin, smoke at times may become a critical problem. Consider the local terrain and the source of smoke. The other situation, the band of smoke, can occur when the winds are strong, the air is reasonably stable, and the wind direction is such that the smoke from the source is carried across the air base.

Windblown restrictions

Blowing dust, sand, or snow can reduce visibility to zero if the conditions are right. With dust or sand, the problem becomes more involved than the restriction of visibility. The hazard to machinery is quite critical. Blowing sand and dust are strictly local problems, depending on the type of soil in the area and the critical wind speed for lifting the soil. Forecasting blowing dust is different for every station that has this problem. Blowing sand and dust have significant effects on military operations as seen in the Gulf War in 1991 and in continuing operations in southwest Asia. Local forecast studies have data for determining what the critical wind conditions are. From this data, you can formulate your forecasts.

Blowing snow, on the other hand, is more widespread than blowing dust. The occurrence of blowing snow depends on the state of the snow surface and the wind speed. If there is new fallen snow and it is dry and fluffy, a forecast of strong winds (>15 knots) should be accompanied by a forecast of reduced visibility due to blowing snow. The condition of the snow and the strength of the wind determine the degree to which the visibility is reduced. No fixed criterion exists for determining visibility.

Precipitation

Visibility restrictions caused by rain and rain showers can be very difficult to forecast. No strict rules have been developed for relating the intensity of rain with the expected visibility limits. On the other hand, once you have forecast the occurrence and the intensity of drizzle, snow grains, snow pellets, or snow (discussed later), you only need to forecast the visibility so that it agrees with the intensity of the precipitation. The visibility criteria for precipitation vary by precipitation type and are listed in AFMAN 15–111. Precipitatation intensity based on visibility can only be determined with snow and drizzle based precipitation. The table below lists the precipitation intensity expected with the following visibility thresholds.

Intensity	Visibility Limits
Light	> 1/2 mile
Moderate	< 1/4 mile
Heavy	≤ 1/4 mile

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

210. Formation and dissipation of fog

1. Indicate using a “C” those processes that create fog or use a “D” for those that dissipate fog.

    a. Addition of moisture.

    b. Removal of moisture.

    c. Cooling of air.

    d. Warming of air.

2. Indicate, using a “C,” the parameters that may create fog and use a “D” to indicate those that may dissipate fog. Identify how it creates or dissipates fog by listing the process involved from question 1.

    a. Evaporation from falling precipitation.

    b. Combustion of hydrocarbon fuels, such as gasoline.

c. Contact warming of the air near the ground.

d. Orographic, frontal, or turbulent lifting of the air.

e. Advection over warmer surface.

f. Evaporation from a wet surface.

g. Advection over colder surface.

h. Subsidence, down slope motion, or turbulent transfer downward of the air.

i. Nocturnal radiation.

j. Turbulent mixing with warmer air aloft.

211. Types of fog forecasting them

1. What is the most common type of fog in the Northern Hemisphere?

2. What factors determine the rate at which an air mass cools due to nocturnal radiation?

3. Which type of fog is advection fog that forms in warm moist air that is cooled to saturation as it moves across cold water?

4. What type of fog forms in higher elevations and builds downward?

5. What type of fog is usually limited to areas near human habitation?

212. Visibility restrictions caused by haze, smoke, windblown particles, and precipitation

Identify the type of restriction in each situation below.

1. This restriction develops in layers in a stable atmosphere usually of fairly large depth. The visibility is usually between three and six miles. To predict this restriction, you must locate a source of pollution and predict the stagnation of an air mass in your area.

2. This restriction is a localized phenomenon that depends on the wind direction to determine areas of restriction. There must be a strong inversion and light winds. To predict its occurrence, you must locate the source and know the wind direction.

3. This restriction varies with the wind speed and type of restriction. The visibility may be restricted completely or only partially, depending on local conditions.

4. What is the intensity of snow showers occurring at you station with a prevailing visibility of five-eighths of a mile?

1–4. Wind Forecast

Surface winds, because of significant diurnal temperature changes in the lower levels of the atmosphere and the resultant formation or non-formation of a low-level temperature inversion, undergo a definite and readily forecastable diurnal trend. Maximum and minimum surface wind speeds normally occur at the time of maximum and minimum surface temperatures, respectively.

Winds result from a difference in air density. This difference can be easily measured at the surface in terms of the change in temperature and/or pressure with distance. The winds resulting from these differences vary considerably from one geographical location to another, therefore, making it near impossible to accurately forecast the winds based upon temperature and pressure difference alone. However, when you combine other tools, such as low-level lapse rate, gradient wind speed, stability, MOS data, local rules-of-thumb, and your knowledge of the effects of local topography, your ability to accurately forecast the wind increases.

213. General tools for forecasting winds

It’s a challenge being a new weather journeyman at an OWS forecasting winds for many varied locations. However, rest assured that there are plenty of tools to help you in this endeavor. From climatology to topographical maps to atmospheric model output, these are tools or job aids you can use to forecast winds.

Climatology

Climatology is a useful tool in forecasting winds. It provides historic averages of wind speed and direction over a period of years. Consult it first to identify prevailing winds for the location and time of interest. These prevailing or climatological winds are meso- and micro-scale local phenomena such as land and sea breezes and thermal lows. Variations from climatological winds are often the result of migratory systems such as lows, highs, and fronts. Climatological winds can be retrieved from several sources, including station climatic summaries and Surface Observation Climatic Summaries (SOCS).

Topography

Topography can have an important effect on both the direction and speed of winds. Frictional effects due to rough terrain can slow wind speeds and change their direction. Mountains upstream may delay or block winds or trigger strong down slope winds. Use a detailed topographic map from the National Imagery and Mapping Agency to identify the forecast location and the terrain surrounding it. Note any unique features, such as bodies of water, hills, and valleys. There should be a reference file built for every location that forecasts are issued for routinely. The reference file should identify any topographical influence that terrain has on the weather elements at that location.

Trends

If the synoptic situation is stagnant and doesn't change, then use persistence for short-term forecasting. Look back on previous days' observations to notice any diurnal similarities. This method works well in tropical locations, where weather conditions remain much the same every day. In tropical locations diurnal changes usually dominate.

Geostrophic winds

A geostrophic wind is the wind located just above the friction layer. Forecasters can estimate what surface winds will be by knowing the geostrophic wind speed and correcting it for friction. Using geostrophic winds is a short-range, up to 2-hours, forecasting method. Because geostrophic winds are sensitive to changes in the pressure field it makes them unsuitable for long-term forecasting. They are also unstable in areas where the isobars are strongly curved such as near deep low-pressure centers.

To use geostrophic winds, locate a speed value for a forecast location on a geostrophic wind chart from NCEP, a representative upper-air sounding, or from a WSR–88D generated vertical azimuth display (VAD) wind profile (fig. 1–18). Calculate about 2/3 of the geostrophic wind speed value during time of maximum daytime heating. Remember the geostrophic wind direction has no frictional effects. Frictional effects have to be added when forecasting actual surface wind direction and speed. The surface wind may not be representative if the geostrophic wind is less than 15 knots.

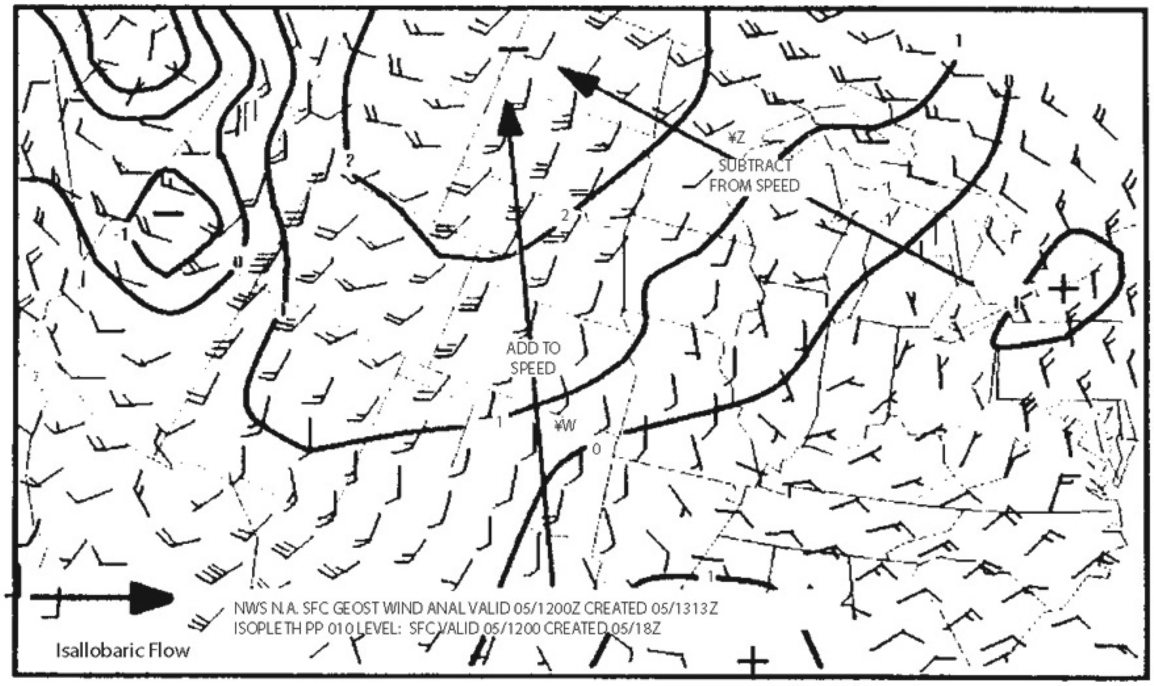


Figure 1–18. Geostrophic wind product.

The mean wind direction in the Northern Hemisphere deviates from the geostrophic direction by minus 10° over ocean areas and up to minus 50° over rugged terrain. An example of this would be if your station is located in rugged terrain, the geostrophic wind is 240° at 30 knots, your surface wind would be approximately 190° at 15 knots. Each station has a locally determined directional deviation based on the topography of the surrounding area.

There are limitations to this method, however. Do not use geostrophic winds to forecast surface winds in vicinity of convective activity. Use geostrophic winds to forecast surface winds after a frontal passage, but not to forecast wind shifts with frontal passage.

When a shallow inversion is present, the surface wind may not be representative of the geostrophic wind. The geostrophic winds may overestimate the actual surface wind when a low-pressure system is within 200 miles of the area.

Numerical output products

Numerical output from atmospheric models such as Model Output Statistics guidance products, are objective tools used to forecast wind speed, direction, and other weather elements. MOS values are produced only for specific locations. If your location is not one of those for which MOS data is generated for then use guidance from a nearby location if it is representative of the location. Once again, remember to initialize the model that the numerical output is produced from.

Use these bulletins to note trends, such as increasing or decreasing wind speeds and/or directional changes, rather than an exact forecast. The model outputs do not handle tropical cyclone winds well.

In the Pacific theater of operations the numerical output from the Navy’s Operational Global Atmospheric Prediction System (NOGAPS) model provides wind speeds and direction at an approximate gradient level of 2,000 feet out to 120 hours. As a general rule, surface winds are 60 to 100 percent of these winds depending on the terrain and synoptic situation.

Meteograms and graphic products

The AFWA produced meteograms based on the WRF model are another tool that can be used to forecast winds. The meteograms display wind direction and speed as wind barbs from the surface to 55,020 feet for every three hours for a specified time period based on resolution. The graphical presentation of the meteorological information is much easier to interpret than standard numerical output bulletins.

The WRF output is incorporated into graphic products to use as guidance to forecast the wind. Figure 1–19 is an example of such a product.

Doppler weather radar

The Weather Surveillance Radar 88 Doppler provides real-time wind data through its velocity products. The Velocity Azimuth Display Wind Profile (VWP), which is covered in detail in a different CDC volume, provides wind direction and speed from surface up to 40,000 feet every five minutes. The product is capable of providing wind data up to 70,000 feet and is similar to receiving the wind data from an upper-air sounding every 5 minutes.

Upper-air soundings

Using the Skew-T wind data from a nearby location is another option or an additional resource. Remember that Skew-T data is usually available only twice a day and that if the synoptic situation is varying rapidly the wind data may become unrepresentative rapidly. Try to use Skew-T data from an upstream location or the most representative site to your location to get the most accurate wind information.

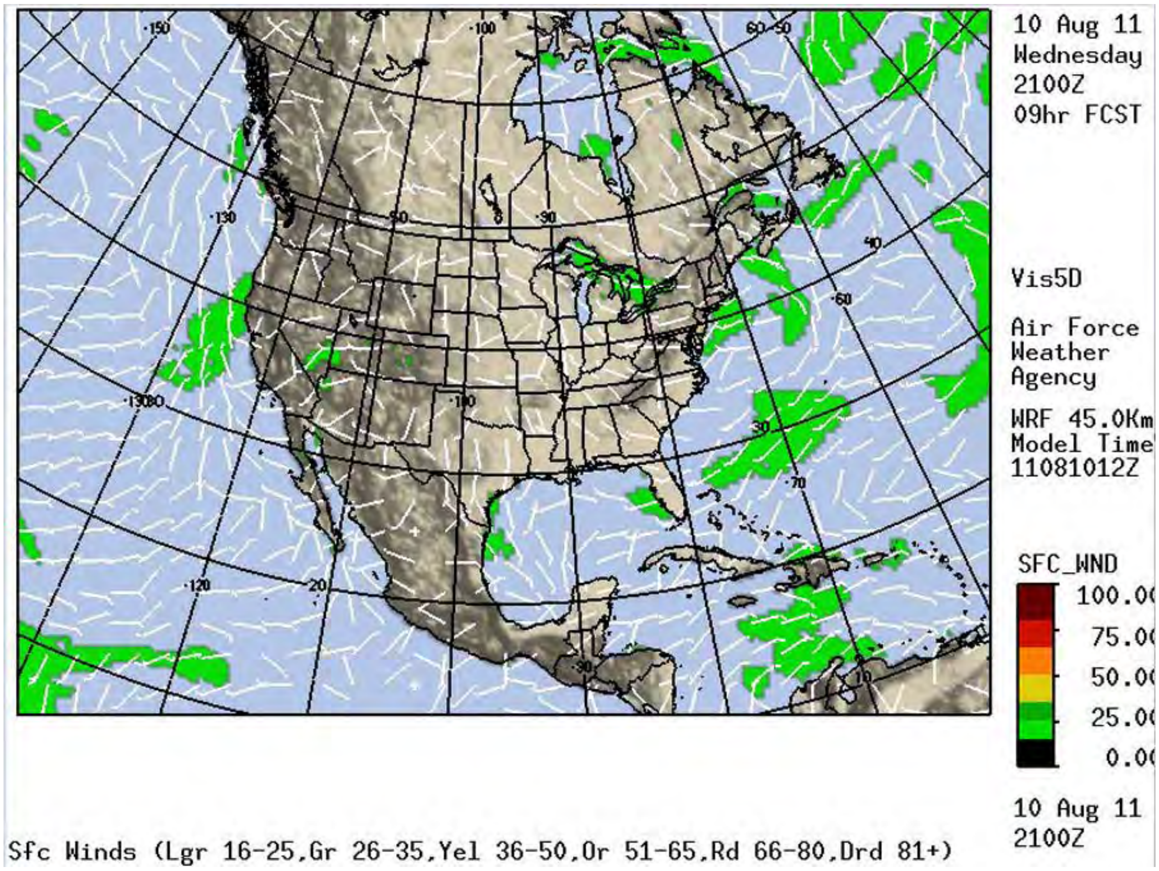


Figure 1–19. WRF-based surface wind product.

Detecting wind speeds from open-cell cumulus clouds

Another method in which to obtain wind data over data sparse areas is by using satellite imagery to analyze the shape of open-cell cumulus clouds. Open-cell clouds are associated with straight-line or cyclonic flow. Since these clouds form in the low-levels of the troposphere, their shape can provide valuable information about the environmental wind field around them. The best types of imagery to use are high 1 to 4km resolutions and visible and infrared images. Keep in mind that the winds at cloud level may not be the same as the surface wind. Using an atlas, topographical, or tactical map, study the terrain of the area covered by the satellite photo. Decide how much frictional effect to apply to the low-level wind indicated by the open-cell cumulus. To estimate approximate wind speeds from open-cell cumulus clouds, determine the shape of the cloud then use the following table or figure 1–20 to determine the wind speed.

Shape W	ind Speed
Doughnut with a hole (Quadrant A)	Less than 10 knots
Elongated doughnut shape with a hole(Quadrant B)	11 to 20 knots
Arc shape (Quadrant C).	21 to 30 knots
Solid and elongated (Quat D)	Greater than 30 s

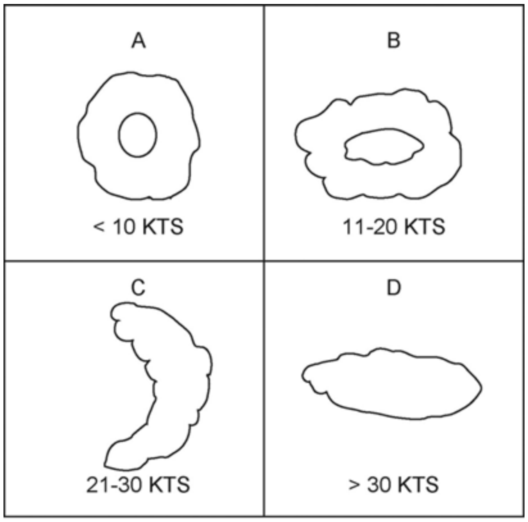


Figure 1–20. Open cell cumulus.

Now that you have general tools for forecasting wind, let’s turn our attention to some of the dynamics associated with forecasting wind.

214. Forecasting surface wind

In this lesson, wind forecasting is presented in relation to two different scenarios:

- Non-frontal surface winds.
- Frontal surface winds.

Forecasting non-frontal surface winds

Within an atmosphere with a flat pressure gradient, the temperature distribution and atmospheric stability may be considered as the major indicators of the winds. With a strong pressure gradient, the winds blow during the night and the day. The strong winds prevent the development of any surface inversion, cause little change between daytime and nighttime wind speeds, and cause the diurnal temperature variation to be much less. When the pressure gradient is weak, the maximum wind speeds occur during maximum heating of the atmosphere, and the minimum wind speeds occur during maximum cooling of the atmosphere. The horizontal temperature gradient does have some influence on the wind speeds, but the vertical gradient is more important.

Forecasting the development of winds during the day under a flat (weak) surface pressure gradient is done by forecasting the time that the surface inversion breaks. If you forecast a relatively stable lower atmospheric layer during the entire period, the winds remain light or become steady at a speed of approximately 40 to 70 percent of the gradient winds that are indicated by the surface pressure gradient, as shown in figure 1–21. The ratio of the wind speed to the gradient wind depends on the surface conditions and should be determined by local forecast studies.

If you forecast the breaking of the surface inversion during the day, your forecast includes indications of considerable insolation. You should forecast the surface wind to attain a speed near the gradient wind speed, and, in areas like the Midwestern US, near the supergradient wind speed for short periods of time, with gusts approaching 80 percent of the 5,000 foot wind. These gusts occur after the surface inversion is completely wiped out, and they last throughout the time of maximum insolation.

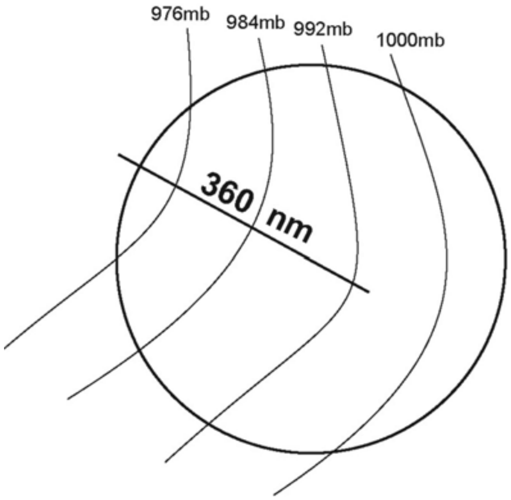


Figure 1–21. Forecasting winds by using pressure gradient.

Forecasting frontal surface winds

Frontal winds are usually forecast by extrapolation of the wind from an upstream station that has the same relative position with respect to the front that you anticipate your station to have at forecast time. This wind, assuming persistence of frontal characteristics, is a close approximation of your station's wind in the future. Since changes in frontal characteristics affect the wind speeds, an account of them must be considered. Deepening or filling of the frontal trough can increase or decrease the wind, and cause changes in moisture content inturn, increasing or decreasing the cloud cover.

Temperature contrast changes resulting from this or other causes alter wind speeds. Normally, there is less of a pure diurnal effect along a front than exists deeper within an air mass because diurnal temperature changes along the front are less pronounced. These things must be considered subjectively, but experience and local studies will help you to weigh these factors effectively in preparing your forecast.

Occasionally, situations arise where surface wind direction and speed are dominated by parameters other than pressure gradient or frontal circulation. This is the case with tertiary circulations.

215. Tertiary circulation patterns

Tertiary circulations are of a third order. In the atmospheric circulation system, tertiary circulations are small-scale wind systems that occur with the general circulation pattern. They are a result of the earth's rough surface and temperature differences between land and water.

Tertiary circulations are frequently called *local winds* and have names that link them to the place where they occur. In spite of their many names, we can lump tertiary circulations into four categories—local-cooling, local-heating, adjacent heating and cooling, and forced circulation winds.

Local-cooling winds

Local-cooling winds are the result of differing specific heat of land surfaces in a generally small place. These are nighttime circulations that develop as one surface radiates heat energy and cools faster than another, such as the mountain breeze shown in figure 1–22. At night, the rocky slopes of the mountain lose their heat more rapidly than the moist, vegetation-covered surfaces in the valley. Air in contact with the mountain cools and becomes denser than the valley air.

This density difference causes a pressure gradient to form with high pressure on the mountains and low pressure in the valley. Winds then blow in the direction of pressure gradient force (PGF), down the slopes into the valley. Wind speeds associated with this type of wind rarely exceed 15 knots, even with a well-developed breeze.

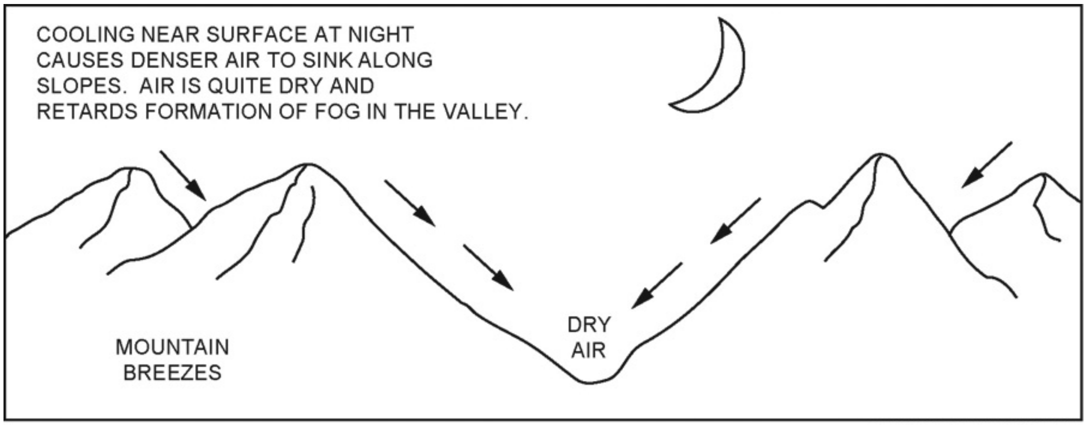


Figure 1–22. Mountain breeze.

Local-heating winds

Local-heating winds are the result of differential heating of two surfaces. Since it is due to local heating, it is a daytime feature dependent on incoming solar radiation. Reverse of the mountain breeze, the valley breeze illustrates this type of wind (fig. 1–23). During a valley breeze occurrence, the sun’s rays strike the mountain slopes before they strike the valley surfaces. Thus, the mountains warm quicker than the valley and lower pressure forms on the mountains. Now the PGF is directed up the slopes and the breeze blows from the valley to the mountain. The valley breeze is usually a more developed wind than the mountain breeze and reaches its maximum intensity in mid-afternoon.

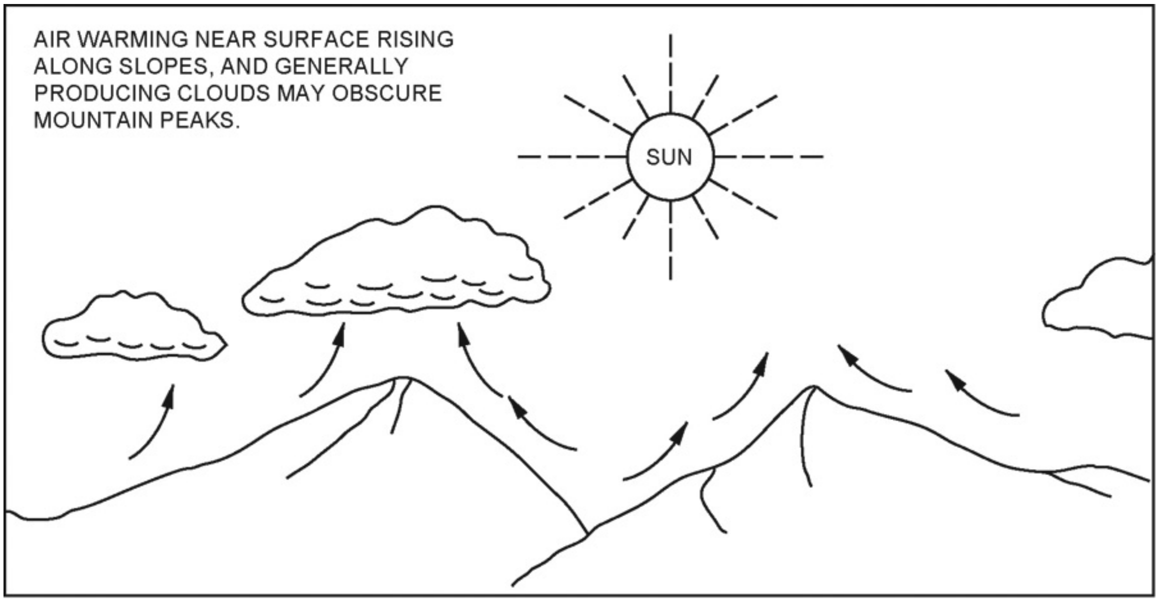


Figure 1–23. Valley breeze.

Adjacent heating and cooling winds

Land and sea breezes illustrate the adjacent heating and cooling winds (fig.1–24). Since land masses absorb and radiate heat three times faster than water masses, land is warmer in the daytime and cooler at night. Differential heating of the land and water causes lower pressure over the warmer land and higher pressure over cooler water. Therefore, PGF is directed toward the land from the water. Thus, on a coast or shoreline, sea breezes (coming from the water) would be expected in the afternoon. At night, the system reverses itself as the land cools more rapidly than the water. Therefore, higher pressure exists over land than the water and the PGF is directed offshore. The best-developed land breeze occurs just before dawn.

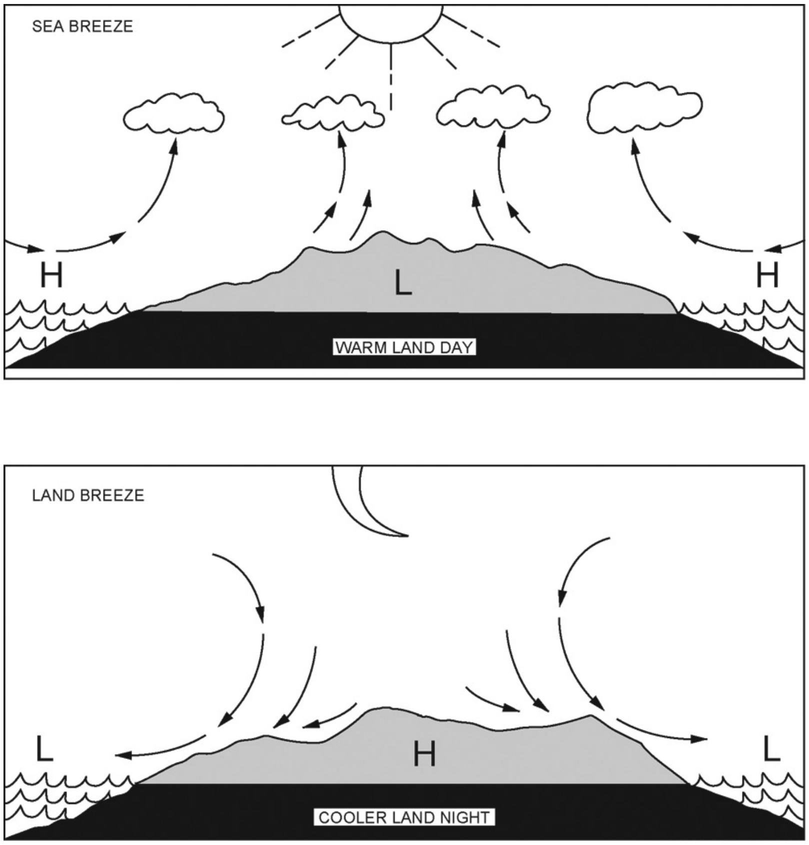


Figure 1–24. Land and sea breeze.

The entire pattern depends on the fact that land and water heat and cool at different rates. The sea breeze reaches maximum velocity between 1400 and 1600 local time and is much stronger than the land breeze. Sea breezes may be felt as far inland as 25 or 30 miles. The sea breeze is also a major factor in afternoon thunderstorm development along coastal regions. Cool air blowing inland creates a boundary (sea-breeze front) between cool, moist marine air and warm inland air. The sea-breeze front is often seen in Florida during the summer. Land and sea breezes are also common along rivers and lake shores.

Forced circulation

Forced-circulation winds are circulations that intensify because of terrain; usually mountain ranges. There are two types of winds in this category—fall and Foehn.

Fall winds

Fall winds (sometimes called *glacier winds*) are cold down-slope winds that remain colder than the air they replace in the lowlands. Fall winds are extremely strong and gusty (60 to 100 knots) and often cause considerable damage. These winds begin in the highlands where air is trapped over cold, snow-covered regions until forced down slope by a migratory pressure system. The winds move down the mountain slope and intensify as they move through the valleys. They are most violent in winter and spring, with single occurrences lasting for several days. Fall winds are frequently called *Mistrals*, the name of the local wind in the lower Rhone valley of southern France. Another name for a fall wind is the Bora that blows from the mountains of Yugoslavia to the Adriatic Sea.

Foehn winds

The Foehn wind, called the *chinook* in North America, is also a down-slope wind, but Foehn winds are warm and dry. The drying results when air is forced up the windward side of a mountain. This rising air cools and loses its moisture due to condensation. It then continues over the crest of the mountain and down the slopes, warming adiabatically as it compresses in the lower elevations. The result is a warm, dry wind that blows downward along the leeward slopes of the mountain. A Foehn wind may raise the temperature as much as 50°F in the low lands on the leeward side of a mountain. The temperature increase may occur in as little as a few minutes. Figure 1–25 shows the sequence of events for a Foehn wind and the clouds that you may observe on the windward side and near the peaks of the mountains. Notice the absence of cloudiness on the leeward side of the mountain. Expect this since the air loses its moisture through condensation as it is forced up the windward side.

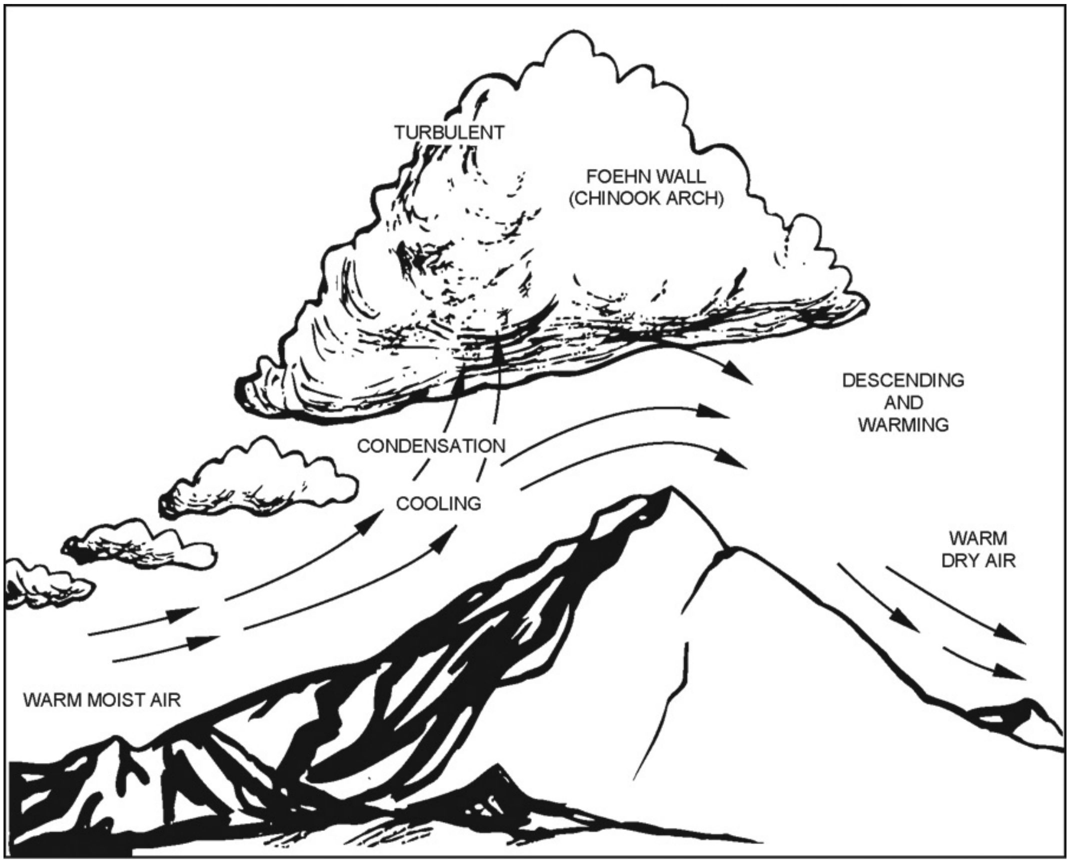


Figure 1–25. Foehn (chinook) wind.

Though considered a local wind, the Foehn wind can be a hazard to aviation. The hazard occurs when the wind direction is perpendicular to the mountain and the wind speed increases rapidly with height. This causes extremely turbulent conditions above the peaks and leeward of the mountain. This condition may be detected by the presence of cap, rotor, and lenticular clouds as depicted in figure 1–26. Also notice in the figure the wind currents associated with mountain-wave conditions.

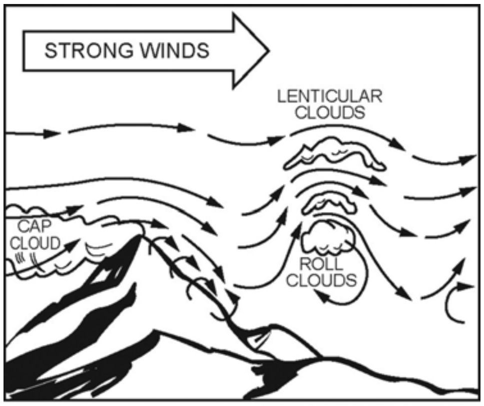


Figure 1–26. Mountain wave.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

213. General tools for forecasting winds

1. Which general tool for forecasting winds can provide information on surface frictional effects?

2. What general tool should you use to forecast winds if the synoptic situation is stagnant and not changing?

3. Which general tool for forecasting winds are graphical in nature, based on the WRF model and provide wind forecasts for three hour intervals?

4. What would the approximate low-level wind speed be if open-cell cumulus in the shape of an elongated doughnut was present?

214. Forecasting surface wind

For each set of data below, predict the speed of the surface wind.

1. Time 2300L, strong inversion, gradient wind of 30 knots, and 5,000-foot wind of 50 knots.

2. Time 1100L, weak inversion, gradient wind of 30 knots, 5,000 foot wind of 60 knots. Temperature needed to break the inversion of 72°, current temperature is 71°. What will the winds be at 1200 LST?

3. Time 2300L, no inversion, gradient wind of 30 knots, and a 5,000 foot wind of 50 knots.

4. Time 1100L, no inversion, gradient wind of 30 knots, and 5,000 foot wind of 50 knots.

5. What is the usual way to forecast frontal winds?

215. Tertiary circulation patterns

1. Name the four categories of tertiary circulations.

2. What is the nocturnal wind that flows into valleys and to what category of tertiary circulation does it belong?

3. Identify the wind caused by warming the air near the mountain slopes and to what category of tertiary circulation does it belong?

4. In which tertiary circulation are land and sea breezes?

5. The Mistral and Bora belong to which category of wind?

6. What category of wind is the fall wind?

7. Cap, rotor, and lenticular clouds are characteristic of which forced-circulation wind?

1–5. Temperature Forecast

The temperature forecast is the forecast that you are requested to make the most often. There is a variety of temperature forecasts made every day. The temperature undoubtedly has a greater influence on other weather elements than any single element. The temperature affects the mission of the Department of Defense (DOD) from weapon system performance to trafficability for ground troops. The forecasts most commonly made are the maximum, minimum, and runway temperatures. Others that are made often are temperatures with frontal passages, critical temperatures for thunderstorm forecasts, "heat waves" or "cold waves," and subfreezing temperatures.

Many things, among which are insolation, radiation, mixing, advection, conduction, and the adiabatic process, cause changes in temperature. There are rules to consider when making a temperature forecast, such as dry air heats and cools faster than moist air or the lack of cloud cover increases radiational cooling. Techniques you use for the accurate forecasting of surface temperatures make use of many tools. Their combination and value vary, depending on season and situation. They range from objective MOS data output to conditional climatology curves to your own subjective adjustments based on insolation and radiation information that you've gleaned from satellite imagery of advecting clouds and water vapor.

216. Measuring temperature

Temperature is another important measure in the study of physics. Although most definitions of temperature are usable, some are much better than others. For example, we may define temperature as a measure of how hot or cold an object is. This is definitely a true statement, but it is somewhat vague for use in physics. A better definition is the one used by the scientific community: *temperature is the measure of molecular activity of a body or object*.

We can illustrate molecular activity by comparing two like bodies at different temperatures; one relatively hot compared to the other (fig. 1–27). Notice that in the colder body, object A, the molecules represented by shivering figures are somewhat motionless. The molecules of object B, represented by jumping figures, possess more thermal energy. As molecules move about, they collide with one another. These collisions produce sensible energy—heat. Temperature is a measure of this energy; the more collisions that occur, the higher is the object's temperature.

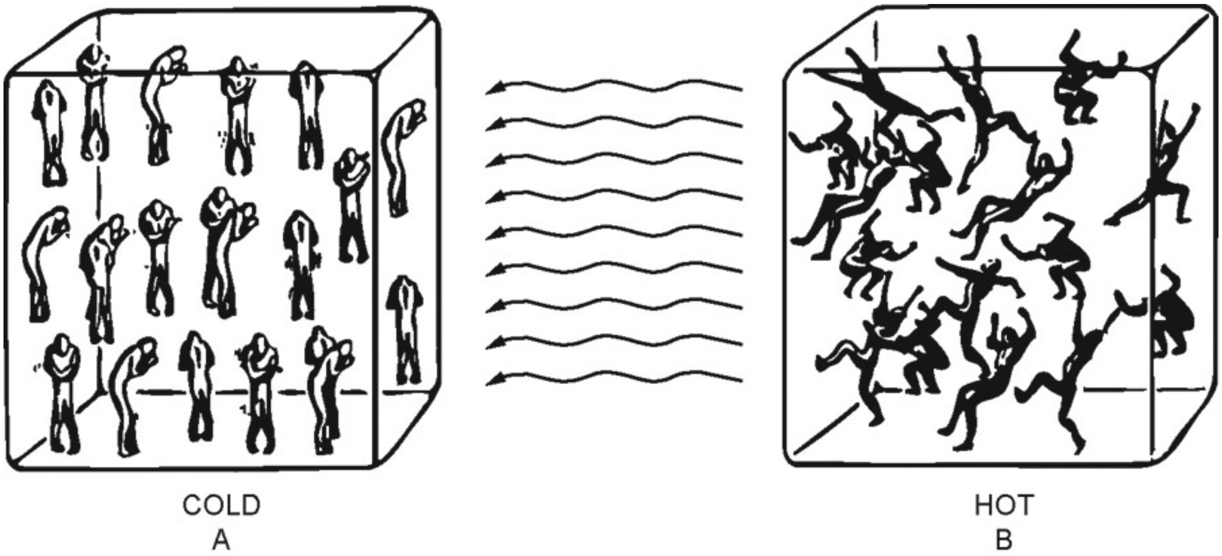


Figure 1–27. Hot and cold bodies.

Thermal energy is transferred between objects in one direction—warm to cold. That is, a warm body naturally transfers some of its thermal energy to a colder body. For example, of the two objects in figure 1–27 object B may transfer some of its energy to object A until the molecules of both objects have an equal amount of thermal energy—that is, until their temperatures are the same. When ice cubes cool a drink, the heat of the drink is transferred to the colder ice cubes. This is another example of thermal energy being transferred from warm to cold.

Scales used to measure temperature

Although there are many temperature scales, in meteorology you need only be concerned with three—Celsius, Fahrenheit, and Kelvin (absolute scale). To compare the three scales, we must define two fixed temperatures and a scale difference between them. The two fixed temperatures are the freezing and boiling temperatures of water. Compare the freezing and boiling points of each scale shown in the following table. Notice that the Fahrenheit scale has 180° between freezing and boiling; in the other two, there are only 100°.

Scale	Freezing Point	Boiling Point
Fahrenheit	32°	212°
Celsius	0°	100°
Kelvin	273°	373°

Conversion of temperature scales

The use of three different temperature scales requires you to know how to convert from one scale to another. There are even times when you must report observed temperatures in two different scales. You'll need to know how to convert Fahrenheit to Celsius, Celsius to Fahrenheit, and make conversions to Kelvin.

Fahrenheit and Celsius conversions

Recall that the Fahrenheit scale has 180° between freezing and boiling, while the Celsius scale has 100°. Therefore, 1.8° on the Fahrenheit scale equals 1° on the Celsius scale. Also remember that the different values for the freezing point are 32°F and 0°C.

Therefore, to convert Celsius to Fahrenheit, multiply the Celsius temperature by 1.8 and add to 32. The formula for this conversion is: F = 1.8C + 32. Now let's test the formula to see if 0°C converts to 32°F.

$$F = 1.8C + 32$$

$$F = 1.8(0) + 32$$

$$F = 0 + 32$$

$$F = 32°$$

We can use the same formula to convert from Fahrenheit to Celsius. This is shown below by converting 32° F to its Celsius equivalent:

$$32 = 1.8(C) + 32$$

$$32 - 32 = 1.8(C) + 32 - 32$$

$$0 = 1.8(C)$$

$$\frac{0}{1.8} = \frac{1.8(C)}{1.8}$$

$$0° = C$$

In simple terms, to convert from Fahrenheit to Celcius use the following steps:

- Take the temperature in Fahrenheit subtract 32.
- Divide by 1.8.
- The result is degrees Celsius.

Kelvin conversions

Conversion to the Kelvin scale is a simple process. Just like the Celsius scale, the Kelvin scale has 100 divisions between freezing and boiling. Therefore, since the Kelvin reading for freezing is 273 and the Celsius reading is 0, you can convert from Celsius to Kelvin (K) by adding 273 to the Celsius reading: $K = C + 273$. Conversion from Kelvin to Celsius is just the reverse or: $C = K - 273$. When you need conversions between Kelvin and Fahrenheit, use the conversion to Celsius as an intermediate step.

217. Moisture measurement in the atmosphere

Moisture in our atmosphere is one of the most important elements in the study of meteorology. Obviously, without it there would be no clouds or rain; beyond that, the amount of moisture in the atmosphere has a great effect on how the atmosphere behaves. For example, as water vapor content increases, the air becomes less dense and rises—a sign of an unstable atmosphere. As you know, an unstable atmosphere is conducive to thunderstorm development.

Water vapor pressure

Water vapor pressure is a measure of the amount of water vapor in the atmosphere. From the last lesson you know that *atmospheric pressure* is the weight of a column of air over a given location. Since water vapor is a gas, it is included in the measure of atmospheric pressure. The English physicist Dalton proved that the pressure of a fixed volume of gas was equal to the sum of the partial pressures exerted by each different gas within the volume. The amount of water vapor pressure in a column of air depends greatly on the temperature of the column of air. Thus, as you may recall from earlier schooling, warm air can hold more water vapor than cold air.

Vapor pressure

Vapor pressure (not to be confused with water vapor pressure) is a measure of the ability of a moisture source to exchange water molecules with the surrounding air. For example, the vapor pressure of a body of water or ice refers to the vapor pressure in the minute space immediately above the surface of the body of water or ice. Comparing the two, we find that vapor pressure is greater over water than over ice. The reason for this is that the molecules of water, warmer than those of ice, move about more freely. This increased molecular activity allows the molecules of water to escape or evaporate into the atmosphere more readily.

Saturation vapor pressure

Water molecules are constantly in motion; constantly changing the water's state from liquid to gas, solid to gas, liquid to solid, and so forth. When the number of molecules escaping and the number returning are equal, saturation vapor pressure is reached. Since saturation vapor pressure is temperature dependent, it varies directly with temperature.

Relative humidity

Relative humidity (RH) is the ratio of actual water vapor in the air to the amount it can hold. You can also determine relative humidity from the ratio of vapor pressure to saturation vapor pressure:

$$RH\,\% = \left(\frac{e}{e_s}\right)(100)$$

where:

RH = relative humidity.

e = vapor pressure.

e_s = saturation vapor pressure.

100 = value used to convert decimal to percent.

Therefore, if the vapor pressure (e) equals 35mb and the saturation vapor pressure (e_s) equals 80mb, compute relative humidity as follows:

$$RH\,\% = \left(\frac{e}{e_s}\right)(100)$$

$$RH\,\% = \left(\frac{35}{80}\right)100$$

$$RH\,\% = 0.4375(100)$$

$RH = 43.75\%$

Relative humidity is the most commonly used expression of water vapor content in the atmosphere.

Mixing ratio

Because saturation vapor pressure and relative humidity vary with temperature, they are not exact measures of water vapor content. Therefore, we often use another measure—mixing ratio. Mixing ratio is the mass of water vapor (M_v) compared to the mass of dry air (M_d) contained in a volume of natural air.

NOTE: Dry air is air with all the moisture removed. Natural air is dry air plus water vapor.

Mixing ratio is a more accurate measure of water vapor because of the nature of mass. Remember that the only way to change mass is to add or remove matter. The mixing ratio is expressed in grams per kilogram (g/kg), and the formula for computing it is:

$$w = \frac{M_v}{M_d}$$

where:

w = mixing ratio.

M_v = mass of water vapor.

M_d = mass of dry air.

Although we can use this formula to compute mixing ratio, the simplest way to compute this value would be by using a plotted Skew-T, Log P diagram. The procedure for finding the mixing ratio for a given pressure is accomplished by reading the value directly (or by interpolation) of the saturation mixing-ratio line that crosses the dew-point curve (T_d) at that pressure. For example, if the 700mb T_d is –13˚ C; the intersection of the saturation mixing-ratio line and the –13˚ C T_d isotherm at 700mb would equate to a mixing ratio of:

$$\frac{2.0\text{ g}}{\text{kg}}$$

Absolute humidity

Absolute humidity is another expression of mixing ratio. We express it as a percent; therefore, to compute absolute humidity, multiply mixing ratio by 100.

Specific humidity

Specific humidity is another measure of water vapor content that uses comparison of masses—it is the ratio of water vapor to the total mass of the natural air. We also express specific humidity in terms of grams per kilogram; but, because specific humidity is the ratio of water vapor to total mass, it is always a lower value than mixing ratio or absolute humidity. The formula for specific humidity is:

$$q = \frac{(0.622\, e)}{(P - 0.378\, e)}$$

where:

q = specific humidity.

0.622 = specific humidity constant.

e = vapor pressure.

P = atmospheric pressure.

0.378 = specific humidity constant.

Below is an example of a specific humidity computation using an atmospheric pressure of 988mb and a vapor pressure of 35mb.

Formula:

$$q = \frac{(0.622\, e)}{(P - 0.378\, e)}$$

Input factors:

$$q = \frac{(0.622)35}{[988 - (0.378)(35)]}$$

Calculate:

$$q = \frac{21.77}{[988 - 13.23]}$$

$$q = \frac{21.77}{974.77}$$

Answer:

q = 0.022 grams per kilogram

NOTE: Specific humidity calculation is preferable for very precise physical and theoretical work. For synoptic purposes, however, the mixing ratio is sufficiently representative and is easier to evaluate.

218. General tools for temperature forecasting

Just like every other surface weather element presented in this volume, there are many tools that you can pull out of the weather tool box to assist you in forecasting. Many tools can be used for forecasting more than one element. The list of tools that you use for forecasting temperature include climatology, model output statistics, and AFWA meteograms.

Climatology

Your first step in forecasting temperatures is to use climatology. By using climatology you get an idea what is likely to occur and what is not likely to occur. Several sources of temperature climatic data is available from the 14th Weather Squadron. Let’s look at a few of the climatic tools you can use to forecast temperatures.

Modeled curves

Modeled curves (MODCURVES) is a computer program that uses climatology as a guide to provide the temperature baseline for the time of year and time of day. The display can be adjusted to meet current or expected weather conditions, such as cloud cover and winds that affect temperature forecasts.

Surface observation climatic summaries

Surface observation climatic summaries (SOCS), Part E, includes temperature, dry-bulb, wet-bulb, dew point, and relative humidity information for specific bases and posts for over an extended period. A station must have a minimum of 5 years of recorded observations to have a SOCS.

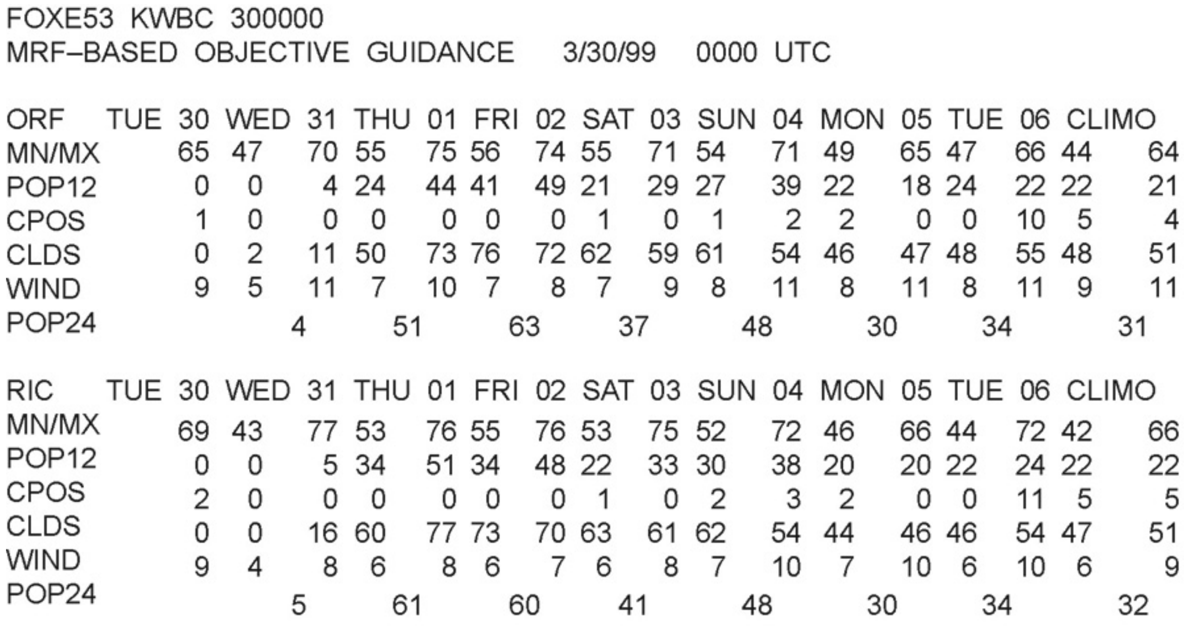
Operational Climatic Data Summary

This climatogy tool provides a summary of monthly and annual climatic data for a station. A forecaster can use the Operational Climatic Data Summary (OCDS) tool to a get a general idea of what type of conditions to expect. OCDS provides the extreme maximum temperature, the mean daily maximum temperature, the mean temperature, the mean daily minimum temperature, and the extreme minimum temperature for any given month. This tool works well for getting an idea of general temperatures at time ranges past model data.

Model output statistics

Model output statistics (MOS) is an excellent tool to help forecast of temperatures (fig. 1–28). MOS guidance derives its forecasting relationships by correlating past model output with station climatology. It is imperative to initialize and verify the model before using its MOS.

MOS can provide guidance on forecasting maximum and minimum temperatures out to the time range of the model. MOS can also provide time-specific, two-meter temperature forecasts. Since two meters is the height of most temperature measuring instruments, we use it in the computations.



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FOXE53 KWBC 300000
MRF–BASED OBJECTIVE GUIDANCE   3/30/99   0000 UTC

ORF    TUE 30 WED 31 THU 01 FRI 02 SAT 03 SUN 04 MON 05 TUE 06 CLIMO
MN/MX      65  47   70 55   75 56   74 55   71 54   71  49   65 47   66 44   64
POP12       0   0    4 24   44 41   49 21   29 27   39  22   18 24   22 22   21
CPOS        1   0    0  0    0  0    0  1    0  1    2   2    0  0   10  5    4
CLDS        0   2   11 50   73 76   72 62   59 61   54  46   47 48   55 48   51
WIND        9   5   11  7   10  7    8  7    9  8   11   8   11  8   11  9   11
POP24               4      51      63      37      48       30      34      31

RIC    TUE 30 WED 31 THU 01 FRI 02 SAT 03 SUN 04 MON 05 TUE 06 CLIMO
MN/MX      69  43   77 53   76 55   76 53   75 52   72  46   66 44   72 42   66
POP12       0   0    5 34   51 34   48 22   33 30   38  20   20 22   24 22   22
CPOS        2   0    0  0    0  0    0  1    0  2    3   2    0  0   11  5    5
CLDS        0   0   16 60   77 73   70 63   61 62   54  44   46 46   54 47   51
WIND        9   4    8  6    8  6    7  6    8  7   10   7   10  6   10  6    9
POP24               5      61      60      41      48       30      34      32
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Figure 1–28. MOS temperature guidance.

Temperature forecasting checklist

Forecasting temperature seems easy but it can be very challenging. Temperature is critical and can mean the difference, for instance, between liquid and frozen precipitation. It’s a good idea is to have a temperature forecasting checklist at your unit (fig. 1–29). The checklist simply collates all the temperature forecasts from several different products or methods and displays them. A forecast worksheet can serve the same purpose. This way you can make your best forecast with all available data at hand.

TEMPERATURE GUIDE

CLIMATOLOGY	EXTREME	MAX ______		MIN ______	
	AVERAGE	MAX ______		MIN ______	
YESTERDAY'S		MAX ______		MIN ______	
FRONTAL PASSING		YES ______		NO ______	
NWS CHART (or equivalent)		MAX ______		MIN ______	
MOS BULLETIN (or equivalent)		MAX ______		MIN ______	
REPRESENTATIVE SKEW–1		MAX ______	AT ______	MIN ______	AT ______
YOUR FORECAST		MAX ______	AT ______	MIN ______	AT ______

Figure 1–29. Temperature forecasting checklist.

219. Factors that affect temperature

Temperature, being subject to marked changes from day to night, is not considered a conservative property of an air mass. Additionally, it does not always display a uniform lapse rate from the surface up through the atmosphere. This means there are many reasons why the surface air temperature may not be representative.

Insolation and terrestrial radiation definitely affect temperature. Low-latitude stations, for instance, receive more heat during the day than do stations at high latitudes. More daytime heat can be expected in summer than in winter, since in the summer the sun’s rays are more direct and reach the earth for a longer period. Normally, there is a net gain of heat during the day and a net loss at night. Consequently, the maximum temperature usually occurs during the day and the minimum at night. Clouds reduce insolation and terrestrial radiation, lowering daytime temperatures and raising nighttime temperatures over what should be expected. With a stable lapse rate there is less vertical extent to heating effects and, therefore, the surface heating occurs more rapidly. With an unstable lapse rate, the reverse is true; surface heating occurs more slowly. If there is an inversion, there is less cooling at night, since the surface temperature is lower than that of the inversion layer.

One of the biggest factors affecting temperature is advection. Temperature advection is particularly noticeable with frontal passage. Advection within an air mass is also important, as it is with the warming from warm low-level flow from the south.

Vertical heat transport is another temperature factor. It is considerably affected by the speed of the wind. With strong winds there is less effect from heating and cooling than with light or calm winds, since the heat energy gained or lost is distributed through a deeper layer when the turbulence is greater.

Evaporation and condensation also affect the temperature of an air mass. When cool rain falls through a warmer air mass, evaporation occurs, taking heat from the air. This often affects the maximum temperature on a summer day where afternoon thunderstorms occur. The temperature at the surface may be affected a little by condensation during fog formation, raising the temperature a degree or so because of the release of the latent heat of condensation to the air.

220. Heat waves

In summer, heat wave forecasts furnish a warning that very high unpleasant temperatures and humidities are impending. The definition of a heat wave varies from place to place. For example, in the Chicago area, temperature above 90°F on three successive days is a heat wave.

Heat waves develop over the Midwestern and eastern part of the US when a long-wave trough stagnates over the Rockies, or the Plains states, and a long-wave ridge lies over or just off the east coast and the belt of westerlies is centered far north in Canada. At the surface, there is a slow and poorly organized low-pressure system over the Great Plains or Rocky Mountains. Pressure is usually above normal over the South Atlantic and Middle Atlantic States. An exception occurs to the scenario described above when the amplitude of the flow pattern aloft becomes very great. As a result of this exemption (the increase in amplitude) several anticyclonic centers develop in the eastern ridge, both at upper levels and at the surface. The heat wave generally continues until the long wave begins to move.

221. Cold waves

A cold wave in the US is a net temperature decrease of 20°F or more in 24 hours with the temperature falling below a preset minimum. The minimum varies with geographical location and the time of the year.

A cold wave is usually caused by very cold continental polar (cP) or arctic air mass over west central Canada, with a low moving eastward from the Continental Divide, and large pressure tendencies (3 or 4mb) occurring behind the cold front. Aloft, a ridge of high pressure develops over the western part of the US or just off the west coast, as a short-wave trough moves into a long-wave trough over the central part of the US and deepens. An increase in intensity of the southwesterly flow over the eastern Pacific frequently precedes the building of the ridge. Often, retrogression of the long wave takes place. Strong northerly to northwesterly flow is established aloft through a deep layer and sets the polar or arctic air in motion southward.

Most cold waves do not persist. Temperatures start upward after about 48 hours. Sometimes, however, the upper ridge over the western part of the US and the trough over the eastern part is quasistationary, and a large supply of very cold air remains in Canada. Then, we experience successive outbreaks with northwest steering that holds temperatures well below normal for as long as two weeks.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

216. Measuring temperature

1. Explain the relationship between temperature and molecular activity in an object.

2. Determine what happens to thermal energy between two objects of different temperature.

3. Name the three temperature scales primarily used in meteorology and give the freezing and boiling point for each.

4. Make the following temperature conversions.
    a. 45°F to degrees Celsius.

    b. 24°C to degrees Fahrenheit.

    c. 30°C to degrees Kelvin.

    d. 425°K to degrees Fahrenheit.

217. Moisture measurement in the atmosphere

1. Match the measures of humidity in column B with the definitions in column A. Column B items may be used more than once.

Column A	*Column B*
____ (1) Indicates an equal exchange of water vapor between the air and a moisture source.	a. Water vapor pressure.
____ (2) Affected by temperature.	b. Vapor pressure.
____ (3) Ratio of actual water vapor to amount of water vapor that a quantity of air can hold.	c. Saturation vapor pressure.
____ (4) Partial pressure of the atmosphere due to the presence of moisture.	d. Relative humidity.
____ (5) Ratio of mass of water vapor to mass of dry air.	e. Mixing ratio.
____ (6) These two measures of atmospheric moisture are the same.	f. Absolute humidity.
____ (7) Ratio of mass of water vapor to mass of natural air.	g. Specific humidity.

218. General tools for temperature forecasting

1. What tool provides a summary of monthly and annual climatic data for a station?

2. What tool can be used to collate temperature forecasts from several sources or products?

219. Factors that affect temperature

1. Is temperature a conservative property in an air mass?

2. Clouds affect the temperature by reducing what?

3. How does cloud cover affect temperatures during the day and at night?

4. How does a stable lapse rate affect heating?

5. How do strong winds affect heating?

6. How does condensation affect surface temperatures?

220. Heat waves

1. What conditions describe a heat wave?

2. How does a heat wave develop?

221. Cold waves

1. What is a cold wave?

2. How does a cold wave develop?

1–6. Atmospheric Pressure

Atmospheric pressure is an extremely important meteorological parameter for aviation. The pressure is used to adjust the aircraft's altimeter to indicate the proper altitude during flight. If the value used to adjust the instrument is incorrect, it could be extremely hazardous to the aircrew and the aircraft. In this section, you'll explore forecasting concepts using these pressure measurements:

- D-values.
- Density altitude.
- Pressure altitude.
- Altimeter setting.
- Sea-level pressure.

Let's begin by looking at forecasting techniques using the pressure reference for all pressure values—sea-level pressure (SLP).

222. Forecasting sea-level pressure

The SLP is the atmospheric pressure at mean sea-level. It can be measured at sea-level or determined from an observed station pressure at other locations and is reported in millibars. Standard atmospheric SLP is 1013.2 mb or 29.92 inches of mercury. Since SLP is not part of a terminal aerodrome forecast TAF, it is not forecast on a regular basis.

However, you may need to forecast the SLP for a location to use in computing forecast winds. The easiest methods to use to forecast SLP are estimating the SLP from centralized charts, numerical bulletins, or meteograms. Since these products rely on a computer model to produce the data, you should always verify the model to see how it is working before using the data. All in all, these methods are accurate and easy to use.

Forecast models and meteograms can also be used as an excellent tool for forecasting surface pressure. Using the JAAWIN website the forecaster can obtain sea-level pressure data out to 384 hours (GFS long range forecast). The surface temperatures/sea level pressure chart forecasts pressure values in 4 millibar isobar intervals. Interpolation of the pressure data is required by the forecaster. Additionally, meteograms are also another tool that can be used to determine sea-level pressure for your station and can be used to graphically depict forecast conditions for a specific location.

223. Altimeter errors

An altimeter is an aneroid barometer calibrated to indicate altitude instead of pressure. When the altimeter is properly set, the altitude indicated corresponds to the equivalent pressure in the standard atmosphere. Thus, the altimeter is always indicating a pressure height and not an actual height. For example, when the aircraft is flying at the 500mb surface, the altimeter indicates that the height in MSL of the aircraft is 18,289 feet, whether it is or not. As you probably realize, this can create hazards for both the pilot and weather journeyman if not careful in determining altimeter settings and setting them accordingly.

The major errors in altimeter settings come from the difference of the actual atmosphere from the standard atmosphere caused by temperature and/or pressure.

Pressure difference errors

The error caused by the deviation of actual sea-level pressure from the 29.92 inches of water gauge (Hg) to sea-level pressure assumed in the standard atmosphere is corrected by adjusting the altimeter setting in the aircraft. Altimeter settings reflect the pressure of the reporting station, in inches of mercury, converted or altered, to make the altimeter read zero elevation at sea level. In other words, it shows the station elevation when the aircraft is on the ground at the station.

Although the correct altimeter setting is used at takeoff, the altitude indicated during a cross-country flight and at the time of landing is usually in error. This is caused by the variation of pressures en route and at the destination. Before landing, a pilot should always reset the altimeter to the correct altimeter setting of the destination. Only then does the altimeter read field elevation on touchdown.

Figure 1–30 depicts an example of the error introduced when a pilot does not change altimeter settings from departure to destination. The current setting at Miami, the departure base, was 30.20 inches of mercury (Hg). The correct setting at New Orleans is 29.70 inches of Hg. Unless the setting is changed en route, an aircraft flying where the indicated altitude remains constant travels along a constant pressure surface, since the pressure altimeter is simply an aneroid barometer.

Let’s say that the 850mb surface was flown. From figure 1–31, you see the 850mb surface is lower over New Orleans than over Miami. Therefore, the aircraft flying a constant pressure surface would be lower in true height at New Orleans than over Miami. However, since the altimeter is still reading the height of the 850mb surface, it would indicate that the aircraft was higher. Considering pressure error alone, flying toward lower pressure at flight level with a constant altimeter setting causes the altimeter to read too high; meaning the aircraft is at a lower altitude than indicated by the altimeter. We’ll see in the discussion of combined errors later that this pressure error is magnified when a low exists over a surface high.

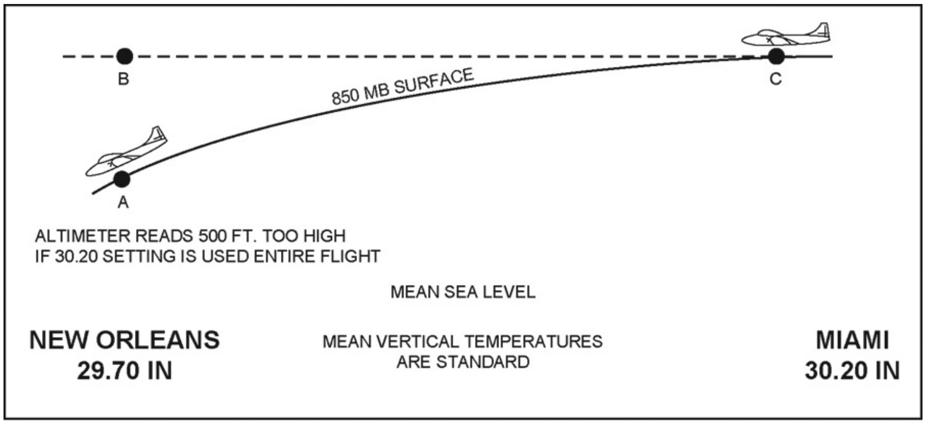


Figure 1–30. Error in altitude caused by not adjusting the altimeter.

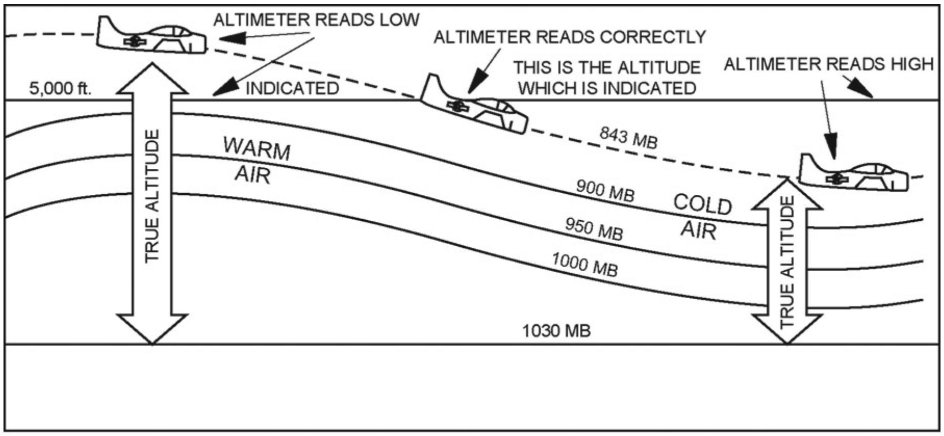


Figure 1–31. Temperature-caused altimeter error.

Temperature difference errors

An altimeter that is correctly set on takeoff is frequently in error at higher altitudes over the departure base because of the deviation of the actual mean vertical temperature from that of the standard atmosphere. If the air temperature through a vertical column is warmer than standard, the air density is less than standard, meaning a pilot must fly higher to reach a given indicated altitude than would be required if the mean temperature were standard. That is, the atmosphere is not so “tightly packed” and the aircraft ascends through its mass less rapidly (per unit of height). Here, the altimeter reads lower than the actual altitude even with the current setting that was set upon takeoff

If the air temperature in the vertical column is colder than standard, the air density is greater than standard, so a pilot would fly lower to attain a given indicated altitude than would be required if the mean temperature were standard. That is, the atmosphere is now “tightly packed” and the aircraft ascends through its mass quite rapidly (per unit of height). Now, the altimeter reads higher than the actual altitude even with the current setting.

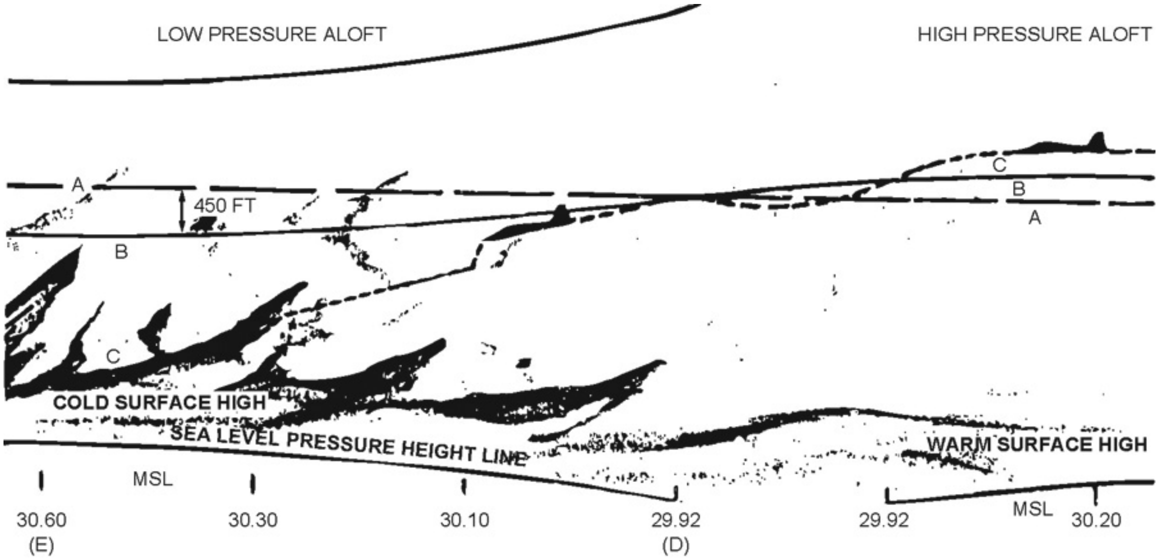
Perhaps, the most critical aspect of temperature variation exists when the mean temperature of a column of air is colder than that of the standard atmosphere. The correction of the altimeter is most important in evaluating terrain clearance, particularly when flying at minimum instrument altitudes. Many aircraft accidents in mountainous areas have been attributed to altimeters indicating higher than the actual altitude on flights in cold air (fig. 1–31).

The greater the departure of the actual vertical temperature lapse rate from the standard, the greater the introduced error. However, it is entirely possible that the temperature at an aircraft’s altitude could be lower than standard with a pronounced temperature inversion between the aircraft's altitude and the surface. Then, it would appear that the mean temperature is colder than standard, when actually it is warmer. Then, correcting pressure altitude for temperature would magnify the error, not correct it.

Combined errors

Under certain conditions, pressure errors and temperature errors may combine to produce an error as great as 2,000 feet between the indicated altitude and the actual true altitude.

The example in figure 1–32 shows a flight proceeding from one area of high surface pressure to another area of high surface pressure. Although altimeter settings are received at regular intervals along the route and flight corrections are made accordingly, the net result in this example is a gradual loss of actual altitude above MSL. This loss results from the altimeter setting being based on surface pressure with no direct consideration of a reversal aloft to lower or higher pressure. Notice that if a pilot made an actual flight under the conditions as illustrated, disaster could result if the aircraft were flying at minimum altitude clearance in mountainous terrain. However, the information needed to make such a flight safely is readily available in the weather station.



A. PROPOSED FLIGHT PATH OF 11,000 FEET MSL

B. ACTUAL HEIGHT OF PRESSURE LEVEL EQUIVALENT TO 11,000 FEET MSL IN STANDARD ATMOSPHERE

C. FLIGHT PATH USING ALTIMETER SETTINGS FROM STATIONS ALONG ROUTE. (INDICATING 11,000 FEET ALL THE WAY)

Is this aircraft flying from a Low toward a High, or from a High toward a Low? Do we consider the pressure pattern at the ground or at flight level? Note in the diagram that, between Points (D) and (E), the pressure pattern at flight level is exactly the reverse of the surface pattern.

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Figure 1–32. Pressure-caused altimeter error.

In our illustration, a flight is planned into an area of a cold high at the surface coupled with a cold low aloft. This is the most dangerous situation for terrain clearance, but is a common occurrence in the atmosphere. If this flight were conducted with an altimeter setting of 29.92 inches, the flight path would closely approximate curve "B" in figure 1–32. However, the pilot has used various altimeter settings along the route as required by FAA regulations. Notice that flying toward the cold surface high requires use of consecutively higher altimeter settings because of the higher surface pressure. Each time the pilot raises the setting, the altimeter reads a corresponding increase in indicated altitude, although the true altitude is actually decreasing. (Remember that the altimeter is reading the standard height of the pressure surface.) Here, this is based on a surface high that does not exist aloft. To maintain a constant indicated altitude of 11,000 feet, the pilot must descend each time a higher setting is used (as the altimeter setting is raised the indicated altitude increases), since the actual pressure height is much lower than standard.

Now let’s see how low our aircraft would be when it reaches the left of the sketch (fig. 1–32). The altimeter setting near the end of the flight is 30.60 inches. Since one inch of pressure is equivalent to approximately 1,000 feet of altitude near sea level in the standard atmosphere, we multiply the 0.68 inch difference (30.60 - 29.92) by 1,000. This gives a difference of 680 feet between the true pressure altitude and the indicated altitude with a 30.60 setting.

224. Difference values

The difference (D)-value is the difference between the true altitude of a pressure surface and the standard atmosphere altitude of this pressure surface.

D-value = true altitude – standard altitude

Suppose a flight is proposed at 11,000 feet MSL, by using the closest constant pressure product to the proposed altitude (here, 700mb), you can determine the D-value. The standard height of the 700mb level is 9,780 feet above sea level. In our case, assume that the upper-air data indicates that the 700mb level is at 9,200 feet. Therefore, the D-value is 580 feet (9,200 - 9,780). The aircraft, then, would be flying 580 feet below the indicated altitude if its altimeter had been set at 29.92 on takeoff, as in figure 1–31.

In a previous lesson, you saw that the aircraft would end up 680 feet lower over its destination because of the change in altimeter setting. Thus, both corrections must be added to determine the actual height of the aircraft over its destination. The D-value (–580 feet) plus the surface correction (–680 feet) equals –1,260 feet. Therefore, the actual height of the aircraft is 9,740 feet (11,000 - 1,260) above sea level.

225. Pressure altitude

The pressure altitude (PA) is the elevation in the standard atmosphere at which a given pressure occurs.

For example, let’s say that the PA for your airfield is +3,000 feet. This means that any aircraft landing or departing performs as though it is at 3,000 feet of elevation, no matter whether the field elevation is 3,000 feet or not.

The PA is easily computed. All you need to know is the field elevation and the altimeter setting. Following is a step-by-step procedure for computing the PA using a field elevation of +750 feet and an altimeter setting of 30.02:

Step 1: Algebraically subtract the altimeter setting from 29.92.

29.92 – 30.02 = - 0.10

Step 2: Multiply the answer in *Step 1* by 1,000.

0.10(1,000) = –100

Step 3: Algebraically add the answer in *Step 2* with the field evaluation.

100 + 750 = 650

Therefore, the pressure altitude is +650 feet.

Suppose you want to compute the PA for an airfield with a field elevation of +330 feet and an altimeter setting of 29.76. The same procedures would be applied.

Step 1: 29.92 - 29.76 = 0.16

Step 2: 0.16(1,000) = 160

Step 3: 160 + 330 = +490 feet

This method can be used to compute the PA for an airfield since you know the field elevation and the altimeter setting for the time of concern.

226. Density altitude

Density altitude is the altitude at which a given density is found in the standard atmosphere. The effects of air density on heavy airlift operations can be mission limiting. For example, if the pressure altitude at a 6,000-foot mountain location is equal to the standard-atmosphere pressure for 6,000 feet, but the air temperature is 100°F, over 60°F warmer than the standard-atmosphere at that height, the density altitude would be near 10,000 feet. Under these conditions, runway length requirements for some fixed-wing aircraft could increase by 50 percent or more, and hover-lift capabilities of some rotary-wing aircraft could be exceeded.

Density altitude cannot be measured directly, but must be computed from pressure and virtual temperature (temperature at which dry air would have the same density as a moist air sample) at the altitude or location under consideration. The easiest way of computing density altitude is to use a Skew-T product that has an overprinted density altitude nomogram or by using the tables presented in AWSTR 165, *Forecasting Density Altitude*.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

222. Forecasting sea-level pressure

1. What sources can you obtain forecasted sea-level pressure values from?

2. Which product can provide you with a graphical indication of a frontal passage in the sea-level pressure?

223. Altimeter errors

1. Why is an altimeter necessary?

2. How does pressure create an error in altimeter readings?

3. How does temperature create an error in altimeter readings?

224. Difference values

1. What is the D-value?

2. What is the D-value for 9,000 feet if the height of the 700mb level is 10,240 feet?

225. Pressure altitude

1. Compute the pressure altitude for each of the following situations:

Altimeter	Field Elevation	Forecasted PA
a. 29.84	+150	_____
b. 29.96	+460	_____
c. 30.05	+105	_____
d. 29.72	+890	_____
e. 29.92	+30	_____

226. Density altitude

1. What is density altitude?

2. What elements are used in computing density altitude?

Answers to Self-Test Questions

201

1. Solid, liquid, and gaseous.
2. (1) d.
    (2) f.
    (3) a.
    (4) e.
    (5) c.
    (6) b.
3. Take.
4. Release.
5. Release.
6. Take.
7. Take.
8. Release.

202

1. Solute; if the droplet has a vapor pressure that is less than the environment ($e_{s\ env} > e_{s\ droplet}$) then the droplet grows due to evaporation. Curvature; Small droplets that are tightly curved have a larger saturation vapor pressure than bigger droplets that are less curved ($e_{s\ small} > e_{s\ big}$). In this case, if the droplet has a larger e_s than the environment, it becomes smaller due to evaporation.
2. Cumulonimbus clouds. The residence time in cloud would be increased as updrafts in the cloud would increase distance and coalescence time of the droplet.
3. Variable droplet fall velocities.

203

1. Adiabatic cooling.
2. $\frac{1}{50}$; of the two, this front has the steeper slope, stronger lifting capabilities, and the adiabatic cooling process increases faster.
3. To the east of your station due to low-level speed convergence occurring there.
4. With westerly winds, orographic lift and adiabatic cooling would produce the clouds on the windward side. However, as the wind continues to flow down the leeward side of the mountains, adiabatic warming would occur and dissipate the clouds.

204

1. Any four of the following: climatology, model and centralized guidance, forecast relative humidity values, extrapolation and weather radar.
2. Climatology.
3. Forecast relative humidity values.
4. Extrapolation.
5. Weather radar.

205

1. Atmospheric stability.
2. Lifting condensation level.
3. The level in the atmosphere where condensation occurs with a parcel of air due to adiabatic cooling.
4. Strong instability.
5. Dew point depression method.

206

1. A stable atmosphere.
2. Where the RH reaches or exceeds 65 percent.
3. Where it decreases below 65 percent.
4. They tend to recur frequently at the same altitude, less than 1,000 feet.
5. The wind may allow fog or it may blow the stratus cloud away, depending on its strength.

207

1. It is not necessary for the system to show on the surface.
2. Only when the air is moving in a cyclonically curved path.
3. After the 500mb ridge has passed.

208

1. (1) No precipitation.
    (2) Yes, but the type is not predictable.
    (3) No precipitation.
    (4) Yes, shower type precipitation.
    (5) Yes, continuous type precipitation.

209

1. Small scale.
2. Polar front.
3. Eastward.

210

1. a. C
    b. D.

- c. C.
- d. D.

2. a. C, addition of moisture.
   - b. C, addition of moisture.
   - c. D, warming of air.
   - d. C, cooling of air.
   - e. D, warming of air.
   - f. C, addition of moisture.
   - g. C, cooling of air.
   - h. D, warming of air.
   - i. C, cooling of air.
   - j. D, removal of moisture/warming of air.

211

1. Radiation fog.
2. The initial temperature and the rate and duration of net long-wave radiation transfer.
3. Sea fog.
4. Upslope fog.
5. Ice fog.

212

1. Haze.
2. Smoke.
3. Windblown restrictions.
4. Light.

213

1. Topography.
2. Trends.
3. Meteograms.
4. 11 to 20 knots.

214

1. 12 to 21 knots.
2. 30 knots with gusts to 48 knots.
3. 30 knots with gusts to 40 knots.
4. 30 knots with gusts to 40 knots.
5. Through extrapolation of current conditions with modifications based on expected changes.

215

1. Local heating, local cooling, adjacent heating and cooling, and forced circulations.
2. Mountain breeze; local cooling.
3. Valley breeze, local heating.
4. Adjacent heating/cooling.
5. Fall winds.
6. Forced circulation.
7. Foehn winds.

216

1. The molecules in a warm body are more active than in a cold body.

2. Thermal energy is transferred from the warm body to the cold body.
3. Celsius scale where freezing is 0° and boiling is 100°; Fahrenheit scale where freezing is 32° and boiling is 212°; and Kelvin or absolute scale where freezing is 273° and boiling is 373°.
4. a. 7.20°C.
   b. 75.2°F.
   c. 303°K.
   d. 305.6°F.

217

1. (1) c.
   (2) a, b, c, d.
   (3) d.
   (4) a.
   (5) e, f.
   (6) e, f.
   (7) g.

218

1. OCDS.
2. Temperature forecasting checklist.

219

1. No, it is nonconservative.
2. Insolation and terrestrial radiation.
3. It means a lower temperature in day and higher at night.
4. Heating occurs more rapidly.
5. There is less heating due to greater mixing.
6. The surface heats only a little bit.

220

1. A heat wave varies from place to place but it generally refers to very high unpleasant temperatures and humidities.
2. The long-wave trough stagnates over the Rockies and a long-wave ridge is over the eastern US. The jet is in Canada and the surface pressure is higher than normal in the eastern US.

221

1. A cold wave in the U S is a net decrease of 20°F or more in 24 hours with the temperature falling below a preset minimum.
2. A very cold continental polar air mass in Canada moves southward near the surface beneath a strong ridge over the western US and a strong trough in the eastern US.

222

1. Centralized charts, numerical bulletins, and meteograms.
2. Meteogram.

223

1. It indicates the height of an aircraft with respect to MSL.
2. The altimeter is based on standard atmosphere. Pressure variances from standard result in altimeter errors.
3. The altimeter is based on the standard atmosphere. When the temperature differs from the temperature expected in a standard atmosphere, the altimeter is in error.

224

1. D = true altitude - standard altitude.

2. D = +460 feet.

225

1. a. +230 feet.
   b. +420 feet.
   c. –25 feet.
   d. +1,090 feet.
   e. +30 feet.

226

1. The altitude that a given density is found in the standard atmosphere.
2. Pressure and virtual temperature at an altitude or location under consideration.

Do the unit review exercises before going to the next unit.

Unit Review Exercises

Note to Student: Consider all choices carefully, select the *best* answer to each question, and *circle* the corresponding letter. When you have completed all unit review exercises, transfer your answers to the Form 34, Field Scoring Answer Sheet.

Do not return your answer sheet to Air Force Career Development Academy (AFCDA).

1. (201) Which term defines a change in state from a solid to a gas?
a. Fusion.
b. Freezing.
c. Deposition.
d. Sublimation.

2. (201) What changes of state absorb heat from the surrounding environment?
a. Condensation, fusion, and deposition.
b. Condensation, freezing, and deposition.
c. Vaporization, fusion, and sublimation.
d. Vaporization, freezing, and sublimation.

3. (202) What is the *most efficient* method of cloud droplet growth?
a. Collision and coalescence.
b. Collision and evaporation.
c. Condensation and collision.
d. Condensation and coalescence.

4. (203) What term is used to define the process where the outer edge of the cloud mixes with air outside of the cloud?
a. Evaporation.
b. Condensation.
c. Dry-air entrainment.
d. Moist-air entrainment.

5. (204) Which cloud forecasting tool is biased towards climatology (average conditions), and may not be accurate during abnormal conditions?
a. Meteograms.
b. Model output statistics.
c. Conditional climatology tables.
d. Forecast relative humidity values.

6. (205) What is the *main* difference in forecasting the occurrence of convective or cumuliform clouds versus stratus clouds?
a. Vertical velocity.
b. Degree of cloudiness.
c. Atmospheric stability.
d. Dew-point and free-air temperature spread.

7. (206) You would *not* expect stratus clouds to form when
a. the curvature is anticyclonic.
b. warm stable air is being lifted over a warm front.
c. onshore winds are in an area of cyclonically curved contours.
d. a large area of slight convergence has cyclonically curved contours.

8. (207) A very *narrow* band of clouds associated with a cold front may indicate the
a. 700-millibars (mb) contours are parallel to the surface cold front.
b. 700mb isotherms are parallel to the surface cold front.
c. 700mb contours are perpendicular to the surface cold front.
d. 1000-500mb thickness lines are parallel to the surface cold front.

9. (208) The type and intensity of precipitation observed at the earth's surface is related to the thickness of the cloud aloft, and particularly to the
a. height of the cloud deck top.
b. height of the cloud deck base.
c. temperatures in the upper part of the cloud.
d. temperatures in the lower part of the cloud.

10. (209) With a quasistationary front in the southern United States under a broad west or southwest flow aloft and a weak surface low, the snow rain zone becomes
a. truncated in the direction of the upper-level current with precipitation rates stretching over long periods.
b. elongated in the direction of the upper-level current with precipitation rates stretching over long periods.
c. truncated in the direction of the upper-level current with precipitation rates stretching over short periods.
d. elongated in the direction of the upper-level current with precipitation rates stretching over short periods.

11. (210) Which condition increases the probability of fog formation?
a. Advecting a warm, moist air mass over a warm surface.
b. Advecting a warm, moist air mass over a cold surface.
c. Turbulent mixing of a moist layer with warmer air aloft.
d. Turbulent mixing of a moist layer with an adjacent dry layer.

12. (210) Which condition decreases the likelihood of fog formation?
a. Constant temperature and increased moisture content.
b. Constant moisture content and decreased temperature.
c. Increased temperature and decreased moisture content.
d. Decreased temperature and increased moisture content.

13. (211) What type of air mass causes a persistent type of continental high-inversion fog to occur in valleys?
a. Maritime polar air.
b. Maritime tropical air.
c. Continental polar air.
d. Continental tropical air.

14. (212) Which visibility restriction is the *most* localized?
a. Haze.
b. Smoke.
c. Precipitation.
d. Wind-blown particles.

15. (213) What would the estimated low-level wind speed be, in knots, from satellite imagery displaying *arc-shaped* open-cell cumulus clouds?
a. Less than 10.
b. 11 to 20.
c. 21 to 30.
d. Greater than 30.

16. (214) If no inversion is present, forecast the *maximum* surface wind gusts to be
a. 50 percent of the 2,000-foot wind speed.
b. 80 percent of the 2,000-foot wind speed.
c. 50 percent of the 5,000-foot wind speed.
d. 80 percent of the 5,000-foot wind speed.

17. (214) In forecasting frontal winds, the deepening or filling of a frontal trough
a. increases the winds.
b. decreases the winds.
c. increases or decreases the winds.
d. does *not* have any effect on the winds.

18. (215) Which tertiary circulation is associated with mountain-wave turbulence?
a. Mountain breeze.
b. Valley breeze.
c. Mistral wind.
d. Foehn wind.

19. (216) Object A has a temperature of 49 degrees (°) Celsius (C), object B has a temperature of 65°C, and object C has a temperature of 88°C. How might thermal energy be transferred between the three objects?
a. B transfers to A and C.
b. A transfers to B and C.
c. B and C transfer to A.
d. A and B transfer to C.

20. (216) How many degrees are there between the freezing point and the boiling point on the Kelvin temperature scale?
a. 273.
b. 212.
c. 180.
d. 100.

21. (217) What type humidity is another expression of mixing ratio?
a. Mixing.
b. Specific.
c. Relative.
d. Absolute.

22. (218) Which forecasting tool provides a summary of monthly and annual climatic data for a station?
a. Operational Climatic Data Summary (OCDS).
b. Weather research forecast (WRF) meteograms.
c. Modeled curves (MODCURVES).
d. Model output statistics (MOS).

23. (219) Surface heating occurs more rapidly
a. with a stable lapse rate.
b. with strong surface winds.
c. with an unstable lapse rate.
d. when an afternoon thunderstorm occurs.

24. (220) A heat wave over the Midwestern and eastern part of the United States develops when a
a. cut-off low forms off the California coast.
b. blocking high is located over the central United States.
c. long-wave ridge stagnates over the Rockies and a long-wave trough lays over the east coast.
d. long-wave trough stagnates over the Rockies and a long-wave ridge lays over the east coast.

25. (221) Which condition *must* exist for the development of a cold wave over the United States?
a. Northeasterly flow over the eastern Pacific.
b. Large pressure tendencies ahead of the cold front.
c. Movement of a low eastward from the Continental Divide.
d. Continental polar air with temperatures above average over east central Canada.

26. (222) How many hours can the global forecast system (GFS) long range forecast predict out to?
a. 120.
b. 240.
c. 312.
d. 384.

27. (223) If the temperature through a vertical column is warmer than standard, then the air density is
a. greater than standard and a pilot must fly higher to reach a given indicated altitude.
b. less than standard and a pilot must fly higher to reach a given indicated altitude.
c. greater than standard and a pilot must fly lower to reach a given indicated altitude.
d. less than standard and a pilot must fly lower to reach a given indicated altitude.

28. (224) How is the difference (D)-value calculated?
a. True altitude minus standard altitude.
b. Standard altitude minus true altitude.
c. True temperature minus standard temperature.
d. Standard temperature minus true temperature.

29. (225) What values do you need to compute the pressure altitude?
a. Difference value and runway length.
b. Field elevation and altimeter setting.
c. Station pressure and density altitude.
d. Sea-level pressure and free-air temperature.

30. (226) Density altitude is the
a. most frequently used method for forecasting sea-level pressure.
b. altitude in the standard atmosphere at which a given pressure occurs.
c. altitude at which a given density is found in the standard atmosphere.
d. difference between the true altitude of a pressure surface and its standard atmosphere altitude.

31. (226) What two parameters are used to compute density altitude?
a. Altitude and pressure.
b. Pressure and virtual temperature.
c. Wet-bulb temperature and altitude.
d. Density and dew-point temperature.

Please read the unit menu for unit 2 and continue ➔

Unit 2. Forecast Flight Weather Elements

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A MAJOR CONCERN that you have as a weather journeyman is the safety of aircraft and their crews. You must make sure that the pilots have all the information necessary to make the correct decisions. The airborne aircraft has few options and very little time when it encounters unsuspected meteorological hazards. It’s the responsibility of the weather forecaster to alert the aircrew to the existence of these hazards.

This unit covers some flight forecasting problems and concentrates on turbulence, icing, and low-level wind shear forecasting.

2–1. Turbulence

You studied turbulence formation and dynamics in forecast apprentice school. This section is devoted to expanding your knowledge on forecasting turbulence. There has been a great deal of material written about turbulence and its effect on aircraft. Turbulence has led to many aircraft accidents, serious injuries, and deaths over the years. It’s your responsibility to alert pilots about potential turbulence so they can be prepared to act.

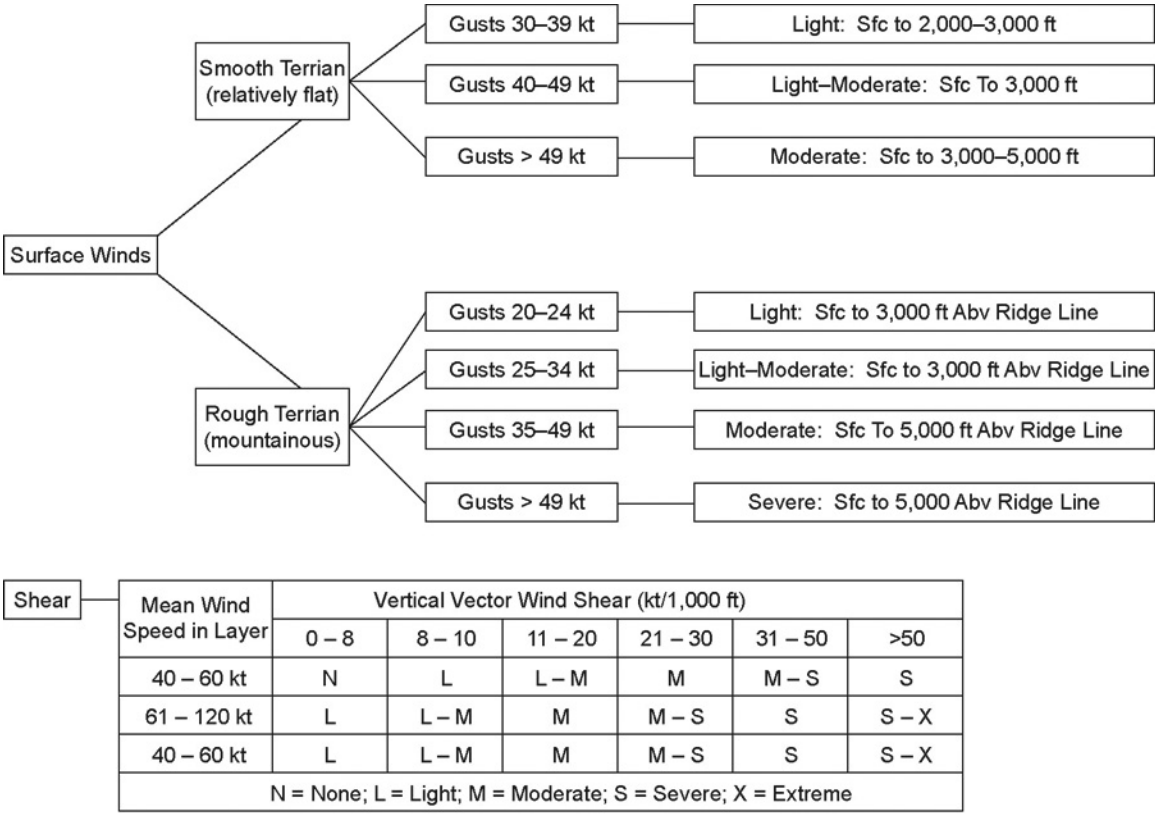
227. Turbulence locations

As with other atmospheric conditions, there are certain meteorological situations and/or atmospheric conditions where turbulence can be expected to occur. Knowing these conditions and situations makes your forecast decision process easier and provides your customer a more accurate forecast. The general location of turbulence should be anticipated in:

- Thunderstorms.
- Areas of strong temperature advection, such as cold-air advection, warm-air advection, strong upper-level fronts, rapid surface cyclogenisis, and outflow area of cold digging jet.
- Areas of considerable horizontal directional and/or speed shear, such as in mountainous areas, developing cutoff lows, and sharp anticyclonic curvature.
- Areas of considerable vertical shear, particularly below strong stable layers in tilted ridges, sharp ridges, tilted troughs, and confluent jet streams.

Basic forecasting checklist for low-level turbulence

Low-level (surface to 10,000 feet) turbulence can dramatically impact flight operations. Aircrews operating in high-speed, low-altitude training routes and those participating in close air support missions must be prepared to make quick corrections to avoid catastrophic accidents. One way to make sure you prepare the flight crew for possible turbulence is to use a low-level checklist (fig. 2–1).



Mean Wind Speed in Layer	Vertical Vector Wind Shear (kt/1,000 ft)					
	0 – 8	8 – 10	11 – 20	21 – 30	31 – 50	>50
40 – 60 kt	N	L	L – M	M	M – S	S
61 – 120 kt	L	L – M	M	M – S	S	S – X
40 – 60 kt	L	L – M	M	M – S	S	S – X
N = None; L = Light; M = Moderate; S = Severe; X = Extreme						

Figure 2–1. Low-level turbulence checklist for Category II aircraft.

Centralized products

Both the Air Force Weather Agency (AFWA) and the National Weather Service produce forecast graphic products to assist you in forecasting turbulence. These charts predict out to 48 hours and are produced for lower- and upper-level turbulence. The low-level turbulence chart forecasts from the the surface up to 18,000 feet. The upper level forecast product is for turbulence above 18,000 feet (figs. 2–2 and 2–3). AFWA also produces graphic visualization products for turbulence based on the WRF model.

The National Oceanic and Atmospheric Administration (NOAA) Aviation Weather Center (AWC) produces several graphic products that forecast areas of turbulence as well. The Graphical Turbulence Product (GTP) uses a color coded, four-dimensional diagnosis and forecast of clear air turbulence potential between 10,000 feet and 45,000 feet MSL.

The turbulence forecast products are just that, a forecast. Sometimes the best-compiled forecast doesn't verify. Pilots believe weather forecasters but they believe other pilots even more! After reviewing the turbulence forecast products check for pilot reports (PIREP) of turbulence in the forecast area. PIREPs are actual airborne observations by aircrews.

AFWA collated PIREPs are available on the JAAWIN and are arranged by geographical region. Always to remember to read the PIREP carefully for the time observed, type of aircraft, flight altitude, and any remarks before applying the information to your forecast.

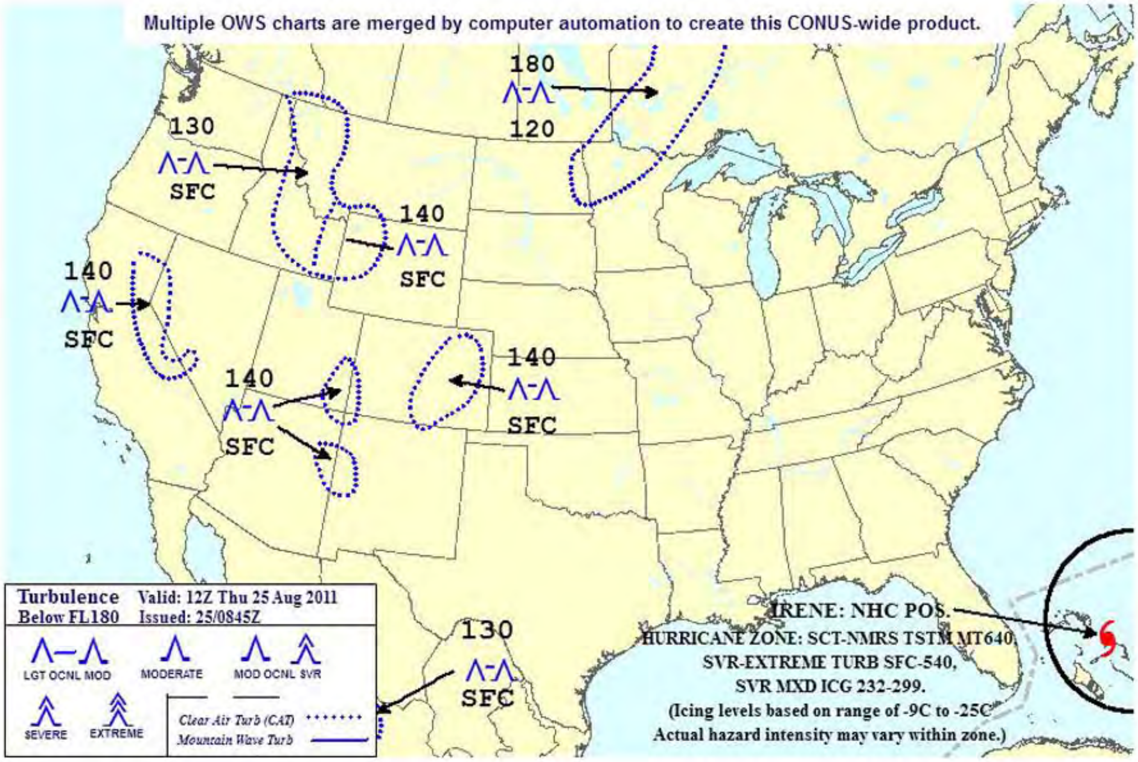


Figure 2–2. Low-level turbulence product.

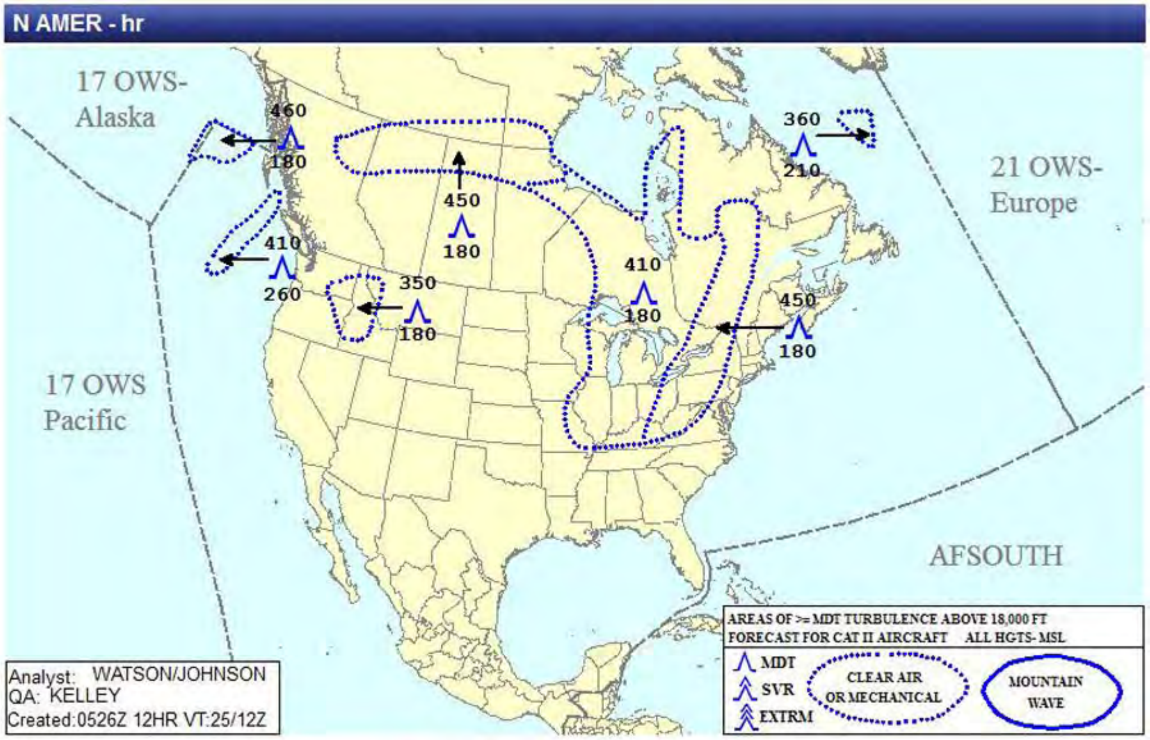


Figure 2–3. Upper-level turbulence forecast product.

Doppler radar

The WSR–88D radar provides a wealth of base and derived velocity products to use in identifying turbulence.

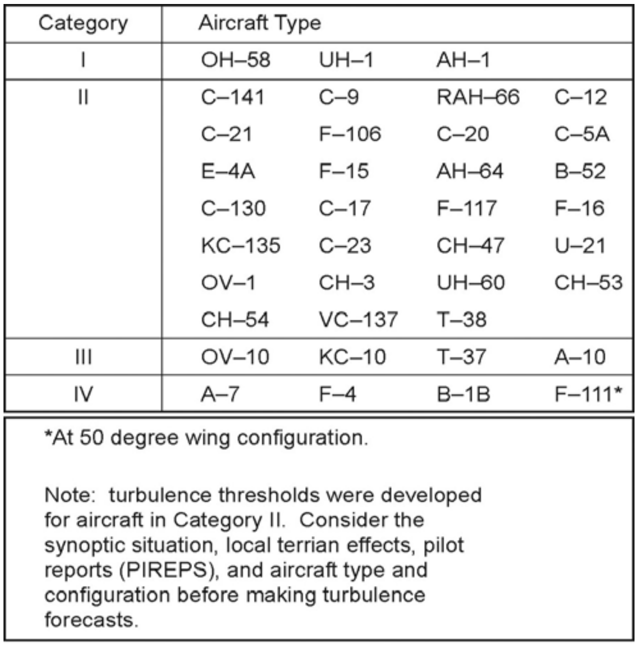
228. Aircraft turbulence sensitivities

In the previous course we discussed the general factors that affect an aircraft’s sensitivity to turbulence. Different aircraft types have different sensitivities to turbulence. The table below lists the categories for most military fixed-wing and rotary-wing aircraft. When you, as a weather journeyman, make a turbulence forecast in a terminal aerodrome forecast (TAF), you are specifying the intensity of turbulence for a Category II aircraft. It is important to know that turbulence forecasts are modified for the specific type of aircraft supported. The table below shows the modifications to the turbulence forecast based on aircraft category.

Legend A	ircraft Category			
N=None	I	II	III	IV
()=Occasional (Less than 1/3 of the time)	Tubulance Reported As			
L=Light	N	N	N	N
M=Moderate	(L)	N	N	N
S=Severe	L	(L)	N	N
X=Extreme	L-(M)	L	(L)	N
NOTE: Use caution when converting extreme turbulence reports between various aircraft types. Extreme turbulence causes a range of effects from a minimum threshold (rapid airspeed fluctuations greater than 25 knots) to a maximum threshold (structural damage). Even though the table considers this, the design is more for the sake of “completeness” rather than observational or scientific evidence.	M	L-(M)	L	(L)
	M-	M	L-(M)	L
	S	M-(S)	M	L-(M)
	S-(X)	S	M-(S)	M
	X	S-(X)	S	M-(S)
	X	X	S-(X)	S
	X	X	X	S-(X)
	X	X	X	X

Aircraft turbulence category type.

An aircraft’s sensitivity varies significantly according to the amount of fuel, cargo, passengers it carries, air density, wing surface area, sweep angle of the wings, airspeed, and flight altitude. So, use caution when applying forecast turbulence (Category II) to a specific aircraft type, configuration, and mission profile. Figure 2–4 is a guide to convert turbulence intensities for different categories of aircraft.



Category	Aircraft Type			
I	OH–58	UH–1	AH–1	
II	C–141	C–9	RAH–66	C–12
	C–21	F–106	C–20	C–5A
	E–4A	F–15	AH–64	B–52
	C–130	C–17	F–117	F–16
	KC–135	C–23	CH–47	U–21
	OV–1	CH–3	UH–60	CH–53
	CH–54	VC–137	T–38	
III	OV–10	KC–10	T–37	A–10
IV	A–7	F–4	B–1B	F–111*

*At 50 degree wing configuration.

Note: turbulence thresholds were developed for aircraft in Category II. Consider the synoptic situation, local terrian effects, pilot reports (PIREPS), and aircraft type and configuration before making turbulence forecasts.

Figure 2–4. Turbulence intensity and aircraft type reference chart.

Turbulence effects on fixed-wing aircraft

As a general rule the effects of turbulence for a fixed-wing aircraft are increased with:

- Increased airspeed.
- Increased wing surface area.
- Decreased weight of the aircraft.
- Increased altitude/decreased air density.
- Non-level flight (climbing or descending).
- Decreased wing sweep angle (wings more perpendicular to fuselage).

Turbulence effects on rotary-wing aircraft

As a general rule the effects of turbulence for a rotary-wing aircraft are increased with:

- Increased airspeed.
- Decreased weight of the aircraft.
- Decreased lift velocity (the faster the lift-off, the less the turbulence.
- Increased arc of the rotor blade (the longer the blade, the greater the turbulence).

As you can see, the impact of turbulence on fixed-wing aircraft versus rotary wing aircraft is quite complex. It’s important that you understand these differences. When disseminating turbulence information to your customer, do so by properly aligning the turbulence intensity with the aircraft category.

229. Distribution of clouds and turbulent regions in a mountain wave

There are specific clouds associated with the mountain wave. They are the cap (Foehn wall), rotor (roll), lenticular, and mother-of-pearl clouds. Figure 2–5 illustrates the structure of a strong mountain wave and the associated cloud patterns. The lines and arrows depict the wind flow from left to right.

Cap cloud

The cap cloud hugs the tops of the mountains and flows down the leeward side with the appearance of a waterfall. This cloud is dangerous because it hides the mountain and has strong downdrafts. The downdrafts can be as strong as 5,000 ft per minute. Essentially, this means that if an aircraft were flying at 20,000 feet above ground level (AGL) it would be at 15,000 feet AGL a minute later due to downdrafts associated with mountain wave turbulence. You can understand how detrimental this can be when aircraft fly within close proximity to mountain peaks.

Roll or rotor cloud

The roll, or rotor, cloud looks like a line of cumulus or altocumulus clouds parallel to the ridge line. It forms on the leeside and has its base near the height of the mountain peak with its top extending considerably above the peak. The tops may extend to twice the height of the highest peak. The rotor cloud often merges with the lenticular clouds above, forming a solid mass to the tropopause.

The rotor cloud is dangerously turbulent, as extreme turbulence should be expected both in and below the clouds. The air in the cloud rotates around a horizontal axis parallel to the mountain range. It has updrafts up to 5,000 ft per minute on its windward edge and downdrafts of 5,000 ft per minute on its leeward edge (fig. 2–6). This cloud is stationary, constantly forming on the trailing edge (updrafts) and dissipating on the leading edge (downdrafts). The roll cloud may form immediately to the lee side of the mountain or it may be a distance of ten miles downwind.

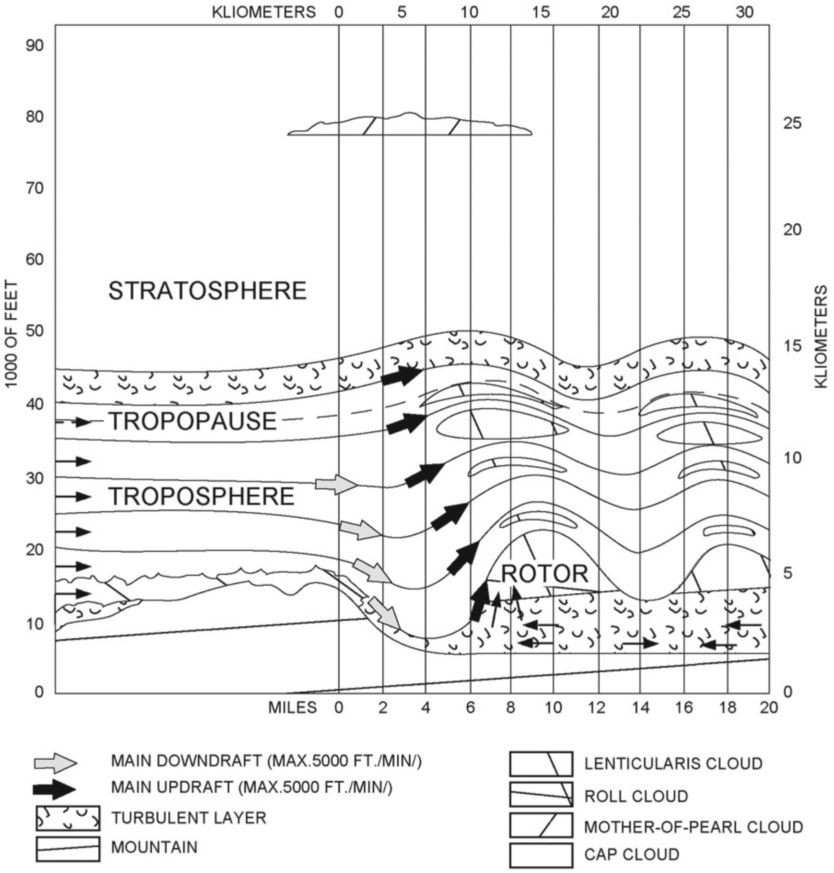


Figure 2–5. Mountain wave structure.

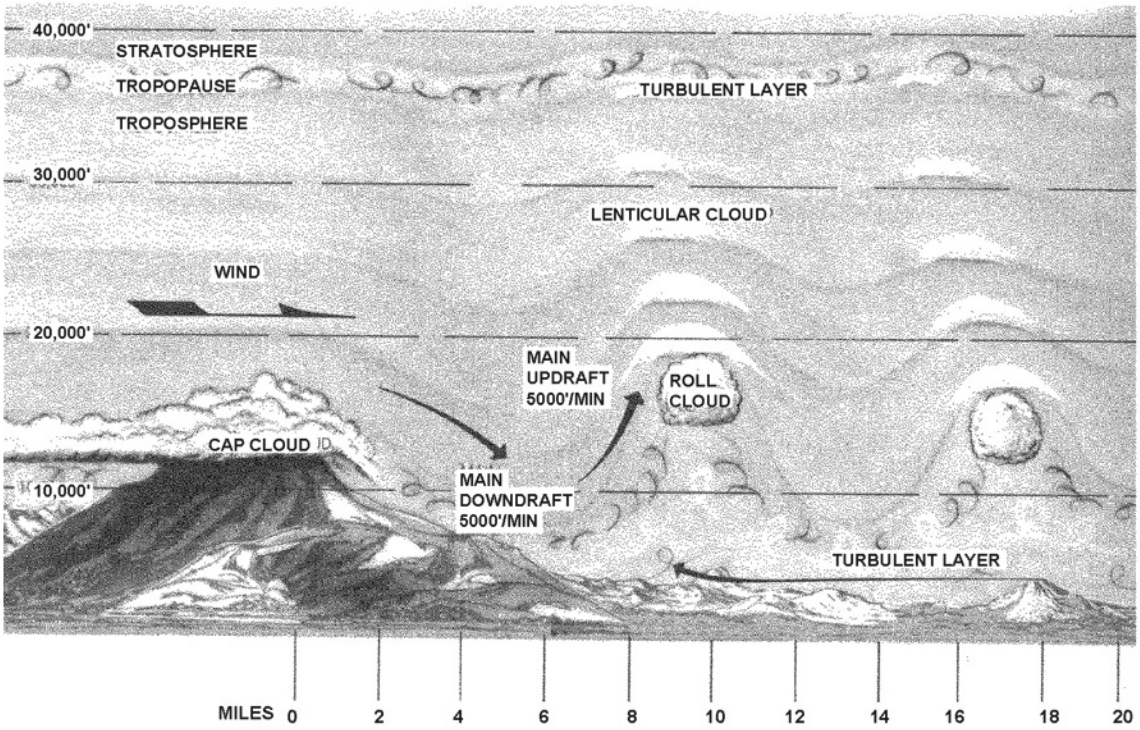


Figure 2–6. Example of mountain wave showing updrafts and downdrafts associated with rotor clouds.

Lenticular cloud

Lenticular clouds form at 80,000 ft and are sometimes called “mother-of-pearl” (nacreous). They are lens-shaped with bases above the roll clouds and their tops extend to the tropopause. These clouds have a tiered or stacked look due to the stratified nature of the moisture in the atmosphere. All lenticular clouds are associated with turbulence. In polar regions, a high stratospheric lenticular cloud is often associated with mountain waves.

Appearance of mountain wave clouds

The clouds in a mountain wave form in pyramids on the lee side of the mountain. These pyramids are usually tilted back toward the mountain. One pyramid may exist, a series of pyramids extending downstream may exist, or no clouds at all may be present. The existence of clouds suggests turbulence; however, *the absence of clouds does not ensure the absence of turbulence*. Turbulence may be present without clouds if there is not enough moisture available to form clouds.

Turbulence associated with mountain wave

The most dangerous features of a mountain wave are the turbulence in rotor and cap clouds. Figure 2–7 shows a cross section of a mountain wave cloud over Boulder, Colorado. The heavy dashed lines separate observations taken at different times; the crosses indicate regions of turbulence. The downdrafts in these clouds can carry a plane into the mountain. Pilots investigating wave clouds relate that they have experienced more hazardous flight conditions in mountain wave clouds than they have encountered in any thunderstorm. Mountain wave clouds have produced gust velocities of 50 feet per second (fps) at 30,000 ft.

Most aircraft experience structural failure when encountering gusts of 50 fps at reduced speeds or 35 fps at ordinary speeds. Calculations have shown that a high-speed jet would experience 8 to 14 times the force of gravity flying downwind through these areas of varying vertical motions.

230. Clear-air turbulence

Clear-air turbulence (CAT) includes all turbulence not thermally induced or associated with convective activity. Thus, mountain wave turbulence is a form of CAT. To get more specific this lesson will discuss turbulence associated with horizontal and vertical wind shear.It has been found that 3 percent of the miles flown in the 20,000 to 40,000 ft. altitudes over land is turbulent (3 miles of turbulent air per 100 miles flown). Approximately 75 percent of this turbulence is light, 15 to 20 percent is moderate (or light to moderate), 5 to 10 percent is severe, and 1 to 3 percent is extreme. The favored region of CAT is from 30,000 to 40,000 ft with 33,000 ft. being the most favorable range. The minimum likelihood of a clear-air turbulence occurring is about at 45,000 ft. CAT is directly related to the heights of the jet stream and the tropopause.

The average thickness of these turbulent “patches” is 2,000 ft. The horizontal width may be from 10 to 40 miles; the length in the direction of the wind averages 50 miles over land and 100 miles over water.

Two-thirds of the occurrences of severe turbulence are associated with jet streams. However, when all turbulence reports are included, only 20 percent are associated with the jet stream. The maximum turbulence occurs below and to the cyclonic side of the jet stream; the minimum turbulence occurs at the level of maximum winds. Over the oceans, the maximum turbulence appears to be to the anticyclonic side of the jet stream.

CAT also occurs near the jet stream, north to northeast of a developing surface low-pressure system (fig. 2–8), and is associated with development of cutoff, upper-level lows (fig. 2–9).

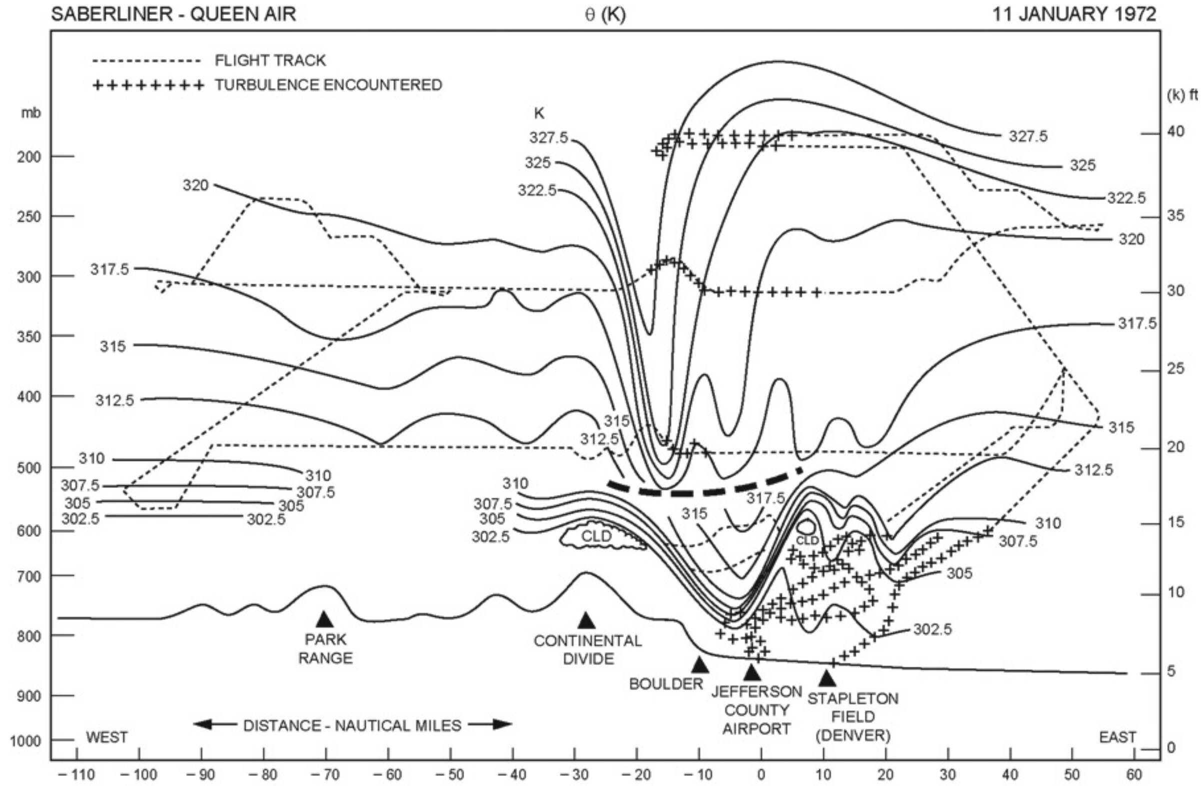


Figure 2–7. Cross-section of the potential temperature field observed in a very strong mountain wave over Boulder, Colorado, on 11 Jan 72.

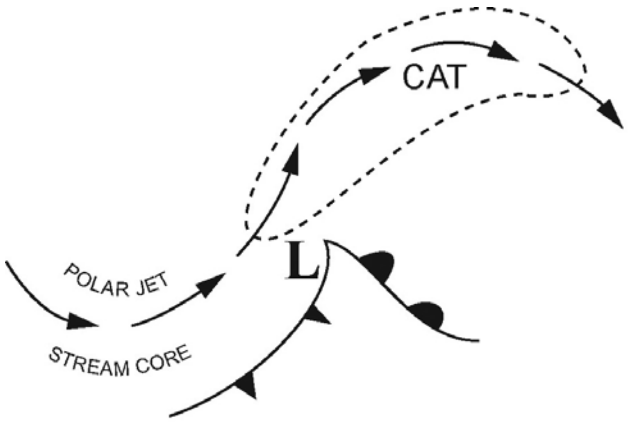


Figure 2–8. CAT N to NE of developing surface low-pressure system.

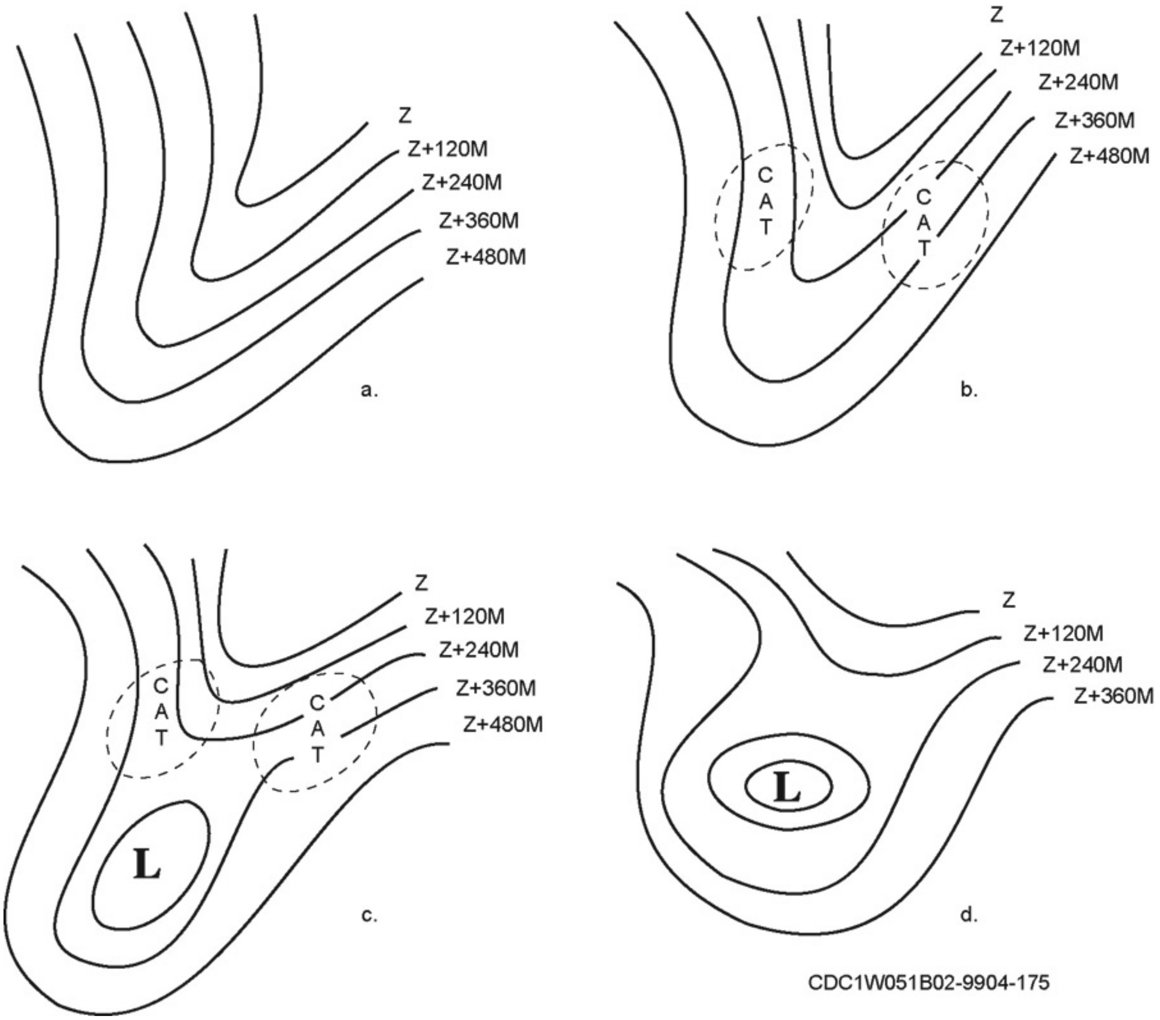


Figure 2–9. CAT development in various stages of development cutoff, upper-level low.

Most CAT is observed during formation of diffluent upper-level wind patterns (fig. 2–10).

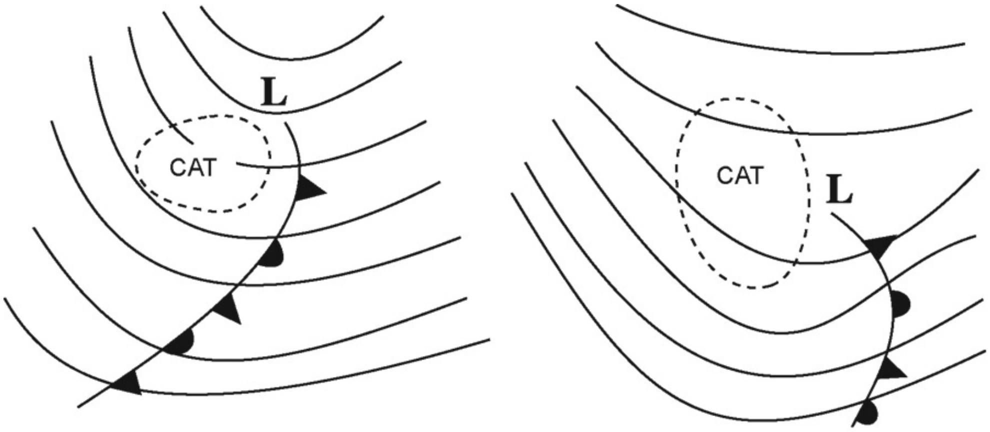


Figure 2–10. CAT associated with diffluent flow aloft.

When two jet stream cores, such as the polar and subtropical jets, converge within 250 nautical miles (nm) or 5°of latitude, the potential for CAT increases (fig. 2–11).

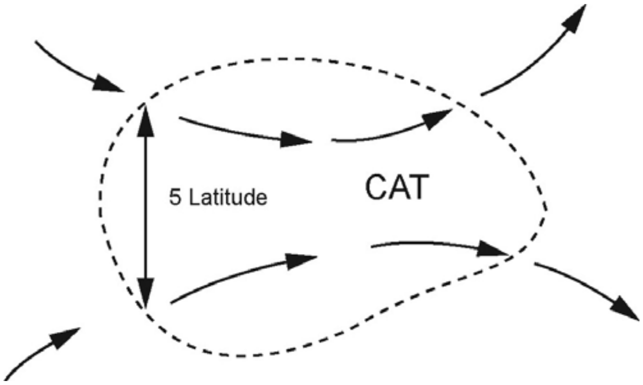


Figure 2–11. CAT associated with converging jet streams.

Normally, your turbulence forecast comes from data obtained from strategic weather centers, such as the Air Force Weather Agency (AFWA) at Offutt AFB. In the event you suspect an area of unforecasted turbulence or begin receiving pilot reports (PIREP) of turbulence; there are several methods available to determine the intensity of CAT. The one method recommended, which is easiest and quickest, is based on wind shear criteria. You should calculate the horizontal and vertical wind shear values. The following tables illustrate the critical values and their associated turbulence intensities:

If the horizontal shear is:	Then the forecasted CAT turbulence is:
25 to 49kt/90nm	Moderate
50 to 89kt/90nm	Severe
>90kt/90nm	Extreme

If the vertical shear is:	The forecasted CAT turbulence is:
3 to 5kt/1,000 ft	Light
6 to 9kt/1,000 ft	Moderate
10 to 14kt/1,000 ft	Severe
> 15kt/1,000 ft	Extreme

A simple conversion can be used in instances where the distance measured is not 90 nautical miles (nm) or 1,000 feet. For example, you calculate that a 55-knot shear (kts) exists over a distance of 120nm in the horizontal. You can now set up a simple proportion, use cross-multiplication, and solve for “X” to compute the shear for 90nm:

$$\frac{55\text{ kts}}{120\text{nm}} = \frac{\text{X kts}}{90\text{nm}}$$

$$55\text{kts} \bullet 90\text{nm} = 120\text{nm} \bullet \text{X}$$

$$4950\text{ kts/nm} = 120\text{nm X}$$

$$\text{X} = 41.25\text{ kts}$$

Therefore, 41.25 knots of horizontal shear exists per 90nm. Referring back to the horizontal shear table, a forecast of moderate intensity CAT would be made.

The criteria in the two tables above are sufficient by themselves to produce the category of turbulence indicated. However, when two of the criteria are present in the same region, a higher intensity should be expected. For example, if moderate horizontal criteria and moderate vertical criteria are present in the same region, severe turbulence should be expected in the region that overlaps.

The significant parameters in the turbulence-producing mechanisms have been summarized in the table below. Some may be detected by synoptic correlation or by direct dynamic measurement, and not all are equally important. The parameters are separated into three headings:

- High-level wind and temperature field.
- Terrain.
- Gravity wave atmospheric.

Heading Ca	**tegory**	**Parameter**	**Turbulance Suitability**	**Remarks**
High-level Wind And Temperature Field	I Mechanism Over Land	Vertical shear (jet-stream vicinity)	XX	Strong double jet
		Cyclonic shear (Cyclonic side of jet) Vicinity of tropopause	X	Approximately under jet stream; difficult to specify type of shear Well below tropopause
		Low static stability (destabilizing differential advection)	XX	Moderate cold-air advection at 300mb but not at 500mb
		Cyclonic curvature and difluence (troughs and exit regions)	X	Trough to west, difficult to determine difluence pattern
	IA Mechanism Over Oceans	Vertical shear (jet-stream vicinity)		
		Anticyclonic shear		
		Anticyclonic curvature		
		Low static stability (destabilizing differential advection)		
		Exit region of isotach maximum		
Terrain	II	Height of ridge (presence of ridge)	X	Not high
		Ridge well-defined, sharp	XX	
		Series of well-spaced ridges	XX	
Gravity Wave Atmospheric	III	Strong low-level winds	XX	About 25 knots
		Low-level winds normal to ridge	XX	Winds 290°, ridge 200 - 20°
		Increasing wind with height (strong winds aloft)	XX	
		Little change of direction with height	XX	Only about 20 - 30°
		Low-level unstable layer (cold-air advection)	XX	Surface to 850mb
		Intermediate stable layer and less stable above	XX	

The heading parameter "High-level wind and temperature field" can be further separated into two categories, depending on which mechanism is operating. The parameters under category I refer to the usual mechanism over land, and those under IA refer to the usual mechanism over oceans. However, category IA can also be operative over land. All headings (I, II, and III) should be considered when forecasting over land, but only IA should be considered when forecasting over oceans. If the parameter is mildly suitable for turbulence, an X or check mark can be entered in the tubulance suitability column, and if strongly suitable, two or more X’s or check marks should be entered. If multiple X’s or check marks were entered, you should anticipate strong or severe turbulence. If only a few X’s or check marks are entered, only light turbulence is likely to occur. Experience allows subjective evaluation and interpretation of the results.

The preceding table shows an actual example of a significant turbulence-causing parameter checklist that indicated the likelihood of severe CAT. As indicated, several aircraft in the region experienced severe turbulence, and one aircraft crashed. The crash was attributed in part to turbulence and in part to structural defects that made the aircraft a likely victim of structural failure under strong stresses.

231. Wake turbulence intensity

Wake turbulence is becoming an increasing problem with the increased use of heavy aircraft. You should be familiar with the cause of wake turbulence and how it forms. Wake vortices are primarily a product of lift; they roll off the wingtips and trail behind the aircraft. Heavy, slow, and clean aircraft produce the strongest vortices.

Every aircraft generates two vortices. They begin when the nose wheel breaks ground and end when the nose wheel is back on the ground during landings. Each vortex forms at the wingtip as air circulates outward, upward, and around the wingtip. The diameter of the vortex core varies with the size and weight of the aircraft. Tests have shown the vortex may be 25 to 50 ft in diameter but the area of turbulence can be much greater. The vortices usually stay fairly close together (about 3/4 of the wingspan) until dissipation.

Other tests have shown that after the vortices are formed aloft, they sink at 500 ft per minute and stabilize about 900 ft below the flight path, where they begin to dissipate. Atmospheric turbulence speeds their dissipation. However, in calm air, the vortices may exist for up to 15 miles behind a cruising aircraft and persist for as long as six minutes.

Ground effect and surface winds alter the low-level vortex characteristics slightly. As the vortex sinks into the ground effect zone, it begins to move laterally at 5 knots. A crosswind decreases the lateral movement of the upwind vortex and increases the movement of the downwind vortex. This could hold one vortex over the runway for an indefinite time or allow one to drift onto a parallel runway.

The major avoidance criteria depend on the aircrew. They must land their aircraft beyond the touchdown point of the aircraft ahead of them. They must lift off ahead of the liftoff point of the aircraft ahead of them. In flight, they should avoid flying behind and below other aircraft. The 1994 accident of a Boeing 737 aircraft on final approach outside Pittsburgh’s International Airport in Pennsylvania underwent close scrutiny and investigation by the National Traffic Safety Board (NTSB). The 737 aircraft was in the traffic pattern and attempted to land only minutes after a previous aircraft had landed at the airport. The exact cause of the accident was not conclusive; however, wake turbulence is believed to be a major contributing factor associated with this aircraft accident.

Understanding the causes and effects of wake turbulence on aircraft in flight is vital and important information that a weather journeyman must be familiar with. By understanding the intricacies associated with this phenomena, weather personnel can play a significant role in determining the length of time that wake turbulence vortices affects an airfield on a specific day.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

227. Turbulence locations

1. Where should turbulence be anticipated?

2. From which agencies are centralized turbulence products available?

3. What source of turbulence reports should you review after checking all forecast products?

228. Aircraft turbulence sensitivities

1. What category of aircraft does a turbulence forecast normally specify?

2. What six factors increase the effects of turbulence for a fixed-wing aircraft?

3. What four factors increase the effects of turbulence for a rotary-wing aircraft?

229. Distribution of clouds and turbulent regions in a mountain wave

1. Draw a figure of the mountain wave indicating the clouds present and relate the clouds to the mountain and to each other.

2. Describe the regions where turbulence should be expected in the mountain wave and the intensity of each region.

230. Clear-air turbulence

1. Define CAT.

2. At what altitude does most CAT occur?

3. How thick is the average layer of CAT?

4. How long is the average length of CAT in the direction of the wind?

5. Where does CAT occur in relationship to the jet stream?

6. Complete the following table below by predicting the intensity of CAT for the conditions listed.

Horizontal shear (knots/90nm)	Vertical shear (knots/1,000 ft)
a. 30kt _____	j. 3kt ______
b. 95kt _____	k. 18kt _____
c. 75kt _____	l. 12kt ______
d. 20kt _____	m. 7kt ______
e. 60kt _____	n. 6kt ______
f. 49kt _____	o. 15kt _____
g. 55kt _____	p. 8kt ______
h. 35kt _____	q. 14kt _____
i. 40kt _____	r. 22kt ______

231. Wake turbulence intensity

1. What causes wake turbulence?

2. How long can surface winds keep a vortex of turbulence over a runway?

3. To avoid wake turbulence, when should an aircraft lift off if it is following another aircraft?

2–2. Aircraft Icing

Ice accumulation on in-flight aircraft has been and still is one of the more treacherous obstacles known to pilots. With the increase in instrument flights, the icing hazard has become more pronounced, and the importance of forecasting icing conditions has increased as well.

In this section we'll cover icing starting with the basics, such as types and intensity, and then move on to the more complex, such as forecast products, synoptic rules, and icing location.

232. Aircraft icing—types and intensity

In-flight icing is the accretion of supercooled liquid water (SLW) on the airframe of aircraft. Accretion is the process of growth or enlargement by a gradual buildup. This SLW can be in the form of cloud droplets, freezing rain, or drizzle. Generally, cloud ice and snow do not adhere to the airframe, and graupel and hail may actually help to remove the accreted ice.

Icing causes adverse affects on the flight characteristics of an aircraft. Icing can increase drag, decrease lift, and cause control problems. The added weight of the accreted ice is generally only a factor to light aircraft.

In this lesson, you'll learn some information about aircraft icing and its intensity that will help you better understand the information on forecast products, synoptic rules, and icing location.

Types of icing

There are three different types of aircraft icing:

- Rime.
- Clear.
- Mixed.

Rime

When SLW drops are small, such as in a light drizzle or moisture in stratified clouds, the liquid remaining on the aircraft after initial impact freezes quickly before the liquid has time to spread out over the surface. The small, frozen drops trap air between them, giving rime ice a rough, milky, opaque appearance—the look of your freezer (if not a frost-free) before you defrost it. Although lighter than clear ice, rime ice has an irregular shape and a surface roughness, both of which reduce aerodynamic efficiency—reduces lift and increases drag. Because it is brittle, rime ice is easier to remove with aircraft in-flight deicing equipment than clear and mixed ice.

Clear

If after initial impact the remaining liquid from the SLW drop flows out over the aircraft's surface gradually freezing forms a smooth sheet of ice. Large drops, as found in rain or in cumuliform clouds, create clear ice. Clear ice (sometimes translucent) is hard and glossy, heavy and difficult to remove. Clear ice, as the accumulation increases, can build up in a horn-like shape on the leading edges of the aircraft, either as a single horn or a double horn; consequently, aircraft in-flight deicing equipment may be incapable of removing such large build-up of clear ice. This horn-like accumulation produces a very large increase in drag and a decrease in lift.

Mixed

When supercooled drops vary in size or are mixed with snow or ice particles, a combination of clear and rime ice can form very rapidly. Ice or snow particles can actually imbed in the clear ice and create highly irregular shapes on the airfoil leading edges of the aircraft.

Regardless of the form the icing takes, the amount of the accumulation is directly proportional to the amount of liquid water in clouds. The worst case scenario involves large water droplets, temperatures close to freezing, and clouds with a high liquid-water content.

Ice intensities

All three different types of icing can be categorized by one of four ice intensity categories. Icing intensity is determined by the rate of accumulation and its impact on deicing and anti-icing equipment. The four intensity categories are:

- Trace.
- Light.
- Moderate.
- Severe.

Trace

A trace of ice is defined as the ice becomes noticible. Rate of accumulation is slightly greater than the rate of sublimation. It is not hazardous even though deicing or anti-icing equipment is not utilized. However, it can become hazardous when encountered for an extended period of time—over an hour.

Light

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

Moderate

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

Severe

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

Now that you have good understanding of the types and intensities of aircraft icing, we’ll move on to explore the hazards that icing imposes on in-flight aircraft.

233. Aircraft icing—impact on aviation and countermeasures

In-flight icing poses a significant threat to aviation. In fact, it is such a significant threat that aircraft icing can result, and has resulted, in aircraft crashes and fatalities. On October 31, 1994 American Eagle Flight 4184 crashed in Roselawn, Indiana following a rapid descent from an uncommanded roll while on autopilot. The airplane was flying in a holding pattern in freezing drizzle and was descending to a newly assigned altitude. Ice accumulated on the wing’s leading edges at a rate so overwhelming the ice protection system failed resulting in an in flight loss of control and subsequent uncommanded roll.

As you can see, aircraft icing is extremely dangerous. Weather journeymen are charged with the duty to provide weather forecasts for icing and other parameters that assist pilots in arriving at their final destination safely. This lesson is designed to give you some background information that will help you better understand your role in the process of forecasting aircraft icing. The lesson is divided into two topics: effects of aircraft icing and counter measures against aircraft icing.

Effects of aircraft icing

In-flight icing is bad news. If the pilot’s flying strategy or anti-icing fails or the deicing equipment is overwhelmed, there are numerous complications that can result. Let’s concentrate on six key effects that can be caused by aircraft icing:

1. Tail stall.
2. Engine stoppage.
3. Collateral damage.
4. Windshield obstruction.

5. Hinders aircraft control.
6. Destroys the airfoil effect.

Tail stall

How quickly a surface collects ice depends in part on its shape. Icing effects on thin, modern wings is more critical than thick, older wings. The tail surfaces of an airplane normally ice much faster than the wing. If the tail's weight becomes too great or its electronic components fail due to ice and airflow disruption, recovery is unlikely at low altitudes.

The tail acts as a horizontal stabilizer. It balances the tendency of the nose to pitch down by generating downward lift on the tail of the aircraft. When the tail stalls, this downward force is lessened or removed, and the nose of the airplane can severely pitch down. One of the worst things about tail stall situations is the ice accumulation cannot seen by the pilot.

Engine stoppage

Ice can also cause engine stoppage by either icing up the carburetor or, in the case of fuel-injected engine, blocking the engine's air source.

Collateral damage

Ice ingested into the engine can cause foreign object damage (FOD) by ruining turbines and other engine components. Also, ice can cause communication antennas to vibrate so severely that they break.

Windshield obstruction

A number of aircraft accidents have occurred when flights have successfully negotiated airfield approach, but the pilot could not see ahead well enough to land through an iced-up windshield.

Hinders aircraft control

Icing also affects aircraft control surfaces such as the ailerons, flaps, and elevators by restricting their movement, not to mention the negative effects on the associated hydraulic systems. Ice accumulation on rotors and propellers can result in extreme vibrations making control of the aircraft very difficult, if not impossible.

Destroys the airfoil effect

Ice destroys the smooth flow of air over the various aircraft surfaces which increases the drag while decreasing the lift. In addition to adding weight to the aircraft, ice blocks the flow of fuel in carburetors and restricts the flow of air in jet intakes. But, the most significant result of ice is that it destroys the efficiency of the airfoil by altering the shape of the aircraft (fig. 2–12).

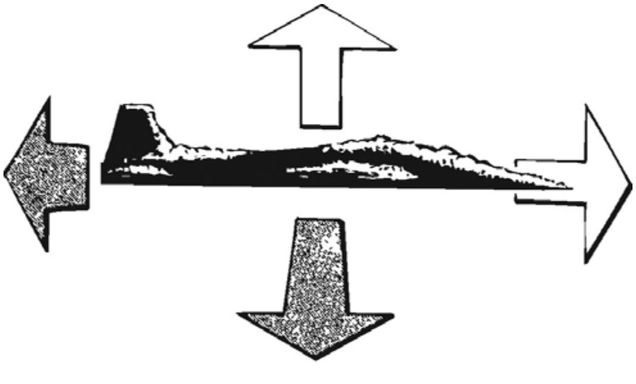


Figure 2–12. Effects of aircraft icing.

Wind tunnel and flight tests have shown that frost, snow, and ice accumulations, no thicker than a piece of coarse sandpaper, on the leading edge or upper surface of the wing can reduce lift by 30 percent and increase drag up to 40 percent. The reduced lift and increased drag can cause a multitude of problems for the pilot.

As power is increased to compensate for the additional drag and the nose is lifted to maintain altitude, the increased flight angle allows the undersides of the wings and fuselage to accumulate additional ice. As this ice builds up on unprotected areas, it increases the overall amount of drag. A study at the National Aeronautics and Space Administration (NASA) revealed that up to 50 percent of the total drag caused by ice accumulation is from unprotected areas of the aircraft. Unprotected surfaces include antennas, flap hinges, control horns, fuselage frontal area, windshield, windshield wipers, wing struts, fixed landing gear, and so forth.

If ice control measures are not taken, the drag and reduced lift caused by an excessive accumulation of ice can produce a situation where flight is no longer possible. In the case of freezing precipitation, the inability of flight can happen in just a few minutes. Freezing rain and drizzle are the ultimate enemy that can distort airfoil shapes. Icing can coat an aircraft so quickly that the time frame for action is extremely short.

Icing counter measures

The counter measures against icing consist of the pilot's flying strategy, anti-icing, and deicing equipment.

Pilot's flying strategy

Smart "ice flying" begins on the ground. This is where you, the weather journeyman, play a vital role in the pilot's formulation of an ice-avoidance flight plan.

It's critical that you inform the pilot of any potential icing areas along their planned flight route. Carefully analyze icing forecast products to determine the impact icing has on their flight path. (These products are covered in the next lesson.) Be on the lookout for any Pilot Reports (PIREP) reporting icing. When icing is reported make sure you brief the aircrew. Your overall objective is to be comprehensive and vigilant; be comprehensive in your review of the weather products to develop an accurate icing forecast and be vigilant in your monitoring of pilot reports for ice encounters.

Anti-icing and deicing equipment

The second line of defense against aircraft icing is anti-icing and deicing equipment. There are times when complete ice avoidance by an aircraft is impossible. In some cases the only surefire way to avoid icing would be to cancel the flight. In situations where aircraft must fly through icing conditions, anti-icing and deicing equipment helps minimize the impact.

Anti-icing is turned on before the aircraft enters the icing condition. Typically this includes carburator heat, prop heat, fuel vent heat, windshield heat, and fluid surface deicers (in some cases).

Deicing is used after the ice has built up to an appreciable amount. Typically this includes surface deicing equipment. There are three major types of wing deicers described in the table below.

Type of Deicer	Description
Boots	Boots are inflatable rubber strips attached to the leading edges of the wing and tail surfaces. The rubber strips inflate to break the accumulated ice off the wing.
Weeping wing system (fluid deicing system)	Pump a deicing fluid from a reservoir through a mesh screen embedded in the leading edges of the wings and tail.
Heated wings	Heating elements located in the wings.

Now that you have a better understanding of what icing is, its dangers to aircraft, and the three types of wing deicers let's move onto some detailed information related to the forecast process for icing.

234. Icing forecast products

When developing your forecast you'll have centralized products available to aid you in deciding if certain atmospheric parameters and conditions will be present and to what extent. Icing is no different from any of the other parameters discussed in this volume in that there are products and tools available from strategic centers and other sources to guide and assist you through the forecasting process. Some of those tools are centralized charts, pilot reports, and Doppler radar.

Centralized charts

AFWA produces high- and low-level hazard charts that contain forecast geographical areas of icing. Annotated by areas are the valid time, height levels, and intensity of the icing expected. AFWA also produces graphic visualization products for icing based on the WRF model.

Also check any AFWA hazard charts for areas of freezing precipitation which suggests moderate (freezing drizzle) or severe (freezing rain) icing.

The Aviation Weather Center’s Airmen’s Meteorological Information (AIRMET) and significant meteorological information (SIGMET) also provide information on any moderate or worse icing.

Pilot reports

Like turbulence the most convincing evidence of icing is a report from another pilot. Use these reports to verify forecasted icing areas and to identify unforecasted areas of icing. Make sure you solicit the pilots, when disseminating aircrew briefings, to report any significant meteorological phenomena whether forecasted or not. Pilot reports are available through the Joint Air Force and Army Weather Information Network (JAAWIN) and through various Internet sites.

Doppler radar

The WSR–88D Doppler radar provides a reflectivity base and derived products that can identify icing areas. One of these products is the composite reflectivity product. Layers of icing are displayed as rings of higher reflectivity values (fig. 2–13).

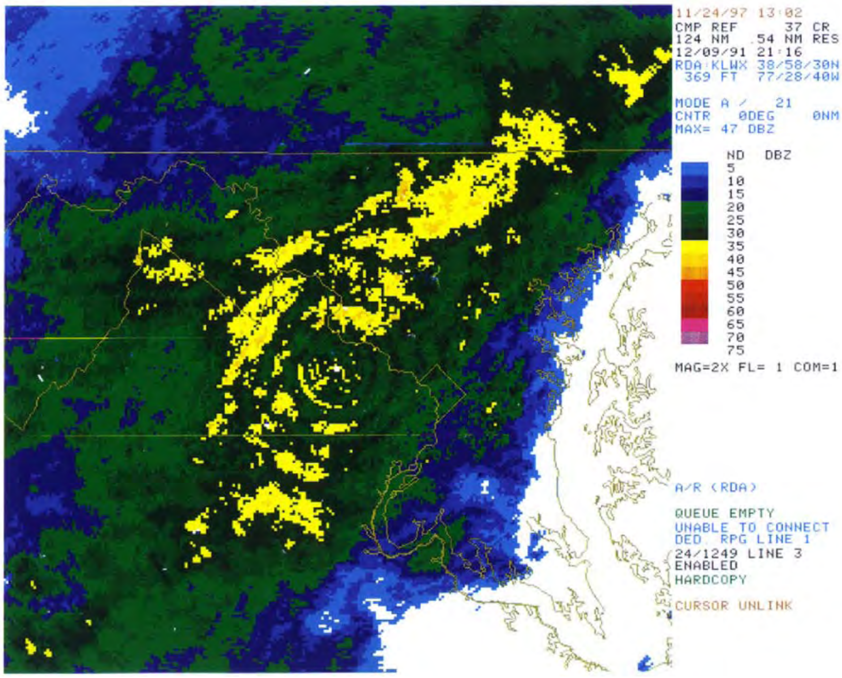


Figure 2–13. Composite reflectivity product showing icing layers.

235. Synoptic rules for forecasting icing

This lesson covers different rules that can be used in ice predictions. Let’s first discuss various icing senarios using upper air data and then determine the forecast of specific icing intensity caused by frontal or orographic lifting.

Icing intensity from upper-air data

The dew point spread and the type of advection occurring provide good indicators of icing potential. Below are rules that have been developed and tested through empirical data.

1. With a temperature of 0° to –7°C and a dew point spread of more than 2°C, there is an 80 percent probability of no icing.
2. With a temperature of –8°C to –15°C and the dew point spread of more than 3°C, forecast no icing with 80 percent probability.
3. With a temperature of –16° to –22°C and a dew point spread greater than 4°C, there is a 90 percent probability of no icing.
4. With temperatures colder than –22°C, there is a 90 percent probability of no icing, despite the dew point spread.
5. If the dew point spread is 2°C or less and a temperature of 0° to –7°C or a dew point spread of 3°C or less and a temperature of –8°C to –15°C:
    a. Forecast trace icing in zones of neutral or weak cold air advection with 75 percent probability.
    b. Forecast light icing in zones of strong cold air advection with 80 percent probability.
    c. Forecast light icing in areas with vigorous cumulus buildup due to surface heating.

Forecasting icing intensity in clouds caused by frontal or orographic lifting

Use figure 2–14 in association with the following rules to forecast icing within clouds caused by frontal or orographic lifting.

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

Type of icing forecast

Use the table below in association with the following rules to forecast the type of icing.

Type of Cloud	**Type of Icing**			
Stratiform	Rime	Rime	Trace Rime	None
Cumuliform	Clear*	Mixed	Rime	None
Temperature ranges for the type of icing				
	0°C to –8°C	–9°C to –15°C	–16°C to –21°C	-22°C and below
NOTE: * Forecast clear icing when flying in freezing precipitation (applies to both freezing rain and freezing drizzle)				

1. Forecast rime icing when the temperatures at flight altitudes are colder than –15°C in any type cloud. For stratiform clouds at temperatures < –15°C, forecast a trace of rime icing if the temperature/dew point spread is ≤ 4°C.
2. Forecast rime icing when the temperatures at flight altitude are –1°C to –15°C in stable, stratiform clouds.
3. Forecast clear icing when temperatures at flight altitude are between 0°C and –8°C in cumuliform clouds.
4. Forecast mixed icing when flight temperatures are between –9°C and –15°C and in unstable clouds.

Two of these rules are illustrated in figure 2–15. Between approximately 9,500 ft and 12,000 ft there are cumuliform clouds with temperature between 0°C and –8°C, so clear ice should be forecast. Above 12,000 ft, a stratiform cloud exists, so rime icing should be forecast.

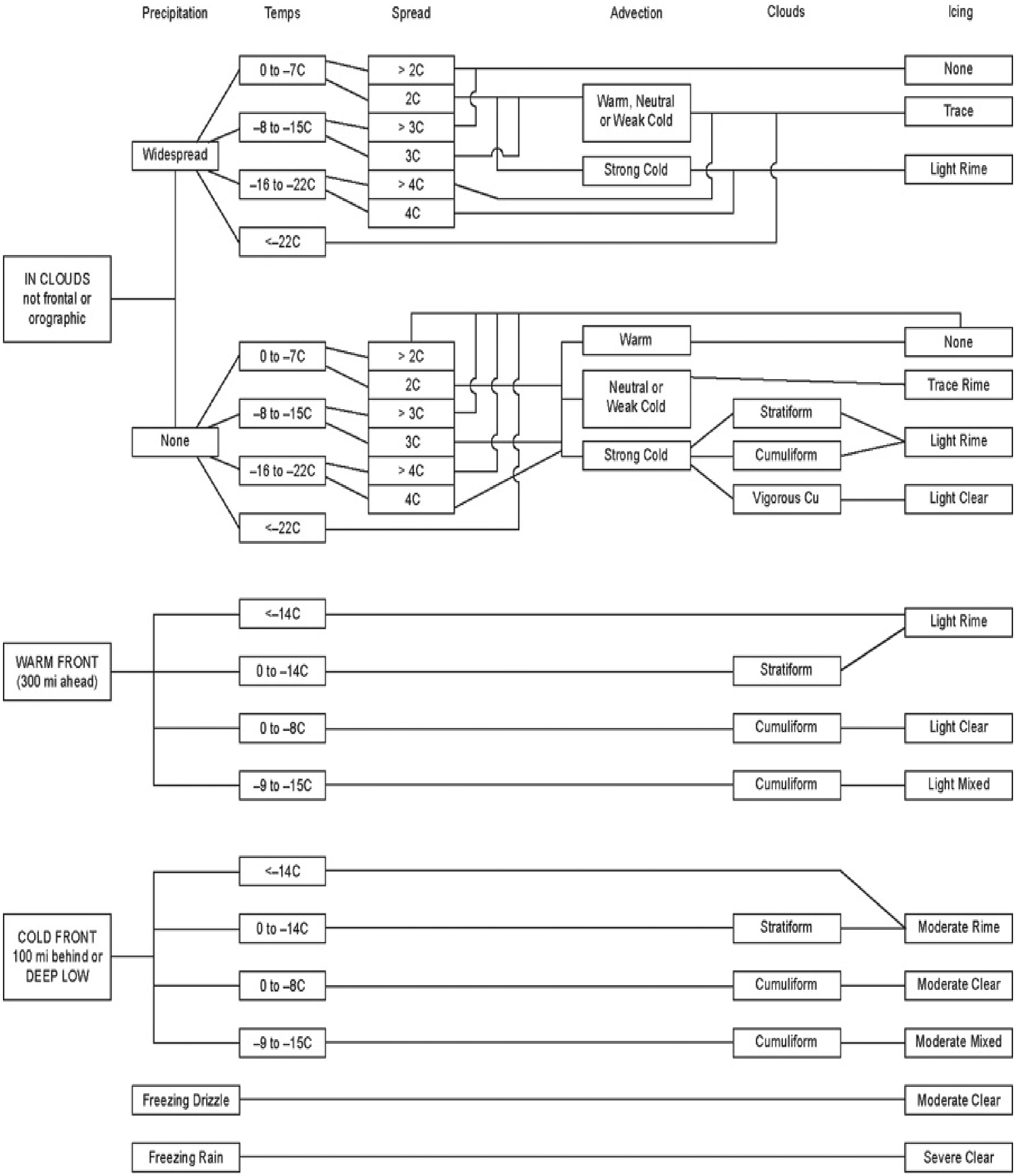


Figure 2–14. Icing forecast flowchart.

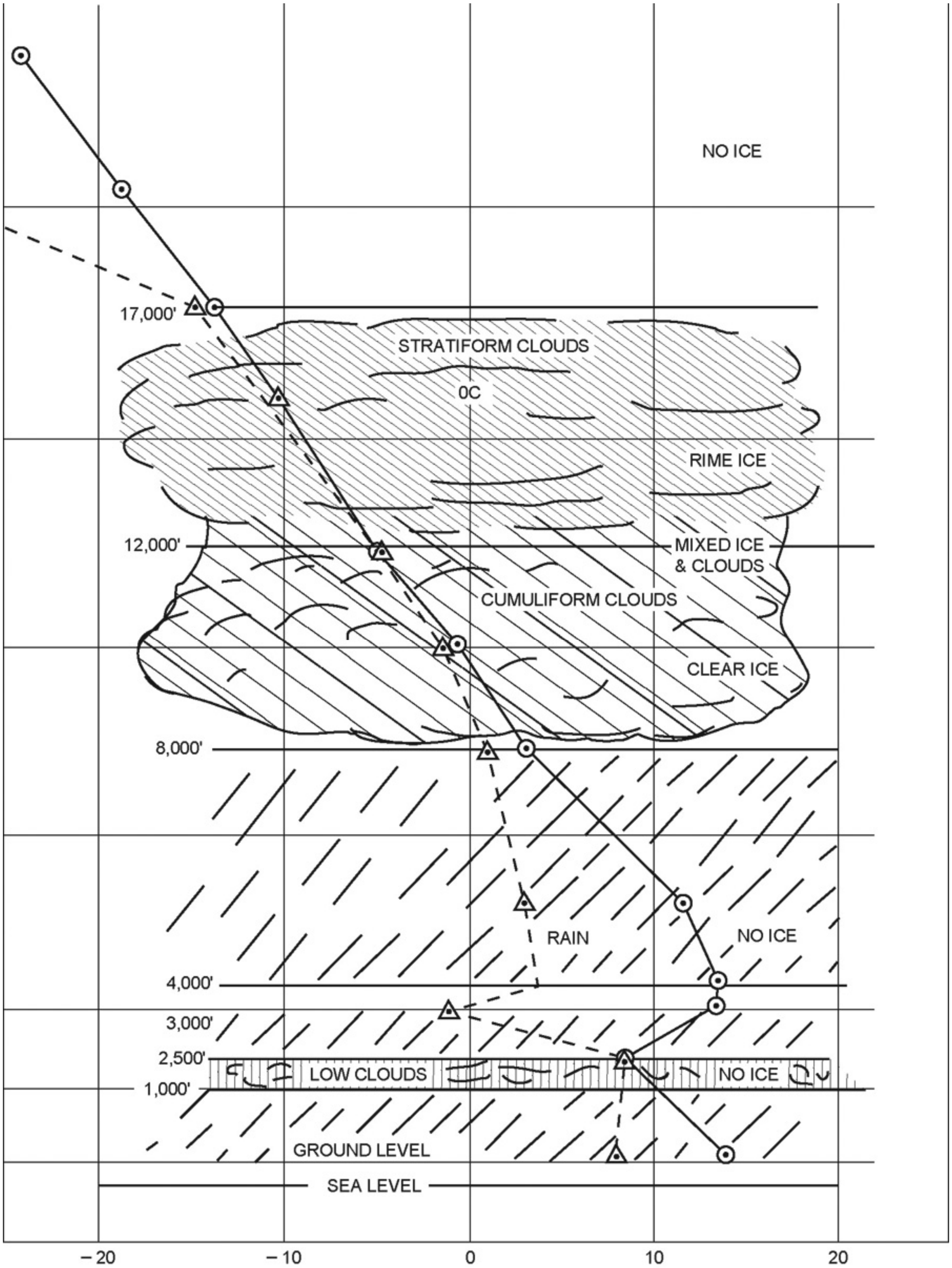


Figure 2–15. Icing indicated on a RAOB sounding.

236. Location of icing in thunderstorms and frontal systems

Figures 2–16, 2–17, and 2–18 illustrate the probable icing areas in the three stages of thunderstorm development. As you can see, significant icing occurs in the developing and mature thunderstorm but little occurs in the dissipating thunderstorm. This happens because the greatest icing areas are associated with strong updrafts and the dissipating thunderstorm has mostly downdrafts.

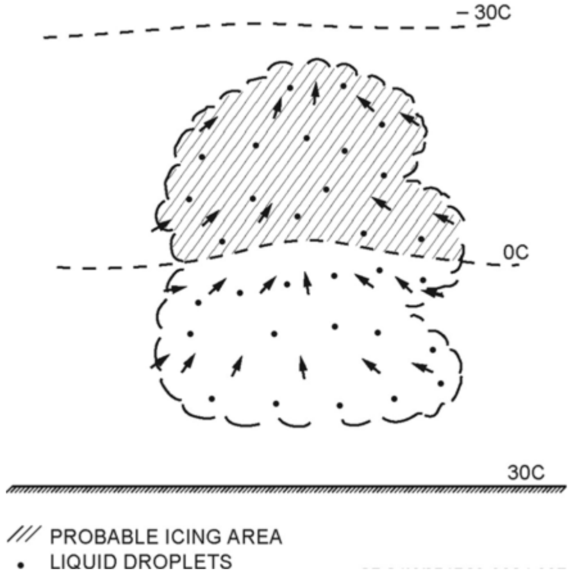


Figure 2–16. Probable icing area in a developing CB cloud.

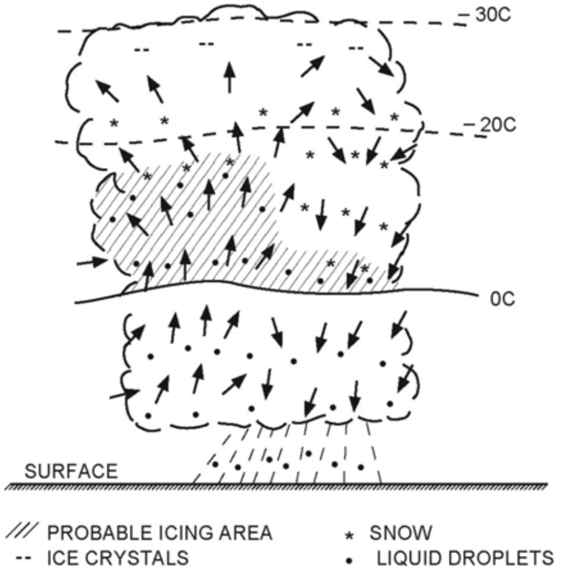


Figure 2–17. Probable icing area in a mature CB.

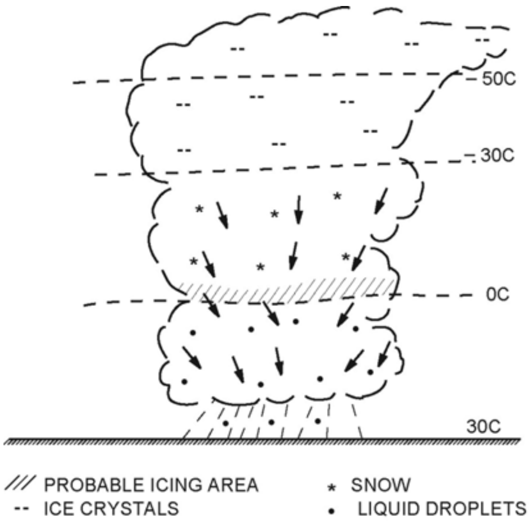


Figure 2–18. Probable icing area in a dissipating CB.

Figures 2–19 and 2–20 illustrate the icing with an active cold front and a warm front, respectively. Again, the greatest icing occurs when updrafts exist. The only exception is the icing associated with freezing precipitation. A horizontal depiction of general icing locations and intensities found in a mature cyclone can be seen in figure 2–21.

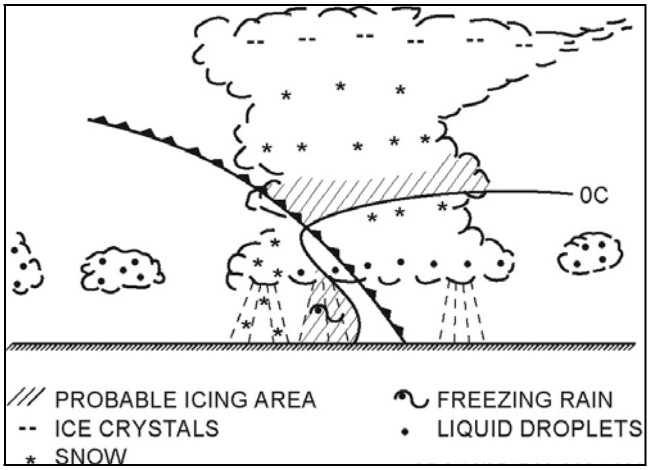


Figure 2–19. Probable icing area along an active cold front.

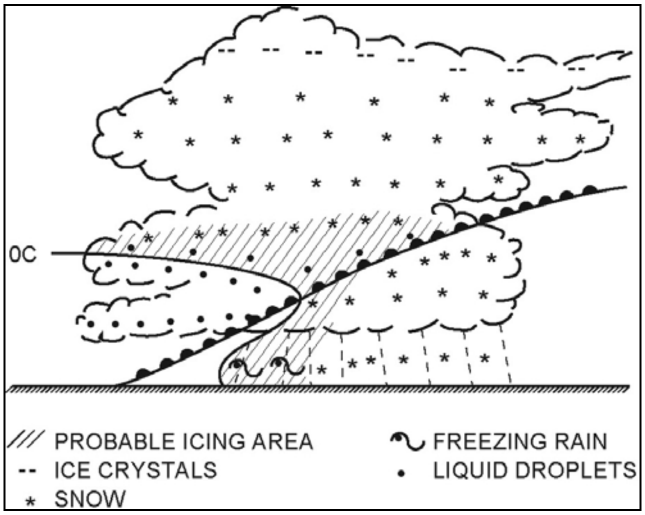


Figure 2–20. Probable icing area along a typical warm front.

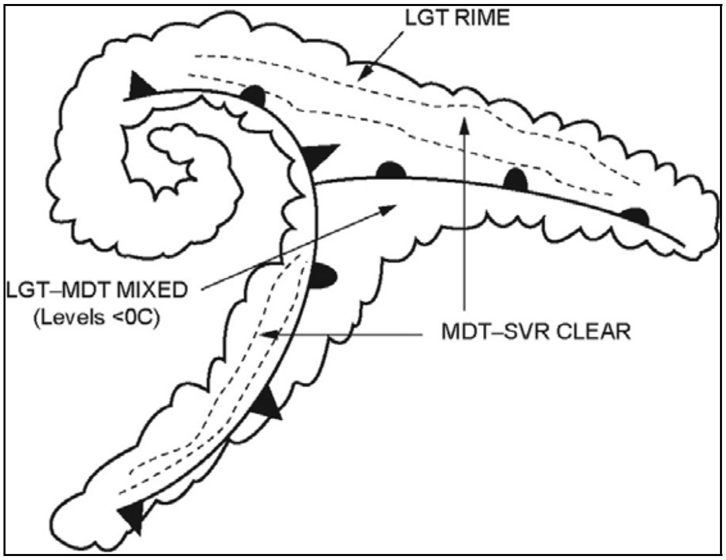


Figure 2–21. Probable icing intensity and type in a mature cyclone.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

232. Aircraft icing—types and intensity

1. What is in-flight icing?

2. Name and describe the three types of aircraft icing.

3. What determines the intensity of in-flight icing?

4. List the four intensities used to categorize in-flight icing.

233. Aircraft icing—impact on aviation and countermeasures

1. What is the greatest impact caused by in-flight icing?

2. Describe tail stall.

3. Describe how icing disrupts the airfoil properties of an aircraft’s wings.

4. In what capacity does the weather journeyman assist pilots in developing an ice-avoidance flight plan?

5. Describe some of the anti-icing and deicing equipment used on aircraft.

234. Icing forecast products

1. What are the two AFWA produced graphic products that forecast icing?

2. What types of icing can the Military Weather Advisory (MWA) alert you to the possibility of?

235. Synoptic rules for forecasting icing

Indicate the icing intensity, probability, and type for the data below:

1. Icing location is 200 miles ahead of a warm front, temperature of –18°C, 5°C point spread, unstable clouds, and neutral advection.
2. Icing location is 50 miles behind a cold front with flight altitude temperature of –9°C, 2°C dew point spread, unstable clouds, and neutral advection.
3. In freezing rain.
4. In unstable clouds with temperature –9°C, 3°C dew point spread, and strong cold advection.

236. Location of icing thunderstorms and frontal systems

1. Where does the greatest icing occur in thunderstorms?
2. Where does the greatest icing occur in frontal zones?

2–3. Low-Level Wind Shear

Today, more emphasis is being placed on forecasting low-level wind shear (LLWS). A review of aircraft accident reports has identified it as a major accident factor. It is likely that many past accidents were caused by this phenomenon but were attributed to pilot error because investigators lacked a more plausible explanation. It is possible to compensate for the effects of LLWS if the pilot is forewarned of its presence. You are responsible for providing this information.

In this section we discuss the effects of wind shear on aircraft performance, the meteorological conditions conducive to LLWS, and some forecasting techniques that enable you to forecast this phenomenon. Many of the rules discussed have been formulated using a rather small database. Therefore, these rules may require modification.

237. Wind shear

Wind shear is the change in the vector wind field in any direction in space, such as along the level flight path or glideslope of an aircraft. Forecasters have been computing wind shear for many years and applying the values obtained to the turbulence forecast. However, the existence of wind shear does not necessarily mean that turbulence will be encountered. Wind shear without turbulence still affects aircraft performance, sometimes with drastic results. This is particularly true during takeoff and landing when the aircraft is within 2,000 ft of the ground. When the wind environment changes faster than the mass of the aircraft can be appropriately accelerated or decelerated, aircraft performance is affected.

Large, abrupt wind shears cause changes in aircraft performance and attitude. An aircraft may gain or lose altitude or deviate from level flight by pitching up or down. The indicated airspeed, a primary indicator of performance, fluctuates. The crew could shorten the recovery time by adding power or pitching down the nose of the aircraft to accelerate to the original indicated airspeed. However, when the aircraft is within a few hundred feet of the runway, these corrective actions may not be possible.

Figure 2–22 illustrates the effect of wind shear on a landing aircraft. The aircraft is on final approach with a tailwind over the outer marker. When the aircraft penetrates the frontal boundary and encounters a headwind, there is an immediate increase in the indicated airspeed, and the aircraft rises above the glideslope. This can cause the aircraft to land farther down the runway and at an indicated airspeed faster than desired, resulting in a long, fast landing. This could leave insufficient stopping distance and the plane could go off the end of the runway.

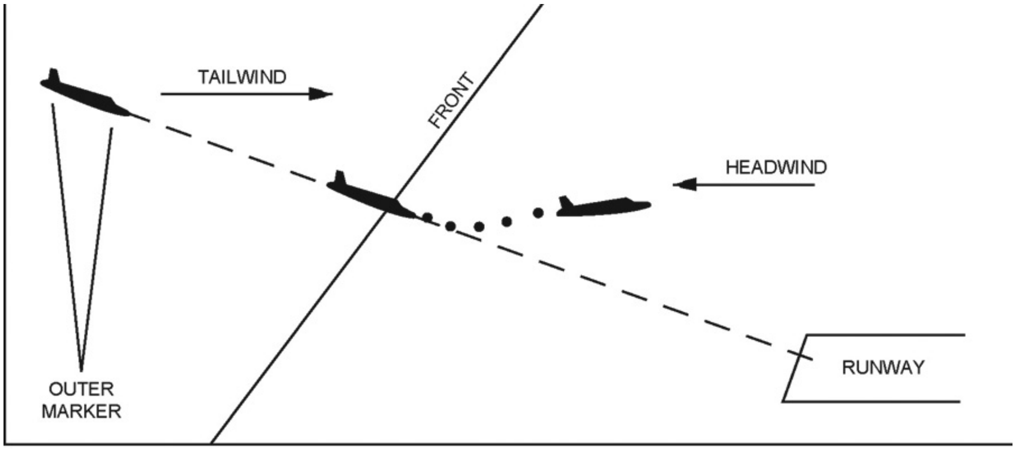


Figure 2–22. Effects of wind shear on an aircraft during approach to a runway as a tailwind changes to a headwind.

When the aircraft begins to deviate from the planned glideslope, the aircrew makes power or configuration adjustments to bring the aircraft back to its proper position on the glideslope. The landing can still be long and fast if the power reduction is insufficient. On the other hand, if the initial power reduction is excessive and not compensated for, the aircraft lands short of the planned touchdown point and at a lower than desired indicated airspeed. This is aptly described as a short, hard landing.

An example of the opposite wind pattern is given in figure 2–23. In this example, the aircraft is flying from a headwind over the outer marker into a tailwind over the runway. Because of the tailwind, the indicated airspeed decreases and the aircraft drops below the planned glideslope unless power or configuration adjustments are made.

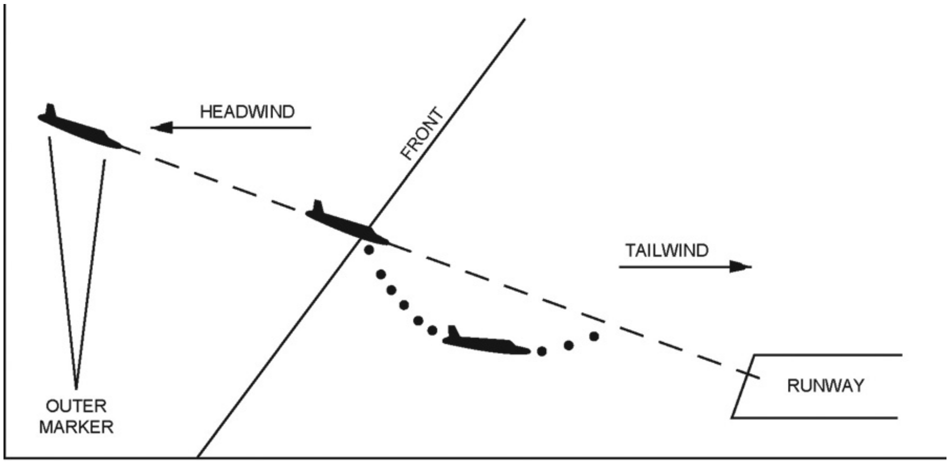


Figure 2–23. Effects of wind shear on an aircraft during approach to a runway as a headwind changes to a tailwind.

To get back on the glideslope, the aircrew is likely to add power. If the added power is insufficient, the aircraft touches down short with a slower than desired indicated airspeed, resulting in a short, hard landing or worse—a crash. When the pilot over-corrects the power adjustment, the aircraft can rise above the glideslope and land beyond the planned touchdown point—resulting in a long, fast landing.

In figures 2–22 and 2–23, the effects of wind shear were illustrated, using frontal zones as a cause of the shear. This, however, is not always the case. As we'll see later, other meteorological conditions can also cause hazardous wind shear.

238. Meteorological conditions favorable for low-level wind shear

Wind shear is a common occurrence at all atmospheric levels, but there are certain meteorological phenomena that can alert us to the potential for LLWS. It is important that the forecaster identify these conditions so that the pilot can be forewarned of the possible hazard.

Thunderstorm gust fronts

One of the most dangerous LLWS conditions occurs with a thunderstorm gust front (fig. 2–24). Cold outflow from thunderstorms forms a mesoscale frontal-type boundary marked by temperature differences and appreciable changes in low-level wind direction and speed. This outflow occurs at the surface in all directions from the thunderstorm cell. The major problem is predicting wind direction and speed with respect to the aerodrome or approach zone.

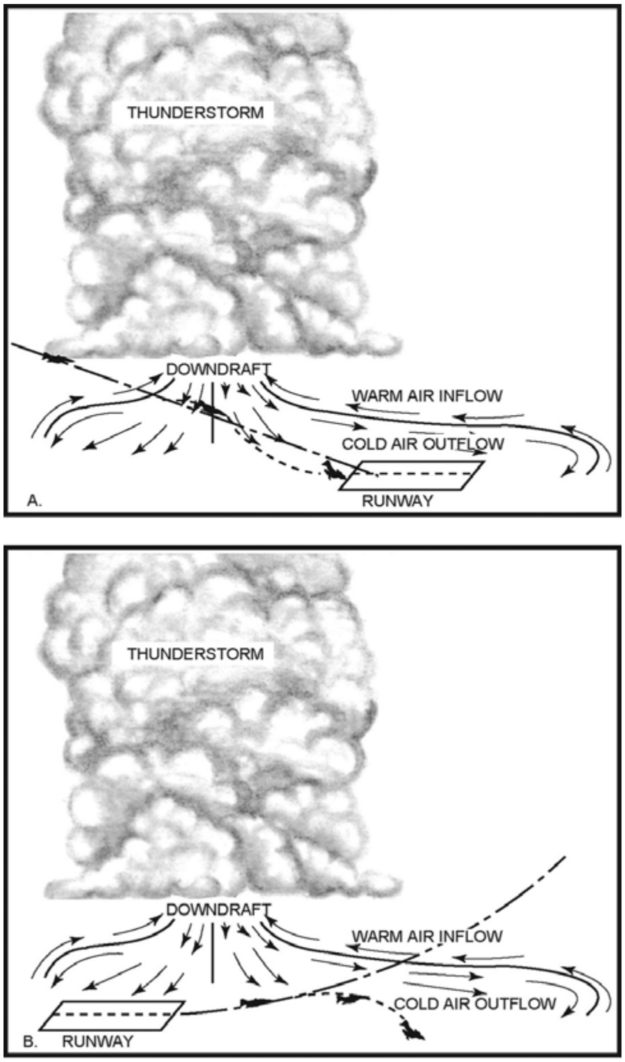


Figure 2–24. Thunderstorm gust front with associated wind shear.

The total operational impact of the low-level shear produced by the thunderstorm gust front depends on the position and movement of the storm with respect to the aerodrome as well as the aircraft flight path. The forecaster should have a good understanding of the following gust front characteristics:

- A surface wind shift may precede the gust front.
- WSR–88D may show the gust front as a thin line echo on radar.
- Upper portions of the gust front may precede the surface portion by one or two miles.
- Pressure jumps usually precede the gust front, and pressure altitude changes near the surface may be 200 to 300 ft.
- The gust front moves faster than the generating thunderstorm and can precede the storm radar echo by five to ten miles.

Since it is difficult to predict the exact onset of the thunderstorm gust front, forecasters must make aircrews aware of the shear threat caused by the gust front. Aircraft may be affected in the approach zone or on the glideslope, yet airfield sensors may not indicate the gust front passage or existence. The forecaster must use sound judgment in forecasting the thunderstorm gust front.

Frontal boundaries

Frontal boundaries are another common cause of potentially hazardous wind shear. The forecaster must examine the vertical structure of the front for shear as it approaches or passes an airfield.

Wind speed and direction changes across a warm frontal boundary can be especially dangerous. It is not unlikely for the shear to be on the magnitude of 90° or more (for example, 90° at 10 knots at the surface veering to 200° at 30 knots above the frontal inversion). When this shear occurs within a few hundred feet of the ground, advisories to aircrews are essential to flying safety. Warm frontal wind shear may persist for six hours or more over a particular airfield because of the small slope and slow movement of the warm front. Another important consideration of warm frontal wind shear is that low ceilings and visibilities frequently associated with warm fronts force more reliance on instruments for safe aircraft operations.

Cold frontal wind shear may be equally dramatic near the surface. Because cold fronts have a greater slope and usually move faster than warm fronts, the duration of LLWS is usually only one to two hours.

Low-level jet

Still another potential wind shear problem for aviation is the nocturnal low-level jet (LLJ). It is observed in all parts of the world, at all times of the year. In the US, it is common in the Great Plains and the central states.

The LLJ occurs above very stable air. The flow is laminar and turbulence is suppressed. Large changes in wind direction and speed can occur in just tens of feet. A typical profile is a light surface wind of less than 10 knots, a surface-based temperature inversion, and a maximum wind of 25 to 40 knots from 600 to 1,500 ft above ground level (AGL). The winds then decrease from 15 to 25 knots up to the gradient level. The core of the jet is just above the top of the inversion layer.

Gusty surface winds

Fluctuations of 10 knots or more from the mean sustained wind speed and strong winds blowing past buildings and structures near a runway produce eddies and turbulence. The resultant shear within a few hundred feet of the surface presents problems for routine aircraft operation. Gusty surface winds frequently accompany some hazards already discussed, namely thunderstorms and fronts. However, other causes must be examined. Chinook winds develop on the lee side of mountains because of simultaneous warm air advection, dynamic heating by subsidence, and rapid destruction of the shallow nighttime inversion. These winds frequently occur along the eastern slope of the Rocky Mountains. Santa Ana winds in southern California and the Foehn wind in the Alps are other examples.

You must also be aware of the topography located within the vicinity of the airfield. High winds can be channeled between peaks and valleys of the nearby terrain. The result is higher wind speeds and gusts that can detrimentally affect the performance of an aircraft on take-off and landings (fig. 2–25). This is especially true when the prevailing wind of the runway is quite different from the gusty winds produced by the nearby terrain.

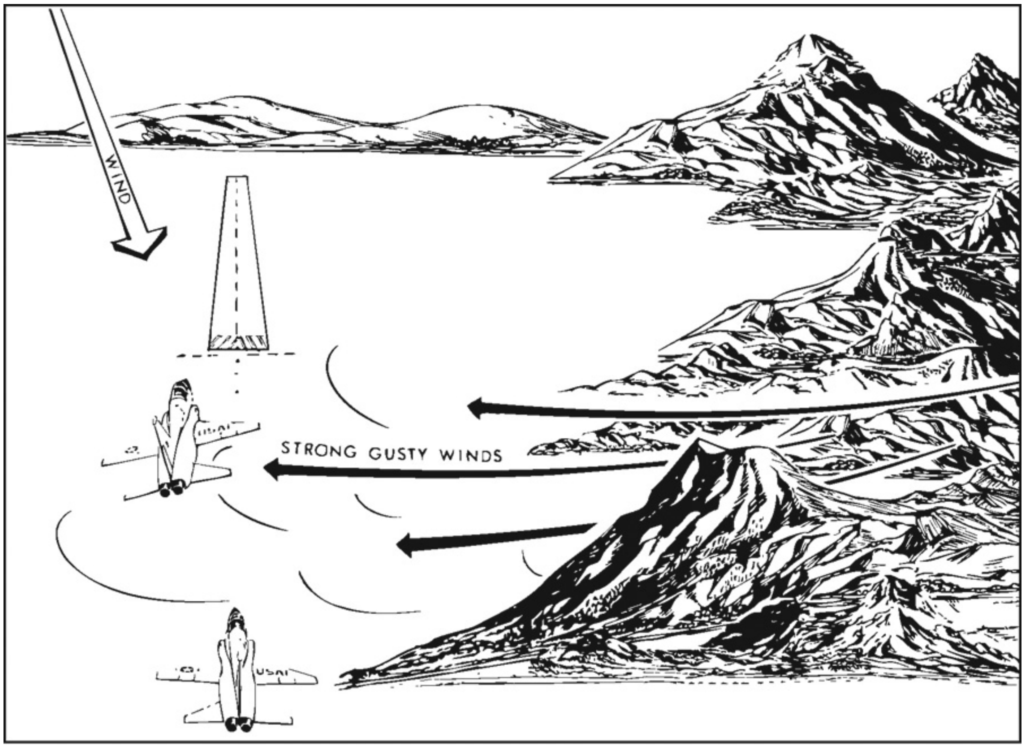


Figure 2–25. Effects of winds being channeled by nearby terrain.

Land and sea breezes

The sea breeze creates a mesoscale frontal boundary that can move inland 10 to 20 miles. It usually reaches its maximum penetration in mid or late afternoon. The depth of the breeze is approximately 2,000 ft. The sea breezes can reach magnitudes of 15 to 20 knots and, occasionally, are observed 75 to 100 miles inland. Land breezes usually occur at night when the land becomes cooler than the water due to radiational cooling. The land breeze has less intensity than the sea breeze and unless aircraft penetrate it on a long, low approach over water, there is little threat to flying safety.

239. Wind shear forecasting rules of thumb

AFWA has modified the rules for forecasting LLWS for use in the LLWS advisory program. The rules may be applied to either forecast or observed conditions. The forecaster should expect LLWS when any of the following conditions are met. Assume the gradient level to be 2,000 ft AGL.

1. When the sustained surface wind is 30 knots or greater.
2. When the sustained surface wind is 10 knots or greater and the difference between the gradient wind speed and twice the surface wind speed is 20 knots or greater.
3. When the sustained surface wind is less than 10 knots and the absolute value of the vector difference between the gradient and surface wind is 30 knots or greater and an inversion or isothermal layer is present below 2,000 ft.
4. When the sustained surface wind is less than 10 knots and the absolute value of the vector difference between the gradient wind and the surface wind is 35 knots or greater.
5. When thunderstorms are observed or forecast within 10nm of the aerodrome.

6. When there is a frontal surface approaching or passing the base with either
    a. A vector wind difference across the front with a magnitude of 20 knots or more per 50nm.
    b. A temperature difference across the front of 10°F (5°C) or more per 50nm.
    c. A frontal speed of 30 knots or more.
7. When a significant LLJ is suspected or reported below 2,000 ft.

The meteorological parameters outlined in these seven rules were observed during real-world instances of LLWS. After meteorological parameters were documented for the first time, they were monitored for future reoccurrence during later instances of LLWS. These seven rules are made up of the most commonly reappearing parameters identified during the study of LLWS. Given the deadly importance of LLWS, it's crucial that you give credence to these rules and learn them as they can save the lives of the aircrew and others.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

237. Wind shear

1. Define "wind shear."

2. How would an aircraft be affected when flying from a tailwind to a headwind during landing?

3. How would an aircraft be affected when flying from a headwind to a tailwind during landing?

238. Meteorological conditions favorable for LLWS

1. Match the items in column A with the items in column B. Some column B items will be used more than once.

Column A

____ (1) Low-level shear may persist for six hours or more due to the small slope and slow movement.
____ (2) Occurs at night and causes very little threat to flying safety.
____ (3) Causes one of the most dangerous LLWS conditions.
____ (4) Wind shear associated with this condition usually lasts for only one to two hours due to large slope.
____ (5) Most intense shear can be expected in the mid to late afternoon.
____ (6) Moves faster than the generating storm.
____ (7) Chinook, Santa Ana, and Foehn are examples of this shear producing phenomenon.
____ (8) Characterized by light surface winds and strong winds just above the surface inversion.

Column B

a. Thunderstorm gust front.
b. Cold frontal boundaries.
c. Warm frontal boundaries.
d. Low-level jet (LLJ).
e. Gusty surface winds.
f. Land breezes.
g. Sea breezes.

239. Wind shear forecasting rules of thumb

In each condition listed below, indicate whether LLWS should be expected.

1. An approaching cold front has a temperature difference across the front of 18°F per 75nm.

2. The sustained surface wind is 12 knots and the gradient wind speed is 30 knots.

3. The sustained surface wind is 8 knots and the absolute value of the vector difference between the gradient wind and the surface wind is 40 knots.

4. You expect a low-level jet to form at 2,500 ft AGL.

5. A thunderstorm is situated 6nm north of your station.

6. A cold front moving at 35 knots has just passed your station.

Answers to Self-Test Questions

227

1. In thunderstorms, areas of strong temperature advection, areas of considerable horizontal directional and/or speed shear, and areas of considerable vertical shear, particularly below strong stable layers.
2. AFWA and NWS Aviation Weather Center.
3. Pilot reports (PIREPS).

228

1. Category II.
2. Non-level flight, increased airspeed, increased wing surface area, increased altitude/decreased air density, and decreased wind sweep angle.
3. Increased airspeed, decreased weight of the aircraft, decreased lift velocity and increased arc of the rotor blade.

229

1. Compare the figure you have drawn with figure 2–5.
2. The most turbulent regions in the mountain wave are in the rotor and cap clouds. The cap cloud has downdrafts of 5,000 ft per minute and the rotor cloud has both updrafts and downdrafts of 5,000 ft per minute. Lenticular clouds are also turbulent.

230

1. CAT is all turbulence not associated with convective activity.
2. 30,000 to 40,000 ft.
3. 2,000 ft.
4. 50 miles over land and 100 miles over water.

5. Below and to the cyclonic side over land but on the anticyclonic side over water.
6. a. Moderate.
   b. Extreme.
   c. Severe.
   d. Light if any.
   e. Severe.
   f. Moderate.
   g. Severe.
   h. Moderate.
   i. Moderate.
   j. Light.
   k. Extreme.
   l. Severe.
   m. Moderate.
   n. Moderate.
   o. Extreme.
   p. Moderate.
   q. Severe.
   r. Extreme.

231

1. It is the result of flow around an aircraft wingtip.
2. Indefinitely.
3. It should lift off before the aircraft ahead of it.

232

1. The accretion of supercooled liquid water (SLW) on the airframe of aircraft.
2. (1) Rime ice – Formed when SLW drops are small, such as in a light drizzle or moisture in stratified clouds, the liquid remaining after initial impact with the aircraft freezes quickly before the liquid has time to spread out over the surface. The small, frozen drops trap air between them, giving rime ice a rough, milky, opaque appearance.

   (2) Clear ice - If after initial impact the remaining liquid from the SLW drop flows out over the aircraft's surface and gradually freezes, it forms a smooth sheet of ice known as clear ice. Large drops, as found in rain or in cumuliform clouds, create clear ice. Clear ice (sometimes translucent) is hard and glossy, heavy and tenacious.

   (3) Mixed ice – A combination of clear and rime ice. When supercooled drops vary in size or are mixed with snow or ice particles, a combination of clear and rime ice can form and form rapidly.
3. Icing intensity is determined by the rate of accumulation and its impact on deicing and anti-icing equipment.
4. Trace, light, moderate, and severe.

233

1. It destroys the airfoil effect.
2. The tail surfaces of an airplane normally ice much faster than the wing. The tail acts as a horizontal stabilizer. It balances the tendency of the nose to pitch down by generating downward lift on the tail of the aircraft. When the tail stalls from ice accumulation, the downward force is lessened or removed, and the nose of the plane can severely pitch down.
3. Icing distorts the shape of airframe surfaces, which destroys the smooth flow of air, increasing drag while decreasing the ability of the airfoil to lift.
4. By informing pilots of any potential icing areas along their planned route of flight.

5. Anti-icing equipment consists of carburetor heat, prop heat, fuel vent heat, windshield heat, and fluid deicers. Deicing equipment consists of three major types: boots, weeping wing systems and heated wings.

234

1. High and low-level hazard charts.
2. Moderate or severe icing.

235

1. 10 percent probability of light rime icing.
2. 75 percent probability of moderate rime icing.
3. 100 percent probability of severe clear icing.
4. 80 percent probability of light mixed icing.

236

1. In the updrafts where the temperature is between 0°C and –20°C.
2. In the updrafts where the temperature is between 0°C and –20°C and in freezing precipitation.

237

1. The change in the vector wind field in any direction in space.
2. The indicated airspeed increases and the aircraft rises above the glideslope.
3. The indicated airspeed decreases and the aircraft drops below the glideslope.

238

1. (1) c.
   (2) f.
   (3) a.
   (4) b.
   (5) g.
   (6) a.
   (7) e.
   (8) d.

239

1. Yes.
2. No.
3. Yes.
4. No.
5. Yes.
6. Yes.

Do the unit review exercises before going to the next unit.

Unit Review Exercises

Note to Student: Consider all choices carefully, select the *best* answer to each question, and *circle* the corresponding letter.

32. (227) Turbulence should *not* be anticipated in areas of
a. inflow in a digging jet.
b. considerable vertical shear.
c. strong temperature advection.
d. considerable horizontal directional and/or speed shear.

33. (228) As a *general* rule, the effects of turbulence for a fixed-wing aircraft are increased with decreased
a. airspeed.
b. wing surface area.
c. weight of the aircraft.
d. altitude and air density.

34. (229) Roll or rotor clouds form
a. parallel to the mountain ridge and are located on the leeside.
b. perpendicular to the mountain ridge and are located on the leeside.
c. parallel to the mountain ridge and are located on the windward side.
d. perpendicular to the mountain ridge and are located on the windward side.

35. (229) The *most* dangerous features of mountain wave clouds are the turbulence in the cap and
a. rotor cloud.
b. nacreous cloud.
c. lenticular cloud.
d. mother-of-pearl cloud.

36. (230) All turbulence not thermally induced or associated with convective activity is classified as
a. severe.
b. clear-air.
c. mechanical.
d. mountain-wave.

37. (231) How can an aircrew avoid the effects of wake turbulence?
a. Land beyond the touchdown point of a leading aircraft, lift off behind the liftoff point of a leading aircraft, and fly above leading aircraft.
b. Land ahead of the touchdown point of a leading aircraft, lift off behind the liftoff point of a leading aircraft, and fly below leading aircraft.
c. Land beyond the touchdown point of a leading aircraft, lift off ahead of the liftoff point of a leading aircraft, and fly above leading aircraft.
d. Land ahead of the touchdown point of a leading aircraft, lift off ahead of the liftoff point of a leading aircraft, and fly below leading aircraft.

38. (232) Which type of aircraft icing occurs in cumuliform clouds and is hard, glossy, and heavy?
a. Rime.
b. Clear.
c. Mixed.
d. Layered.

39. (232) What is the intensity for in-flight icing when deicing or anti-icing equipment fails to reduce or control the hazard and immediate diversion is necessary?
a. Trace.
b. Light.
c. Moderate.
d. Severe.

40. (233) How does in-flight icing on an airplane effect drag and the ability to lift when considering airfoil, respectively?
a. increases; decreases.
b. decreases; increases.
c. increases; increases.
d. decreases; decreases.

41. (233) Which ice protection system is activated *before* the aircraft enters the icing condition?
a. Deicing equipment.
b. Ice removal blowers.
c. Anti-icing equipment.
d. Ionization equipment.

42. (234) What icing intensity corresponds to freezing drizzle?
a. Light.
b. Moderate.
c. Severe.
d. Extreme.

43. (234) Which item is *not* a guidance tool or product for icing?
a. Pilot’s report (PIREP).
b. Meteorological satellite (METSAT) imagery.
c. Airmen’s Meteorological Information (AIRMET).
d. Air Force Weather Agency (AFWA) centralized charts.

44. (235) What type icing, if any, will an aircraft encounter if icing conditions are favorable and flying above 12,000 feet in a stratiform cloud?
a. Mixed.
b. Clear.
c. Rime.
d. None.

45. (236) With a frontal system, the *greatest* icing occurs when updrafts exist. The only exception to this rule is when icing is associated with
a. supersaturated clouds.
b. freezing precipitation.
c. a cold frontal occlusion.
d. a warm frontal occlusion.

46. (237) How would an aircraft be affected during landing if it were flying from a headwind over the outer marker to a tailwind over the runway?
a. Indicated airspeed decreases, and aircraft drops below the glide slope.
b. Indicated airspeed increases, and aircraft drops below the glide slope.
c. Indicated airspeed decreases, and aircraft rises above the glide slope.
d. Indicated airspeed increases, and aircraft rises above the glide slope.

47. (238) How does the upper portion of a thunderstorm gust front compare to the surface portion?
a. Is vertically stacked with the surface portion.
b. Follows the surface portion by one or two miles.
c. Precedes the surface portion by one or two miles.
d. Follows the surface portion by two or three miles.

48. (239) When forecasting weather you would *expect* low-level wind shear when
a. sustained surface wind is 10 knots or greater.
b. thunderstorms are observed or forecast within 20 nautical miles of the aerodrome.
c. sustained surface wind is 10 knots or greater and the difference between the gradient wind speed and twice the surface wind speed is 20 knots or greater.
d. sustained surface wind is less than 10 knots and the absolute value of the vector difference between the gradient wind and the surface wind is 30 knots or greater and an inversion or isothermal layer is present below 2,000 feet.

Please read the unit menu for unit 3 and continue ➔

Student Notes

Unit 3. Convective Severe Weather

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THE MOST DIFFICULT and most demanding duties of a weather journeyman's job are to predict the occurrence and severity of such phenomena as tornadoes, thunderstorms, hail, and damaging winds. These products of atmospheric instability are difficult to exactly forecast because they usually occur within a larger area of potentially severe activity. They are comparatively short-lived, because of rapid development and dissipation, and they occur because of a combination of many atmospheric influences that are often difficult to assess. In this unit, we examine the synoptic conditions and dynamics needed for severe convective weather to form.

Interacting synoptic and mesoscale surface and upper-level weather features such as fronts, troughs, outflow boundaries, and drylines can cause an unstable air mass to erupt into a volatile environment like a spark to gasoline.

Knowing beforehand the conditions needed for severe convective weather to form, helps better prepare you to forecast severe weather.

3–1. Climatology and Formation of Convective Storms

Thunderstorms are one of the most formidable weather hazards for flying operations and are equally significant to ground operations such as troop movements and refueling vehicles or aircraft. Thunderstorms are the result of unstable conditions in the atmosphere. Instability is created when the air is heated in the low levels, cooled aloft, or is forced to ascend the slopes of mountains or frontal surfaces. If rising air is warmer than the surrounding environment in unstable air, convective currents are created. When these currents are strong enough, and when moisture and other factors are favorable, convection occurs.

In this section, we'll cover seven different topics that affect the formation and sustainability of severe convective weather.

- Atmospheric stability.
- United States seasonal variations of severe convective weather.

- Basic conditions necessary for the development of severe convective weather.
- Thunderstorm types.
- Stages of non-severe thunderstorm development.
- The types and development of severe thunderstorms.
- Tornado statistics.

240. Atmospheric stability

Most clouds form as air rises and cools. Why does air rise on some occasions and not on others? And why does the size and shape of clouds vary so much when the air does rise? Let’s see how knowing about the air’s stability helps us answer these questions.

When we speak of atmospheric stability, we are referring to a condition of equilibrium. For example, rock A resting in the depression in figure 3–1 is in stable equilibrium. If the rock is pushed up along either side of the hill and then let go, it quickly returns to its original position. On the other hand, rock B, resting on the top of the hill, is in a state of unstable equilibrium, as a slight push starts it moving away from its original position. Applying these concepts to the atmosphere, we can see that air is in stable equilibrium when, after being lifted or lowered, it tends to return to its original position—it resists upward and downward air motions. Air in unstable equilibrium, when given a little push, moves farther away from its original position—it favors vertical air currents.

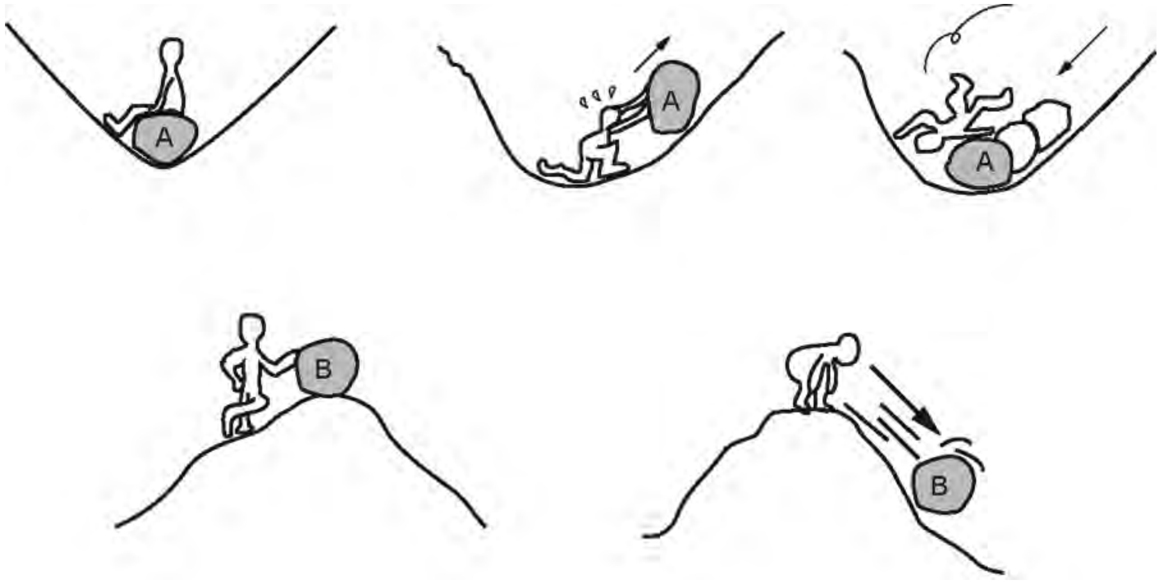


Figure 3–1. Stable and unstable equilibrium

To explore the behavior of rising and sinking air, we must first put some air in an imaginary thin elastic container like a balloon. This small volume of air is referred to as a *parcel* of air**.** Although the air parcel can expand and contract freely, it does not break apart, but remains as a single unit. At the same time, neither external air nor heat can mix with the air inside the parcel. The space occupied by the air molecules within the parcel defines the air density. The average speed of the molecules is directly related to the air temperature, and the molecules colliding against the parcel walls determine the air pressure inside.

At the earth’s surface, the air parcel inside the balloon is the same temperature as the air surrounding it. Suppose we lift the balloon or air parcel up into the atmosphere. We know that air pressure decreases with height. Consequently, the air pressure surrounding the balloon lowers. The lower pressure outside allows the air molecules inside to push the balloon walls outward, expanding the air parcel.

Because there is no other energy source, the air molecules inside the balloon must use some of their own energy to expand. This results in slower than average molecular speeds, which results in a lower air parcel temperature inside the balloon.

If the balloon is lowered to the surface, it returns to a region where the surrounding air pressure is higher. The higher pressure squeezes (compresses) the balloon back into its original (smaller) volume. This squeezing increases the average speed of the air molecules and the air parcel temperature rises. Hence, a rising parcel of air expands and cools, while a sinking parcel is compressed and warms (fig. 3–2).

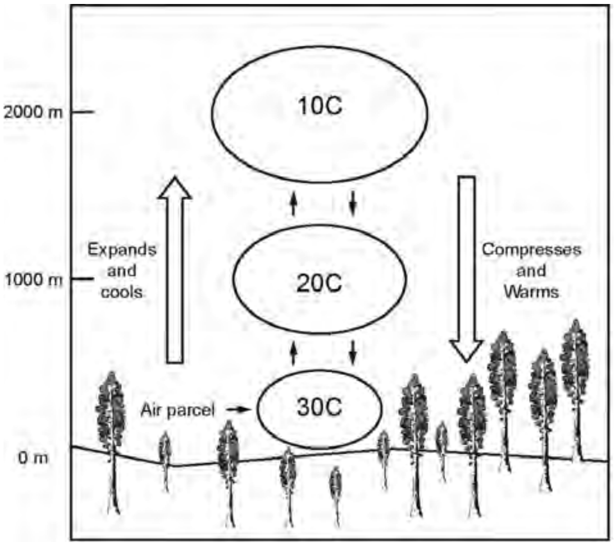


Figure 3–2. Expansion and compression of air parcel.

If a parcel of air expands and cools, or compresses and warms, with no exchange of heat with its surroundings it's called an adiabatic process. As long as the air in the parcel is unsaturated (the relative humidity is less than 100 percent), the rate of adiabatic cooling or warming remains constant. This rate of heating or cooling is about 10° C for every 1,000 meters of change in elevation (5.5°F per 1,000 ft) and applies only to unsaturated air. For this reason, it is called the dry adiabatic rate (fig. 3–3).

As the rising air cools, its relative humidity increases as the air temperature approaches the dew-point temperature. If the air cools to its dew-point temperature, the relative humidity becomes 100 percent. Further lifting results in condensation, a cloud forms, and latent heat is released into the rising air. Because the heat added during condensation offsets some of the cooling due to expansion, the air no longer cools at the dry adiabatic rate but at a lesser rate called the moist adiabatic rate (fig. 3–3). (Because latent heat is added to the rising saturated air, the process is not really adiabatic.) If a saturated air parcel containing water droplets were to sink, it would compress and warm at the moist adiabatic rate because evaporation of the liquid droplets would offset the rate of compressional warming. Hence, the rate at which rising or sinking saturated air changes temperature—the moist adiabatic rate—is less than the dry adiabatic rate.

Unlike the dry adiabatic rate, the moist adiabatic rate is not constant, but varies greatly with temperature. So, warm saturated air produces more liquid water than cold saturated air. The added condensation in warm, saturated air releases more latent heat. Consequently, the moist adiabatic rate is much less than the dry adiabatic rate when the rising air is warm; however, the two rates are nearly the same when the rising air is very cold. Although the moist adiabatic rate does vary, we use an average of 6°C per 1,000 m (3.3°F per 1,000 ft) in most of our examples and calculations.

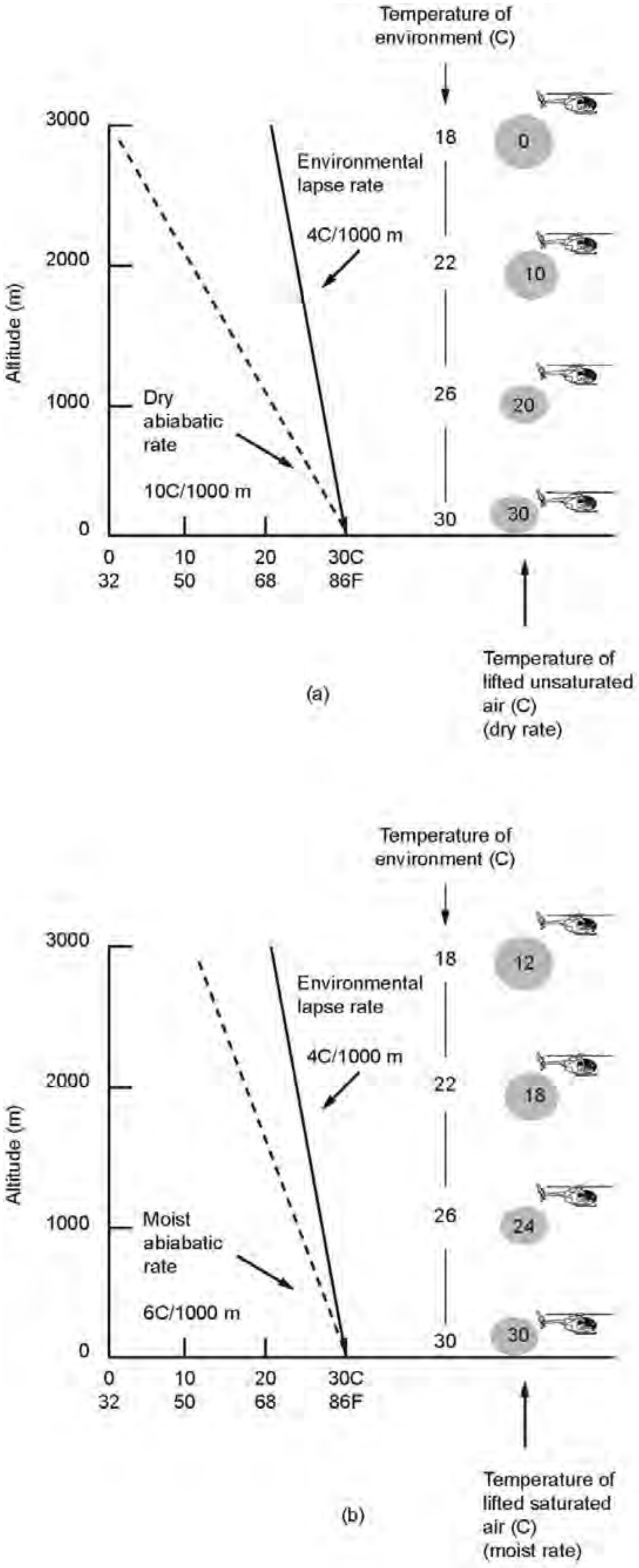


Figure 3–3. Adiabatic lapse rates and stable air parcels.

Determining stability

We determine the stability of the air by comparing the temperature of rising air to that of its surroundings. If the rising air is colder than its environment, it's denser (heavier) and tends to sink back to its original level. In this case, the air is stable because it resists upward displacement. If the rising air is warmer and, therefore, less dense (lighter) than the surrounding air, it continues to rise until it reaches the same temperature as its environment. This is an example of unstable air. To figure out the air's stability, we need to measure the temperature of both the rising air and its environment at various levels above the earth.

Stable air

We can release a balloon-borne instrument called a radiosonde which will send back temperature data. This vertical profile of temperature is called a *sounding*. If we measure the vertical air temperature and find that it decreases by 4°C for every 1,000 meters (2°F per 1,000 ft) it's called the lapse rate; the rate at which the air temperature changes with elevation. Because this is the rate at which the air temperature surrounding us would be changing if we were to climb upward into the atmosphere, we'll refer to it as the environment lapse rate.

Now suppose that a parcel of unsaturated air with a temperature of 30°C is lifted from the surface. As it rises, it cools at the dry adiabatic rate (10°C/1,000 m). At 1,000 meters the temperature inside the parcel is 20°C; 6°C lower than the air surrounding it. Notice that as the air parcel rises higher, the temperature difference between it and the surrounding air becomes even greater. Even if the parcel is initially saturated, it's colder than its environment at all levels. In both cases, the rising air is colder and heavier than the air surrounding it. In this example, the atmosphere is *absolutely stable*. The air is always absolutely stable when the environmental lapse rate is less than the moist adiabatic rate.

Since absolutely stable air strongly resists upward vertical motion, it will, if forced to rise, tend to spread out horizontally. If clouds form in this rising air, they, too, spread horizontally in relatively thin layers and usually have flat tops and bases. We might expect to see clouds such as cirrostratus, altostratus, nimbostratus, or stratus forming in stable air.

A stable atmosphere

How does the atmosphere become stable? As we just discussed, air is stable when the environmental lapse rate is small; that is, when the difference in temperature between the surface air and the air aloft is relatively small. In other words, when you look at a sounding and the temperature curve is more vertical, the atmosphere is more stable. The sharper the temperature curve angles to the upper left corner of a sounding the more unstable the atmosphere is.

The atmosphere can be made more stable by cooling at low levels and/or warming at upper levels. If the air aloft is being replaced by warmer air (warm advection), and the surface air is not changing appreciably, the environmental lapse rate decreases and the air becomes more stable when the lower layer cools. The cooling of the surface air can be caused by nighttime radiational cooling of the surface; an influx of cold air brought in by the wind (cold advection) and by air moving over a cold surface.

Stable atmospheric conditions cause conditions such as fog or haze to persist close to the surface. Another way the atmosphere becomes more stable is when subsidence occurs. For example, if a layer of unsaturated air over 1,000 meters thick that is covering a large area subsides, the entire layer warms up by adiabatic compression. As the layer subsides, it becomes compressed by the weight of the atmosphere and shrinks vertically. The upper part of the layer sinks farther, and hence, warms more than the bottom part as illustrated in figure 3–4. After subsiding, the top of the layer is actually warmer than the bottom, and an inversion is formed.

Inversions that form as by sinking air are called subsidence inversions**.** They sometimes are surface-based but more often are based in the middle and upper levels of the troposphere. They are often associated with large high-pressure areas such as the Bermuda High that dominates the eastern portion of the United States during the summer months. A subsidence inversion, depending on its strength and location in the atmosphere, acts in a manner similar to a cap on a bottle. It limits the vertical transport of moisture and heat (convective currents), and thus hinders thunderstorm development.

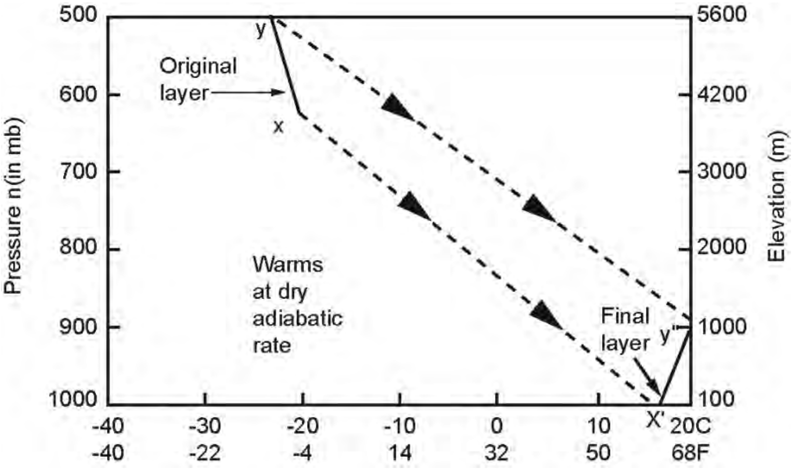


Figure 3–4. Sinking air causing stability.

An inversion represents an atmosphere that is absolutely stable. Within the inversion, warm air overlies cold air, and if air rises into the inversion, it is becoming colder while the air around it is getting warmer. Obviously, the colder air would tend to sink. Inversions, therefore, act as lids on vertical air motion. When an inversion exists near the ground, stratus clouds, fog, haze, and pollutants are all kept close to the surface. In fact, most air pollution alerts occur with subsidence inversions.

Now, let’s examine a condition called *neutral stability***.** If the lapse rate is exactly equal to the dry adiabatic rate, rising or sinking unsaturated air cools or warms at the same rate as the air around it, it is called neutral stability. At each level, it would have the same temperature and density as the surrounding air. Since there is no temperature change between the air parcel and the environment there is no rising or sinking of the air, it is said to be neutrally stable. Neutral stability exists for saturated air when the environmental lapse rate is equal to the moist adiabatic rate.

Unstable air

Suppose a radiosonde sends back temperatures above the earth as shown in figure 3–5. Once again, we determine the air’s stability by comparing the environmental lapse rate to the moist and dry adiabatic rates. In this case, the environmental lapse rate is 11°C per 1,000 m (6°F per 1,000 ft). A rising parcel of unsaturated surface air cools at the dry adiabatic rate. Since the dry adiabatic rate is less than the environmental lapse rate, the parcel is warmer than the surrounding air so it continues to rise vertically, making the atmosphere unstable. Of course, a parcel of saturated air cooling at the lower moist adiabatic rate is even warmer than the air around it. This is shown in figure 3–5, view b.

In both cases, the air parcels, once they start upward, continue to rise just like hot air balloons. This is because they are warmer and less dense than the air around them. The atmosphere in this example is said to be *absolutely unstable***.** Absolute instability results when the environmental lapse rate is greater than the dry adiabatic rate. Typically, absolute instability is limited to a shallow layer of low-level air close to the surface on hot sunny days. In fact, near the ground on hot sunny days, when the environmental lapse rate exceeds about 3.4°C per 1,000 meters (also known as the autoconvective lapse rate), convection becomes automatic.

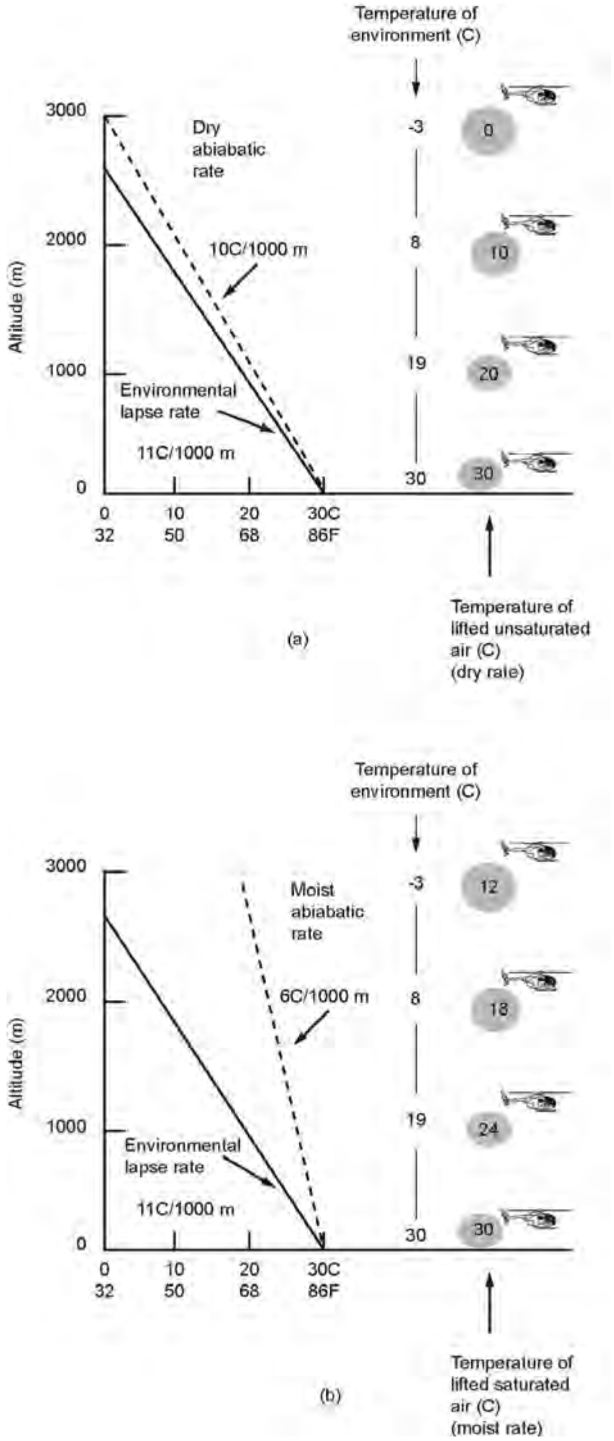


Figure 3–5. Unstable air parcels.

The environmental lapse rate in figure 3–6 is 7°C per 1,000 m (4°F per 1,000 ft). When a parcel of unsaturated air rises, it cools dry adiabatically and is colder at each level than the air around it as in figure 3–6a. Therefore, the parcel tends to sink back to its original level because it is in a stable atmosphere. Now, suppose the rising air is warmer than its environment at each level. This is shown in figure 3–6b.

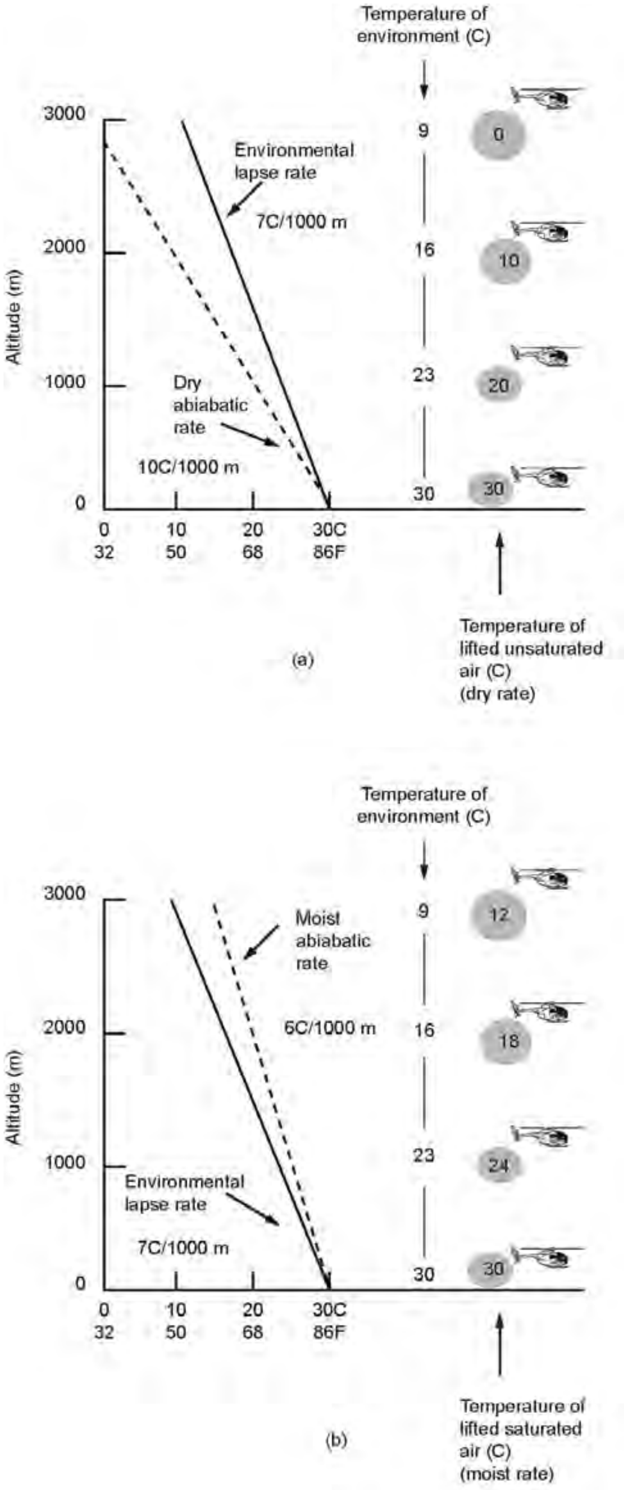


Figure 3–6. Conditionally unstable air.

Once the parcel is given a push upward, it tends to move in that direction; the atmosphere is unstable for the saturated parcel. In this example, the atmosphere is said to be *conditionally unstable***.** This type of stability depends upon whether or not the rising air is saturated. When the rising parcel of air is unsaturated, the atmosphere is stable; when the parcel of air is saturated, the atmosphere is unstable. Conditional instability means that, if unsaturated air could be lifted to a level where it becomes saturated, instability would result.

Conditional instability occurs whenever the environmental lapse rate is between the moist adiabatic rate and the dry adiabatic rate. The average lapse rate in the troposphere is about 6.5°C per 1,000 m (3.6°F per 1,000 ft). Since this value lies between the dry adiabatic rate and the average moist rate, the atmosphere is ordinarily in a state of conditional instability.

Let's review. As shown in figure 3–7, the atmosphere is absolutely stable when the environmental lapse rate is less than the moist adiabatic rate and absolutely unstable when the environmental lapse rate is greater than the dry adiabatic rate. However, a typical type of instability exists when the lapse rate lies between the moist and dry adiabatic rates.

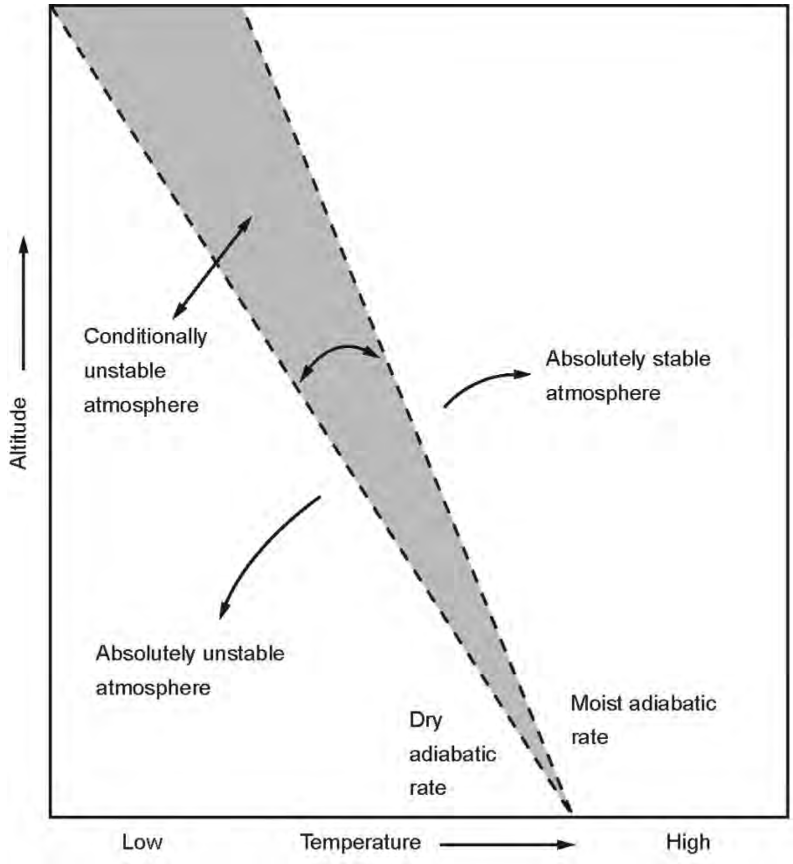


Figure 3–7. Stability identification.

When the atmosphere becomes more unstable, upward vertical motions increase, and cumulus clouds begin to develop. As this development continues, the tops of the clouds increase in height. This action raises two questions: 1) What physical actions cause the atmosphere to become more unstable and 2) What causes upward vertical air currents to rise so that clouds formation takes place? Let's compare unstable air to gasoline. It takes an ignition source to cause gasoline to explode. Likewise unstable air normally needs a "trigger" to start it moving upward through the atmosphere.

Causes of instability

As the atmosphere becomes more unstable, the more the air temperature decreases or drops with increased height. In other words, an environmental lapse rate of 12°C per 1,000 meters represents a more unstable atmosphere than a lapse rate of 6°C per 1,000 meters. Air aloft becoming colder or the surface air becoming warmer may bring on this circumstance. There are three ways that surface air can be warmed:

- Daytime solar heating of the surface.
- Influx of warm air brought in by the wind (warm advection).

- Air moving over a warm surface.

The combination of cold air aloft and warm surface air can produce a sharp lapse rate and an unstable atmosphere.

It’s important to know that the stability of the air changes during the course of a day. During the early morning around sunrise when there are no clouds present and the wind is calm, a radiation inversion normally exists and the surface air is normally colder than the air above it. In this situation, the air is quite stable. Any smoke or haze lingering close to the ground can indicate this.

As the day progresses, sunlight warms the earth’s surface and the surface warms the air above by conduction. As the air temperature rises near the ground, the lapse rate of the lower atmosphere steepens and gradually becomes unstable. Maximum instability usually occurs during the warmest part of the day.

So far, we have explained that a layer of air may become unstable by either cooling the air aloft or warming the air at the surface. A layer of air may also be made more unstable by either atmospheric mixing or mechanical lifting. Atmospheric mixing is caused by such mechanisms as wind currents, turbulence, convection, and advection. In figure 3–8, the environmental lapse rate before atmospheric mixing is less than the moist rate, and the layer is stable as indicated in line A in the figure.

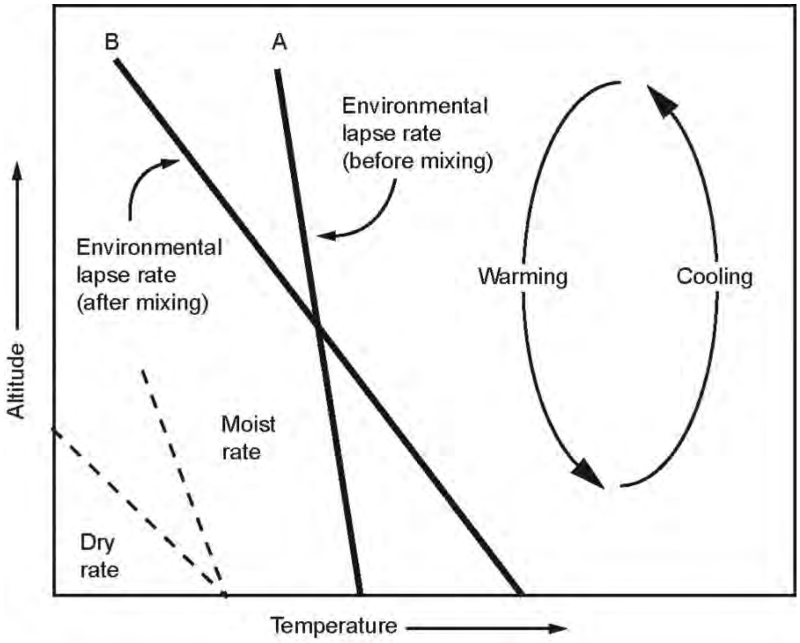


Figure 3–8. Effects of mixing on environmental lapse rate.

Let’s consider an example of mixing. It is a cool, clear morning with a stable atmosphere. As the day progresses the air in the layer is mixed either by convection or by wind-induced turbulent eddies. Air is cooled adiabatically as it is brought up from below and heated adiabatically as it is mixed downward. The up and down motion in the layer redistributes the air in such a way that the temperature in the upper levels of the layer decreases, while in the lower levels the temperature increases. This cooling in the upper levels and warming in the lower levels of the atmosphere steepens the environmental lapse rate and makes the layer more unstable. If this mixing continues for some time, and the air remains unsaturated, the vertical temperature distribution will eventually be equal to the dry adiabatic rate.

Just as sinking air causes a more stable atmosphere, rising air causes a more unstable atmosphere. How the air is lifted doesn’t matter. Suppose on a given day the air lying between 1,000 and 900 mb is initially absolutely stable since the environmental lapse rate of the layer is less than the moist adiabatic rate. The layer is lifted, and as it rises, the rapid decrease in air density aloft causes the layer to stretch out vertically. If the layer remains unsaturated, the entire layer cools at the dry adiabatic rate. Due to the stretching effect, however, the top of the layer cools more than the bottom. This steepens the environmental lapse rate and causes instability.

A very stable air layer can be converted into an absolutely unstable layer if the lower portion of a layer is moist and the upper portion is quite dry. In figure 3–9, the inversion layer between 900 and 850 mb is absolutely stable. Suppose the bottom of the layer is saturated while the air at the top is unsaturated. If the layer is forced to rise, even a little, the upper portion of the layer cools at the dry adiabatic rate and grows cold quite rapidly, while the air near the bottom cools more slowly at the moist adiabatic rate. It does not take much lifting before the upper part of the layer is much colder than the bottom part; the environmental lapse rate steepens and the entire layer becomes absolutely unstable (layer *a’-b’)*. The potential instability, caused by the lifting of a stable layer whose surface is humid and whose top is “dry” is called convective instability**.** Convective instability is associated with the development of severe storms, such as thunderstorms and tornadoes.

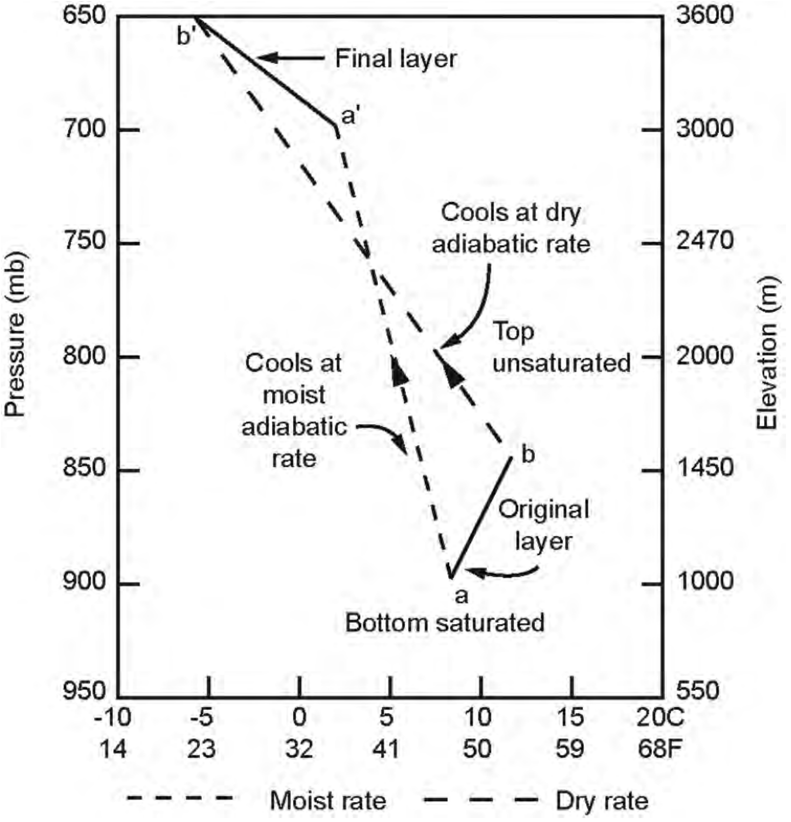


Figure 3–9. Convective instability.

241. United States seasonal variations of convective severe weather

The US is one of the most dominant land areas on the earth in terms of thunderstorm activity. The geography helps to play a large role in this dominance. North America’s large northern land mass combined with the subtropical influence of the Gulf of Mexico and the Atlantic Ocean create an air mass “battle zone” across the US. This is a battle that is fought between the continental polar and maritime tropical air masses. When atmospheric conditions are favorable, the result from these clashing air masses is a thunderstorm. Often, as is the case in the US, these thunderstorms reach severe proportions and wreak havoc on people and property. We (Air Force Weather Agency) can’t stop a thunderstorm, but we can recognize the parameters favorable for convection and give our customers advance notice of the thunderstorm activity.

In this lesson on thunderstorm activity in the US, we'll cover three areas:

- CONUS thunderstorm activity.
- CONUS severe thunderstorm and tornado activity.
- Thunderstorm effects on base resources and aircraft in flight.

CONUS thunderstorm activity

The southeastern United States has the most thunderstorm activity. Such activity is favored during the warmer months (i.e., spring and summer). The primary diurnal activity occurs during the late afternoon and early evening and is associated with maximum surface heating from the sun.

Secondary diurnal activity can occur during the night (nocturnal thunderstorms) during the summer and is associated with cloud remnants of daytime convection. These clouds trap heat in the lower levels. Meanwhile, heat on the top of clouds re-radiates back into space. This creates an unstable condition with cooling of the air in the upper levels above the clouds, as warm air remains trapped in the lower levels. The air contrast causes the warm low-level air to rise and condense, thus forming thunderstorms. The convective activity usually continues until sunrise when the sun heats the upper levels, stabilizes the atmosphere, and dissipates the thunderstorms. However, other dynamic features (i.e., an approaching surface front) can enhance the nocturnal activity and allow it to continue after sunrise.

CONUS severe thunderstorm and tornado activity

Severe thunderstorms are most prevalent from the Continental Divide to the Appalachian Mountains and from the Gulf of Mexico to Canada. However, they can occur anywhere in the US if highly unstable atmospheric conditions and dynamic features exist. The greatest occurrence of severe activity in the US is along "Tornado Alley," located in a region encompassing Texas, Oklahoma, Kansas, and Nebraska. There are other things to consider when discussing thunderstorm and tornado activity such as the time of season, lower- and upper-level dynamics, and finally, occurrences and movement of severe weather.

Seasonal

In the CONUS, March through June appears to be the best months for tornado activity while March through September are the best months for severe thunderstorm activity. However, this severe weather can occur anytime atmospheric conditions are favorable.

Dynamics

Severe thunderstorms and sudden tornadoes develop when strong lower- and upper-level dynamic support exists. Warm, moist, low-level air colliding with cold, dry upper-level air is conducive to creating a volatile situation capable of producing severe weather. These air-mass boundary zones include:

- Cool continental-polar (cPk) air from Canada that clashes with warm maritime-tropical (mTw) air from the Gulf of Mexico associated with the Polar Front.
- Maritime tropical, warm air mass (mT) air from the Gulf of Mexico that tangles with continental-tropical (cT) air from Southwest United States and Mexico associated with a dryline (discussed in detail later).

Occurrence and movement of severe weather

Tornadic activity decreases during the summer due to the lack of strong dynamics aloft associated with the polar-front jet (PFJ) and the weak low-level thermal contrast. The favored area shifts with the position of the PFJ to the north as the strong solar radiation encompasses North America. Regions shift from the Gulf Coast states in the late winter, to the Southern Plains states in early spring, to the Central Plains and Midwest states in the late spring, and finally, to the Northern Plains and Great Lakes region in the summer. If you think about the PFJ and its associated energy shifting northward, it's easy to see why the severe activity shifts northward with the jet.

Thunderstorm effects on base resources and aircraft in flight

Thunderstorms look the same to observers, whether those individuals are on the ground or in flight. However, the hazards associated with thunderstorms affect ground base resources and aircraft in flight quite differently.

Categories and effects on base resources

Thunderstorms categorized as *non-severe* are those storms that produce wind speeds up to 49 knots and hail (that reaches the surface) larger than 3/4 of an inch. Thunderstorms categorized as *severe* are those storms that produce winds 50 knots or greater and/or hail 3/4 of an inch or larger. However, both types of thunderstorms could have the capability to produce heavy rains (2 inches or more within 12 hours, subsequent flash flooding, and with dangerous lightning). Heavy rainfall can cause flooding of an installation or deployed location damaging resources and facilities. Additionally, be aware the thresholds for severe thunderstorms may be different from CONUS to overseas. When discussing the categories and effects of thunderstorms on base resources you must consider high winds, hail, and lightning.

High winds

The most apparent problem associated with high winds is that small to moderately sized objects on the flight line and other exposed areas can be moved around if they are not tied down (i.e., tools, fire extinguisher carts, etc.) These moveable objects are dangerous to flight line personnel and can cause damage to aircraft. Visibilities can also be significantly reduced by blowing dust and dirt produced by high winds. Blowing dust and dirt can also penetrate equipment rendering it unusable.

Hail

Hail is not only dangerous to flight line personnel but to aircraft as well. For example, the T–38 aircraft is highly susceptible to hail damage. The hail can pit the aircraft’s “skin” (metal exterior) and also shatter the aircraft’s canopy. Hail damage can be very extensive, making it very expensive. As weather journeymen, forecasting whether high winds or hail will occur is a high priority. Maintenance crews need to have as much lead-time as possible to allow them to hanger aircraft, if such action is needed.

Lightning

Both types of thunderstorms (severe and non-severe) produce dangerous lightning. Lightning in the vicinity stops aircraft refueling and limits outside (flight line) work because of the danger to personnel. Possible power fluctuations due to lightning strikes make it necessary to switch to back-up power to maintain operations and keep weapon systems operating (i.e., weather station, Air Traffic Control, computer systems, etc.). The majority of installations require lightning advisories to be issued when lightning is observed within 10 nautical miles (nm) of the base and 3nm of the flight line. Lightning can travel for several miles and lead to fatal injuries if personnel are struck.

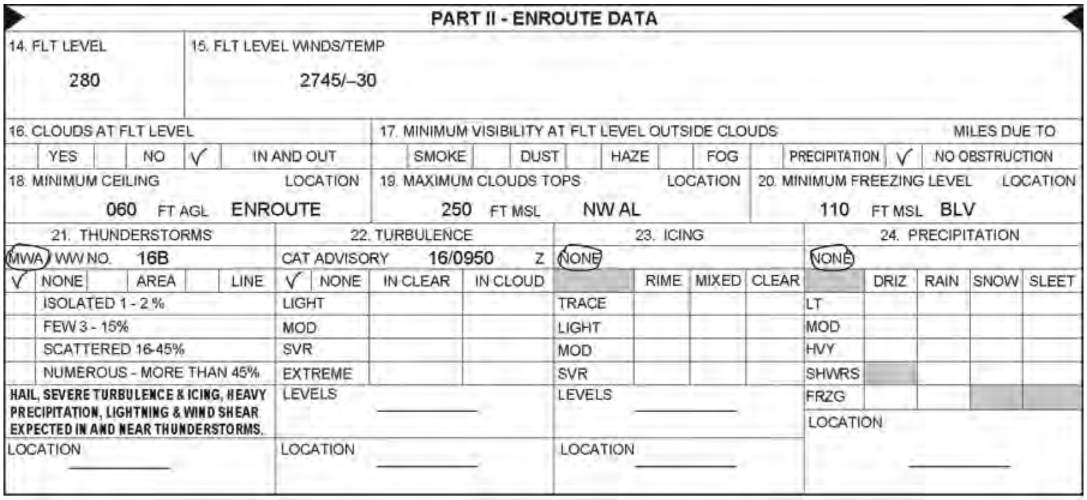
Effects of thunderstorms on aircraft in flight

A weather journeyman must understand that *all thunderstorms*, regardless of severity category, are considered *severe to aircraft in flight*. The weather journeyman must then convey the importance of this information to the pilot. Many pilots have made the mistake of thinking isolated areas of “garden variety” air-mass thunderstorms are not capable of causing damage to their aircraft. Nothing could be further from the truth.

Hail usually melts as it falls through the atmosphere. For example, at the surface, a thunderstorm may be producing 1/4 inch hail. However, at flight level the hail size may be in the 1 to 1-1/2 inch range. In this case, it’s obvious that the hailstone melted as it fell through the warm atmosphere where a high freezing level existed. At the surface, based on hail size alone, the thunderstorm would be considered non-severe. However, to an aircraft in flight, the hail size of 1 to 1 1/2 inches is definitely considered severe. While in flight, the aircraft would not only be susceptible to damage to the aircraft skin, electronic antennas, and the canopy, but would also be susceptible to the engine ingesting the hail and causing damage.

As we discussed earlier, wind gusts and low-level wind shear can be especially dangerous to aircraft on take-off and landings. Within the last few past decades, a number of commercial aircraft have crashed due to downburst and microburst events that were associated with thunderstorms.

Remember, in forecasting the severity of thunderstorms, the weather journeyman is forecasting the expected weather phenomena for the surface; not those aloft. Always brief the aircraft crews that all thunderstorms are considered severe while they are in flight. Figure 3–10 depicts the Part II-Enroute Data section of a DD-Form 175–1, Flight Weather Briefing form. Notice at the bottom of section 21, Thunderstorms, that it specifically states that “hail, severe turbulence and icing, heavy precipitation, lightning, and wind shear are to be expected *in and near* thunderstorms.” Read this statement, verbatim, to all aircrews that might encounter forecasted thunderstorm activity during their flight. By doing so, you make sure that aircrews understand the significance and importance of avoiding thunderstorms.



PART II - ENROUTE DATA

14. FLT LEVEL 280

15. FLT LEVEL WINDS/TEMP 2745/–30

16. CLOUDS AT FLT LEVEL ☐ YES ☐ NO ✓ IN AND OUT

17. MINIMUM VISIBILITY AT FLT LEVEL OUTSIDE CLOUDS — MILES DUE TO ☐ SMOKE ☐ DUST ☐ HAZE ☐ FOG ☐ PRECIPITATION ✓ NO OBSTRUCTION

18. MINIMUM CEILING 060 FT AGL LOCATION ENROUTE

19. MAXIMUM CLOUDS TOPS 250 FT MSL LOCATION NW AL

20. MINIMUM FREEZING LEVEL 110 FT MSL LOCATION BLV

21. THUNDERSTORMS

[✓ MWA] WW NO. 16B

✓	NONE		AREA		LINE
	ISOLATED 1 - 2 %				
	FEW 3 - 15%				
	SCATTERED 16-45%				
	NUMEROUS - MORE THAN 45%				

HAIL, SEVERE TURBULENCE & ICING, HEAVY PRECIPITATION, LIGHTNING & WIND SHEAR EXPECTED IN AND NEAR THUNDERSTORMS.

LOCATION ______

22. TURBULENCE

CAT ADVISORY 16/0950 Z

✓ NONE	IN CLEAR	IN CLOUD
LIGHT		
MOD		
SVR		
EXTREME		

LEVELS ______

LOCATION ______

23. ICING

[✓ NONE]

	RIME	MIXED	CLEAR
TRACE			
LIGHT			
MOD			
SVR			

LEVELS ______

LOCATION ______

24. PRECIPITATION

[✓ NONE]

	DRIZ	RAIN	SNOW	SLEET
LT				
MOD				
HVY				
SHWRS				
FRZG				

LOCATION ______

Figure 3–10. Part II, Enroute Data section of a DD 175–1, Flight Weather Briefing.

242. Basic conditions necessary for the development of convective severe weather

We can summarize the conditions necessary for the development of tornadoes, severe thunderstorms, and their associated destructive phenomena under five interdependent headings; temperature and instability, moisture, winds, lifting, and freezing level.

Temperature and instability

The thermal air structure must be *conditionally unstable* for the development of severe convective weather. The magnitude of the temperature is important in two ways—it controls the ability of the air to hold and transport moisture, and it affects the height of the wet-bulb-zero. The most vigorous storms occur in air having a subsidence type inversion in the morning; a characteristic associated with wind shear.

An unstable atmosphere is also required for thunderstorms to develop. A body is said to be unstable if, after a small displacement, it tends to continue moving away from its original position. In this case, if a parcel of air is displaced upward, it continues to move upward from its original position.

Moisture

Large quantities of moisture *must be available* for severe convective weather to occur. Usually, this requirement is fulfilled by the presence of a low-level tongue of moisture. The strongest storms require drier air above this ridge of moisture or a marked dry source to the windward side in a position to intrude into or over the moist ridge. There are many indications that all tornadoes *may* require some unsaturated air aloft to generate downdrafts and increase explosive overturning.

Warm, moist air in the low levels acts as fuel. The amount of moisture necessary to initiate thunderstorm development is variable. However, the warmer and moister the air is in the low levels, the greater is the potential for thunderstorm development.

The moist air near the surface needs to be lifted to higher levels for condensation to occur. The unstable lapse rate indicates that the environment offers little resistance to the vertical transfer of this moisture to higher levels. The lifting mechanism is important, because, in general, the greater the lift, the greater is the chance for the developing thunderstorm to become severe.

Winds

Strong mid-level winds *are also required* for severe convective weather to develop, except in an equatorial-type atmosphere. Sharp horizontal wind shears favor the development of severe phenomena through instability. Moderate to strong flow in the low level is required for major storm outbreaks of the family type (more than one). The intersection of the maximum low-level winds with a warm front or old squall line is a frequent area of development. From the forecasting prospective, it is extremely desirable for the mid-level and low-level jets to intersect, since the location and movement of this intersection frequently determine the major axis of tornadic activity.

Lifting

A lifting or release mechanism *triggers thunderstorm activity* leading to severe convective weather. Low-level heating, low-level convergence and orographic (upslope) conditions are considered release mechanisms.

NOTE: Positive vorticity advection (PVA) and upper-level difluence are not considered release mechanisms—PVA may indicate upper-level support; upper-level difluence supports and enhances convection but does not cause it.

As you know, severe thunderstorms with tornadoes, hail, or destructive winds do not develop spontaneously. They require a surface-based lifting mechanism, such as one of the following:

- Drylines.
- Cold front.
- Warm fronts.
- Upslope flow.
- Prefrontal squall lines.
- Sea breezes and land breezes.
- Thunderstorm outflow boundaries (fig. 3–11).
- Intersection of any two lines of activity such as squall lines, a squall line and a warm front, a mesocyclone moving along a warm front, or thunderstorm outflow boundaries and a sea breeze.

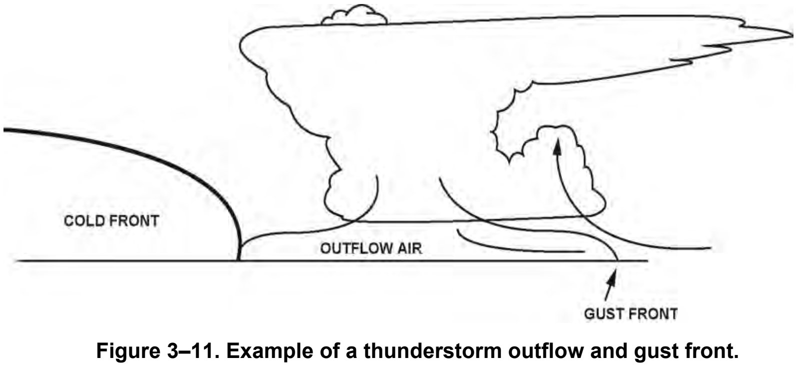


Figure 3–11. Example of a thunderstorm outflow and gust front.

Freezing level

The height of the wet-bulb-zero (WBZ) above the terrain *should be favorable* for severe convective weather to occur. This height in the air mass is assumed to be the height of the freezing level within the storm column. It correlates closely with the type and intensity of the severe phenomenon that reaches the ground, especially hail. The optimum height of the WBZ is about 8,000ft. When this height is below 5,000 or above 11,000ft, the incidence of surface hail is practically negligible and tornadoes, if any, are comparatively weak and short-lived, except in a type I air mass (air mass types were discussed in unit 2).

The occurrences of thunderstorms are dependent on the conditions and atmospheric dynamics available. There must be certain “ingredients” existing in the atmosphere to have thunderstorms develop.

Parcel theory

The parcel theory attempts to describe the process of free convection in the atmosphere leading to thunderstorm formation. The theory assumes no transfer of heat and moisture properties between the parcel and the ambient environment and is considered as an adiabatic assumption based on the lapse rate of the air mass.

If the lapse rate of the surrounding air mass exceeds the rate of cooling of the parcel through part of the layer through which it ascends, the parcel becomes warmer than its environment and buoyant. Basically, if the parcel is warmer than the environment, it rises due to positive buoyancy. Conversely, the parcel sinks if it is colder than the environment due to negative buoyancy. This process is similar to the one used to fly hot-air balloons. The burner is turned on, to heat the air in the balloon so it is a higher temperature than the outside air, and the balloon ascends. If the balloon pilot wants to descend he or she opens a flap at the top of the balloon, allowing the hot air to escape, decreasing the balloon interior air temperature, so it becomes cooler than the outside air.

The convective process

As we previously stated, determining the lapse rate of the air mass is very important when we consider whether thunderstorms develop or not. When it thunderstorms the atmosphere is in a “conditional state.” This can be determined by using a Skew-T for the thunderstorm area. A conditional state occurs when the environmental (actual temperature sounding) lapse rate is greater than the moist adiabat lapse rate but less than the dry adiabatic lapse rate.

The atmosphere is conditionally stable if a parcel is lifted dry adiabatically or conditionally unstable if a parcel is lifted moist adiabatically. For example, if a parcel were lifted dry adiabatically (following the dry adiabat upward from the surface temperature), it would be colder than the environmental temperature and, therefore, sink back to its original position. However, if it were lifted moist adiabatically (following the saturation adiabat upward from the surface temperature), it would now be warmer than the surrounding environmental temperature and, therefore, rise upward from its original position. This confirms our previous statements pertaining to positive and negative buoyancy.

In order for a parcel to condense in a conditionally unstable environment, it must be lifted to the convective condensation level (CCL) or to the lifted condensation level (LCL) and, ultimately, to the level of free convection (LFC). How the LFC is reached depends on the type of lifting.

If lifting is due to low-level surface heating, the CCL is also the LFC. If we determine that the CCL temperature will not be reached, then the LCL will need to be reached for convection to occur. The LCL is reached mainly due to low-level heating. However, additional mechanical lifting in the form of a front, orographic lifting, and so forth, is required for the parcel to reach the LFC.

Once condensation is reached, the parcel releases latent heat of condensation. How much heat is released depends upon the water content of the air parcel. Warmer air can hold more water vapor and, therefore, can release more latent heat when condensation occurs. Conversely, cold air holds little water vapor; thus less heat is released in the condensation process. The release of latent heat causes the parcel to cool at a slower rate. If this rate is less than the environmental lapse rate, the parcel continues to rise after it reaches the LFC because it is then warmer than the environment.

Positive energy area and equilibrium level

An increase in the low-level moisture content (moisture advection) results in lowering the CCL, LCL, and LFC. This results in an increase in the positive energy area (PEA) and greater positive buoyancy. The PEA area occurs from the LFC along the saturation adiabat (where the lifted parcel temperature is warmer than the environmental temperature) to wherever the lifted parcel temperature again crosses the environmental temperature on the parcel's ascent. The area outlined by points A to B′ and B of figure 3–12 indicate a "partial" PEA. It is partial because it has not been extended upward to where the saturation adiabats would cross the environmental temperature curve.

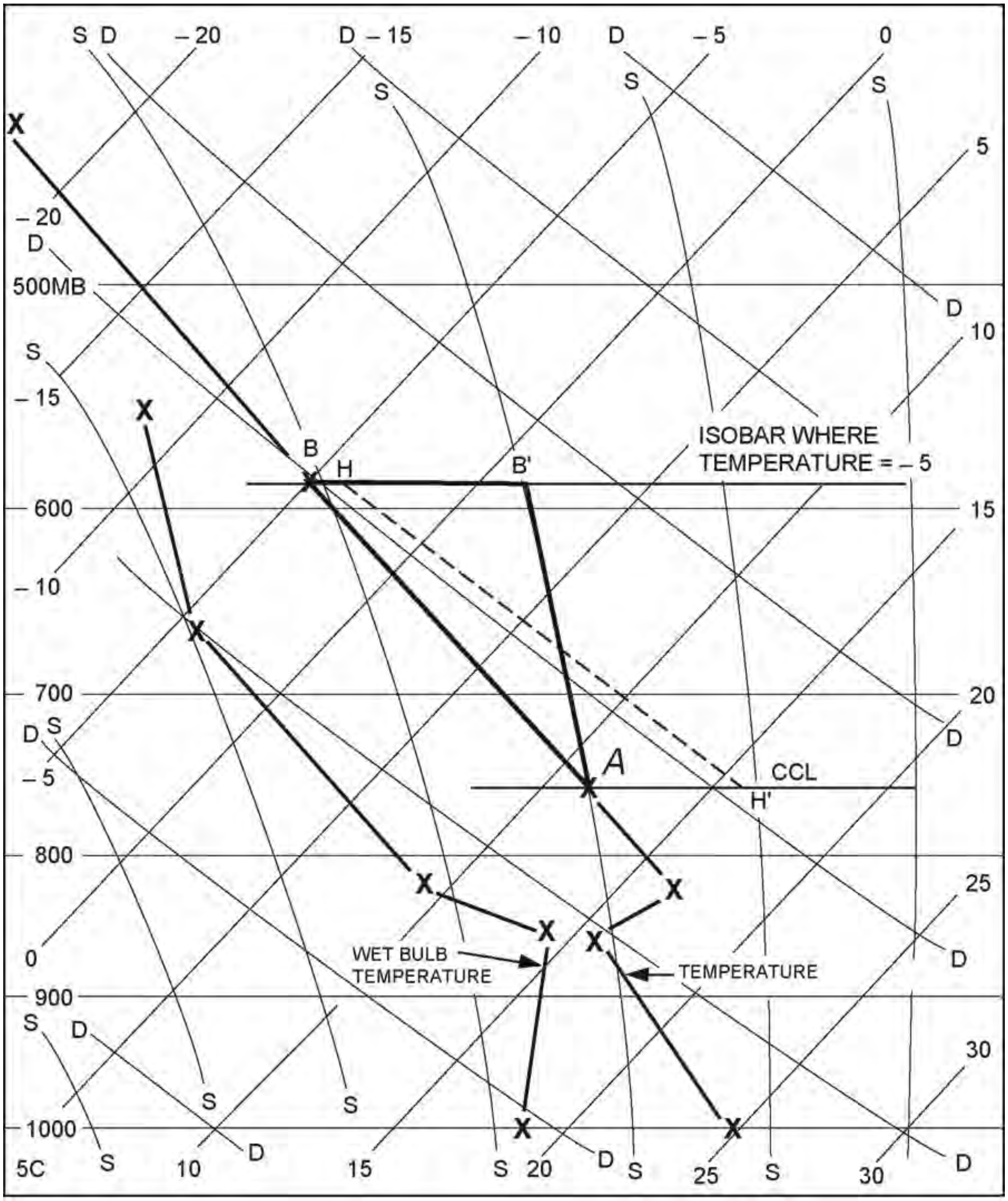


Figure 3–12. Partial example of a PEA (outlined triangle) —points A to B′ and B.

Once the LFC is reached, the parcel rises until it reaches the equilibrium level (EL). This level is the top of the PEA. The equilibrium level is also the top of the convective cloud or cumulonimbus anvil. On occasion, overshooting of the cloud top above the equilibrium level occurs. The overshooting is caused by the conservation of momentum associated with the updraft of the thunderstorm and is a severe thunderstorm indicator.

The rising parcels are referred to as the updraft. The strength of the updraft is determined by the width and vertical extent of the positive energy area. The wider and higher the positive energy area, the stronger the updraft is; this indicates a very unstable condition. The overall size of the positive energy area is a function of:

- The warm, moist air in the low levels.
- Cold air in the middle and upper levels.

In reality, some mixing and radiational heat exchange takes place between the parcel and the environment. As a result, the parcel often cools faster than the parcel theory would suggest. This is why most parcels only make it about halfway through the PEA and develops into towering cumulus clouds instead of cumulonimbus clouds. If the air aloft is too dry, the cloud evaporates, cooling the parcel. The parcel then loses its buoyancy and sinks. As the cloud evaporates, the environment absorbs the moisture. This change explains how initial convection often evaporates while later convection is often deep through the atmosphere.

Entrainment

Dry air is beneficial for enhancing deep convection. The process of mixing drier environmental air with the growing cloud (updraft) is known as entrainment. The outer edges of the cloud mix with the drier environmental air, evaporating the edges of the growing cloud. The evaporational cooling of the cloud-edge air causes the interior air parcels to rise even faster because a greater density gradient is created. Denser, less buoyant parcels would exist on the outer edges of the cloud with less dense, more buoyant parcels rising in the interior of the cloud. The gradient between the outside evaporationally cooling parcels and the interior less dense rising parcels increases. Therefore, the speed of the parcels in the updraft increases.

243. Thunderstorm types

Thunderstorms are classified as air mass, frontal, squall line, dryline, or orographic. Let’s look at each one in detail.

Air mass

Air-mass thunderstorms are the closest to the three-stage, non-severe thunderstorm (discussed later). These storms form in relatively uniform maritime tropical air masses (mT). They form in an area dominated by the subtropical ridge and are *not* associated with upper-level disturbances.

Air-mass thunderstorm cells form due to a combination of low-level (surface) heating and weak, low-level convergence. They are very diurnally dependent and are associated with the time of maximum heating. Weak winds aloft, which are inherent to mT air masses, result in little cell movement. The lack of movement and high water vapor content can result in the production of local heavy rain.

Severe weather is rare, but is possible, especially if an upper-level short-wave trough moves into the area or if there is additional low-level mechanical lift. These thunderstorms are short-lived and usually last an hour or less.

Frontal thunderstorms

Frontal thunderstorms are usually stronger than air-mass thunderstorms due to the synoptic scale forced lifting with the front. They are long lasting (up to 3 hours) due to the mechanical support of the updraft. Normally, thunderstorms form in bands or squall lines in the warm sector of the system and are the most intense thunderstorm variety.

Frontal movements and the presence of winds aloft cause the thunderstorm cells to move. Usual movement is with the 500-millibar (mb) flow at roughly 50 percent of the speed (or 70 percent of 700mb). Frontal thunderstorms favor maximum heating times, but can occur at any time due to mechanical lift support. The five types of frontal thunderstorms we'll discuss are listed in the following paragraphs.

Cold frontal thunderstorms

If unstable air overruns a front, thunderstorms may develop. Thunderstorms frequently form in bands parallel to and ahead of a front, with individual cells moving parallel to a front (multicell structure). The potential for severe weather depends on the air-mass stability, strength of the front, and strength of the upper-level dynamics. Thunderstorms occur with both active and inactive cold fronts, and with squall line activity associated with the inactive cold front.

Warm stationary frontal thunderstorms

Thunderstorm activity associated with warm stationary fronts requires strong overrunning of unstable air. This situation is usually associated with an 850mb maximum wind band intersecting the surface front. Thunderstorm activity may also be associated with an upper-level short-wave trough in the area of a front.

The thunderstorms form in clusters or bands on the cold side of the surface front. They are often embedded in stratiform clouds and precipitation. This creates a very dangerous situation for aircraft flying through the region. Due to radar attenuation, the pilot may have difficulty identifying where the intense storm activity is located along the flight path even with onboard Doppler radar.

Squall-line thunderstorms

Squall-line thunderstorms often form along or ahead of inactive cold fronts in unstable, moist air. Updrafts normally enter from the front of the squall line and tilt vertically due to vertical speed shear. Dry, mid-level air enters the rear of the squall line and is forced downwind behind the updraft. This creates the outflow (gust front) ahead of the line, resulting in low-level convergence and regeneration of the updraft.

The squall line continues to propagate (redevelop) itself forward due to the low-level convergence between the outflow boundary and the southerly flow of the warm air mass. Cells develop on the southern end and move up the line to the northeast with time. The squall line frequently propagates away from the cold front (its original lifting mechanism) and weakens when it outruns the unstable air mass.

The decaying squall line often takes on characteristics similar to a quasi-stationary front, with rain-cooled (stable) air on one side and mT (unstable) air on the other side. Squall line passage is often confused with a frontal passage. The rain-cooled air mass often recovers and returns to mT characteristics after the squall line passage.

Dryline (Marfa Front) thunderstorms

The dryline is a severe weather producing system located over western Texas, Oklahoma, and southern Kansas. This system is sometimes referred to in Texas as the Marfa Front because its genesis is near the town of Marfa, located in the Davis Mountains. The dryline feature is more of a pseudo-front because it does not require upper-level jet stream support to exist. It does, however, separate two very different air masses, cT and mT. Thunderstorms develop along a strong moisture gradient where low-level convergence is present.

During the time of development, a dew-point temperature gradient of 15°C in 2 kilometers (km) across the dryline is not uncommon. The development of the dryline is favored during maximum heating. It migrates eastward throughout the day and recedes to the west at night. Climatology supports spring and early summer as the seasons of prime activity.

Infrequently, severe thunderstorm clusters or lines (multicellular) form along the dryline. The most violent thunderstorms occur when the dryline teams with an upper-level atmospheric disturbance (i.e., a short-wave trough) along with intersection points between the dryline and other boundaries such as lake breezes, outflow boundaries, and so forth. Large hail, damaging winds, powerful and long-lasting tornadoes are common with the enhanced support. Activity is mainly limited to the late afternoon and early evening, although severe thunderstorms can develop along the dryline during its nocturnal retreat.

Orographic thunderstorms

When mountains or hills form an obstruction to the horizontal wind flow, the air is forced to rise vertically. If the air forced upward is convectively and conditionally unstable, thunderstorms may form. These types of storms may develop rapidly and cover large areas when the wind flow and the air mass properties are favorable.

It is characteristic of orographic thunderstorms to remain stationary on the windward side of the mountains. The formation of thunderstorms over mountainous terrain is attributed, in part, to the differential heating (associated with valley breezes) of the mountain tops and the free air at the same altitude. An example is the high frequency of thunderstorms over some mountain ranges when the field is such that the thunderstorms cannot be explained by orographic lift. The eastern Rocky Mountain area of New Mexico and Colorado is such a region.

244. Stages of non-severe thunderstorm development

As previously discussed, thunderstorms develop when warm, moist air near the earth rises and replaces denser air aloft. The lifting results in the condensation of atmospheric water vapor forming a visible cloud of water droplets. Let’s now discuss the development stages of non-severe thunderstorms. The three stages of non-severe thunderstorms consist of the cumulus, mature, and dissipating stages.

Cumulus stage

The cumulus stage lasts about fifteen minutes. During this time the cloud develops (grows) vertically due to the updraft. This stage consists only of updrafts that result from density differences and are enhanced by other dynamic factors, such as low-level convergence and upper-level divergence (fig. 3–13).

Precipitation forms aloft and is suspended by the updraft. The first detectable radar echo appears in the mid and upper levels of the cloud. This can occur in as little as ten minutes from initial cloud formation. When the precipitation aloft becomes heavy enough, gravity allows it to fall through the updraft and eventually reach the surface. This marks the beginning of the second stage of development—the mature stage.

Mature stage

The mature stage is the most violent and intense stage and consists of updrafts and downdrafts. The downdraft results primarily from density differences created by a number of processes; the most important of which is evaporational cooling. Heavy precipitation occurs at the surface beneath the downdraft (fig. 3–14).

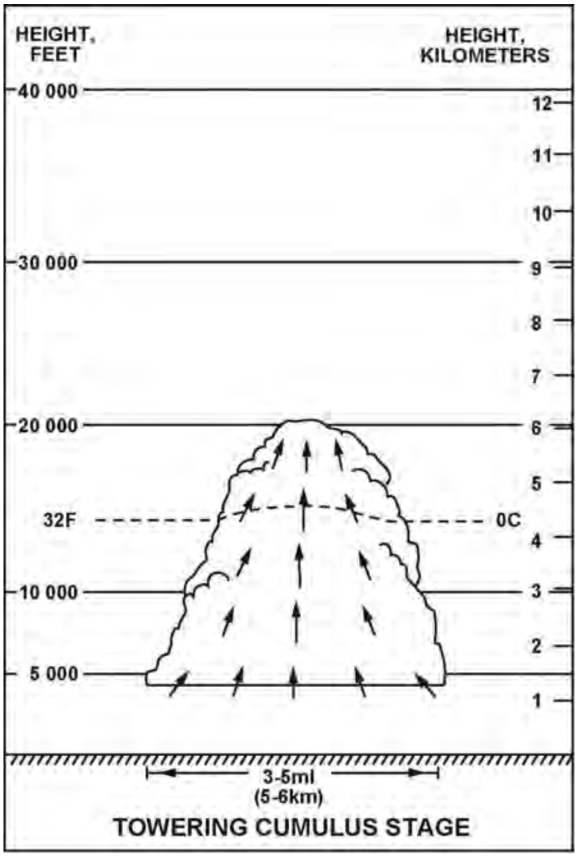


Figure 3–13. Cumulus stage of a non-severe thunderstorm.

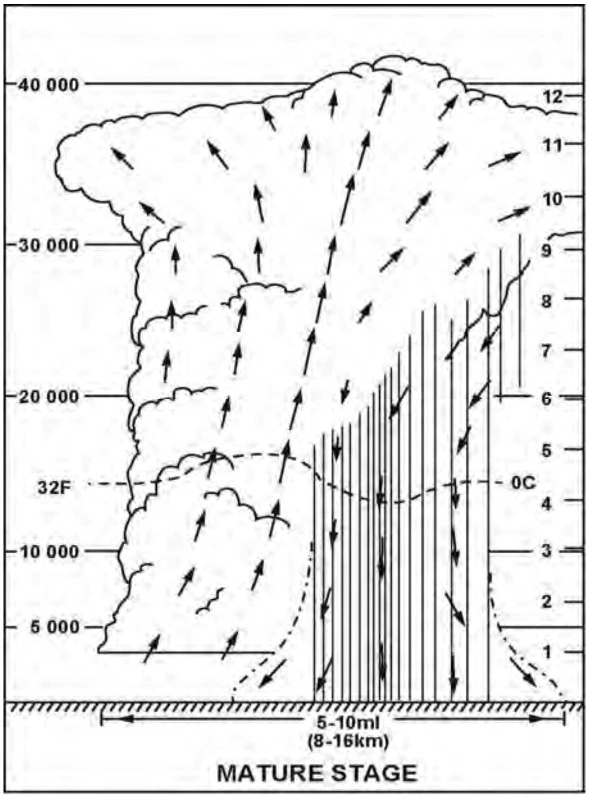


Figure 3–14. Mature stage of a non-severe thunderstorm.

Because the mature stage is the most violent there are other hazards that may occur during the mature stage such as lightning, hail, and strong downdrafts.

Lightning

During the precipitation process, electrical charges (positive and negative) are developing. Relative motions between ice crystals and snow pellets caught in vigorous updrafts and downdrafts causes large-scale charge separation. Once a cloud becomes charged to the point where the electrical field exceeds the local strength of atmosphere to support a separation of electrical charges, a lightning flash results, equalizing the charge distribution. This can occur within the cloud, between the clouds and the surface, between clouds, or between the cloud and clear air.

A quick way to determine the distance to a thunderstorm is by use of the "five second rule." When a lightning flash occurs, count in thousandths of seconds (i.e., one one thousand, two one thousand, etc.) until you hear the associated thunder. Divide the total seconds by five. This number approximates the distance (in miles) that the thunderstorm is from your position. For example, if you counted 15 one thousands, the thunderstorm would be approximately three miles away. However, always verify the distance to the thunderstorm by use of radar or other measuring capability.

Hail

Hail forms in the shear zone between the updrafts and downdrafts near the freezing level in a convective cloud. The developing hailstone grows rapidly due to the accretion of super-cooled water droplets on ice crystals. Hail stone sizes ranging from 1/4 inch to over 5 inches may be encountered anywhere in the proximity of a thunderstorm. In fact, the largest hailstone recorded in the US had a 17.5 inch circumference and occurred at Coffeyville, KS. Imagine the strength of the updraft that had to be present to create such a hailstone! This hailstone had an estimated fall speed of 83 knots.

Hail initially forms as ice crystals above the freezing level. As the ice crystal falls below the freezing level, it partially melts. Hail stones pick up a coating of water while they are below the freezing level and then freeze as they are carried back above the freezing level. This results in an interior of concentric ring patterns. Hail is very common aloft but often melts before it reaches the surface. Hail size at the surface is dependent on two factors:

- The strength of the updraft core (needed to support the hailstone).
- The length of fall from the freezing level to the surface (melting time).

Downdraft

Strong, gusty surface winds are common due to the downdraft. The leading edge of this cold down- and out-rushing air is called the gust front or the outflow boundary. Dangerous wind shear exists at this point. The strength of the downdraft and resulting surface winds is influenced by two factors:

- The heavier the precipitation, the stronger the downdraft.
- Entrainment of dry air causes evaporational cooling. This increases the density of the air in the cloud, causing it to sink and further enhancing the downdraft. Occasionally, the downdraft can be greater than 50 knots.

The formation of the downdraft signals the beginning of the end of the thunderstorm. The precipitation falling into the updraft weakens it causing the precipitation being held aloft to fall. The downdraft strengthens at the expense of the updraft. Since the updraft is the "fuel line" of the thunderstorm, the moisture inflow is reduced. The cool outflow air is very stable. This stable air gets drawn into what is left of the updraft and stabilizes the whole convective column. When the updraft ceases, the final stage of thunderstorm development begins.

Dissipating stage

With the collapse of the updraft, the moisture source is cut-off and the precipitation formation ceases. Without the falling precipitation, the downdraft also begins to weaken. Only light rain falls at the surface. With only downward subsiding air now prevailing, the cloud stratifies and eventually dissipates, leaving only the cool outflow behind which forms a mesoscale high (fig. 3–15).

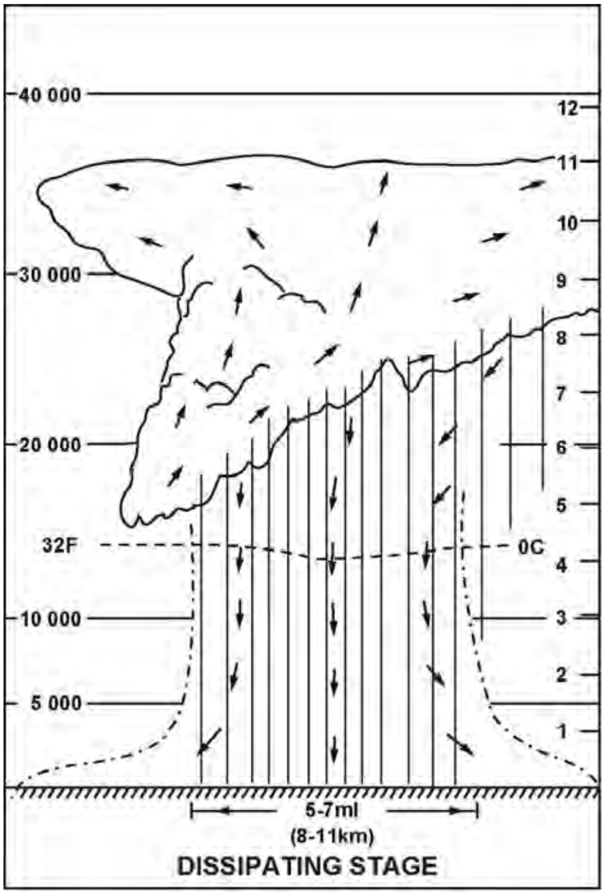


Figure 3–15. Dissipating stage of a thunderstorm.

The meso-high is a “bubble” of higher pressure located at the base of a downdraft and resulting from the increased density of subsiding evaporatively cooled air striking the surface. The outflow is the horizontal wind, which spreads out in all directions from the meso-high at the base of the downdraft as it strikes the surface (fig. 3–16). The average total lifetime of a thunderstorm cell is about 60 minutes; however, a storm consisting of sequence regenerating cells may persist for several hours.

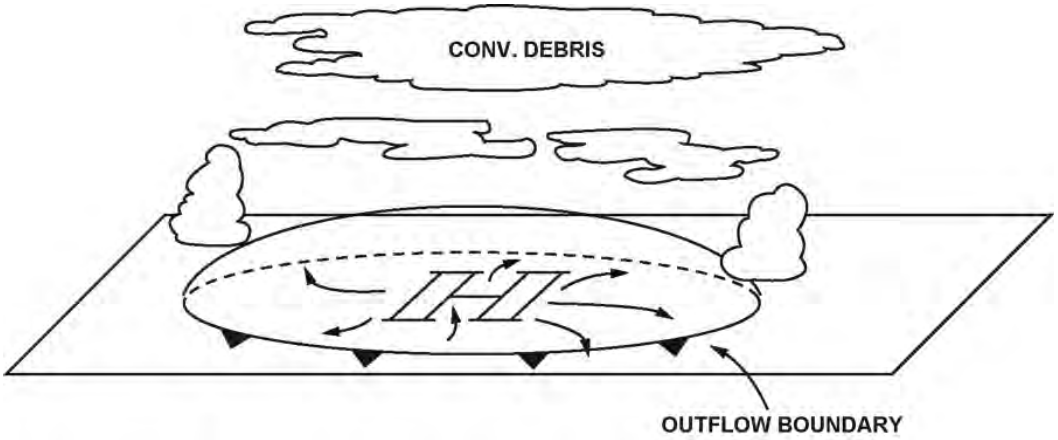


Figure 3–16. “Bubble” high formation.

245. Types and development of severe thunderstorms

Severe thunderstorms require more instability than non-severe thunderstorms. The updrafts must remain unobstructed by falling precipitation, thereby maintaining the updraft core. The presence of strong mid-level winds (varying from 700mb to 300mb, depending on storm height) is essential. These winds carry the precipitation downstream where it falls ahead of the updraft, rather than back into it. These dry mid-level winds also help to dynamically intensify the updraft by increasing the density gradient between the updraft and the mid-level air.

Multicell severe thunderstorms

With low-level flow continuing to feed warm, moist air into the storm, the updraft grows and intensifies especially in the mid-levels, where the strong updraft holds much of the precipitation aloft. Radar often confirms this by depicting a large, intense echo return aloft.

The inflow into the updraft normally enters the storm at the lower levels from the right flank (with respect to its movement). Because it pushes precipitation up to higher levels, there is little or no radar echo return in the updraft core region. This is referred to as the weak echo region (WER) (fig. 3–17). Because precipitation is held aloft in and near the core of the updraft, a WER develops. We can see the middle and upper level echo over a region of weak or absent low-level echo.

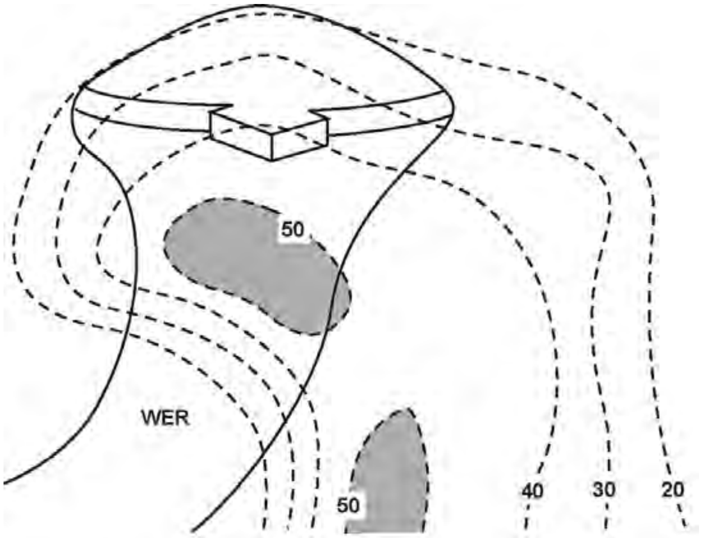


Figure 3–17. Example of a WER.

At this time, the updraft is strong enough to support hail stone sizes that meet the requirements of a severe thunderstorm classification. The largest hail falls just outside and downwind of the updraft core. Smaller hail is carried further downwind with the upper-level winds.

New cells frequently form on the inflow flank to the south, due to strong convergence between the outflow (downdraft) and the inflow. These cells replace older cells that have weakened and moved downwind. Thus, more than one cell updraft exists in the same convective cloud. This is referred to as the multicell severe thunderstorm.

In the vertical cross section of figure 3–18, we see how four cells (I-IV) evolve. The dashed lines in the diagrams represent radar returns in decibels (dBZ). In A of figure 3–18, because cell I is entering its dissipating stage, it is dominated by weakening downdrafts. Cell II is just reaching the stage where precipitation is beginning to form and, therefore, is dominated by updrafts. Cells III and IV are still towering cumulus forms.

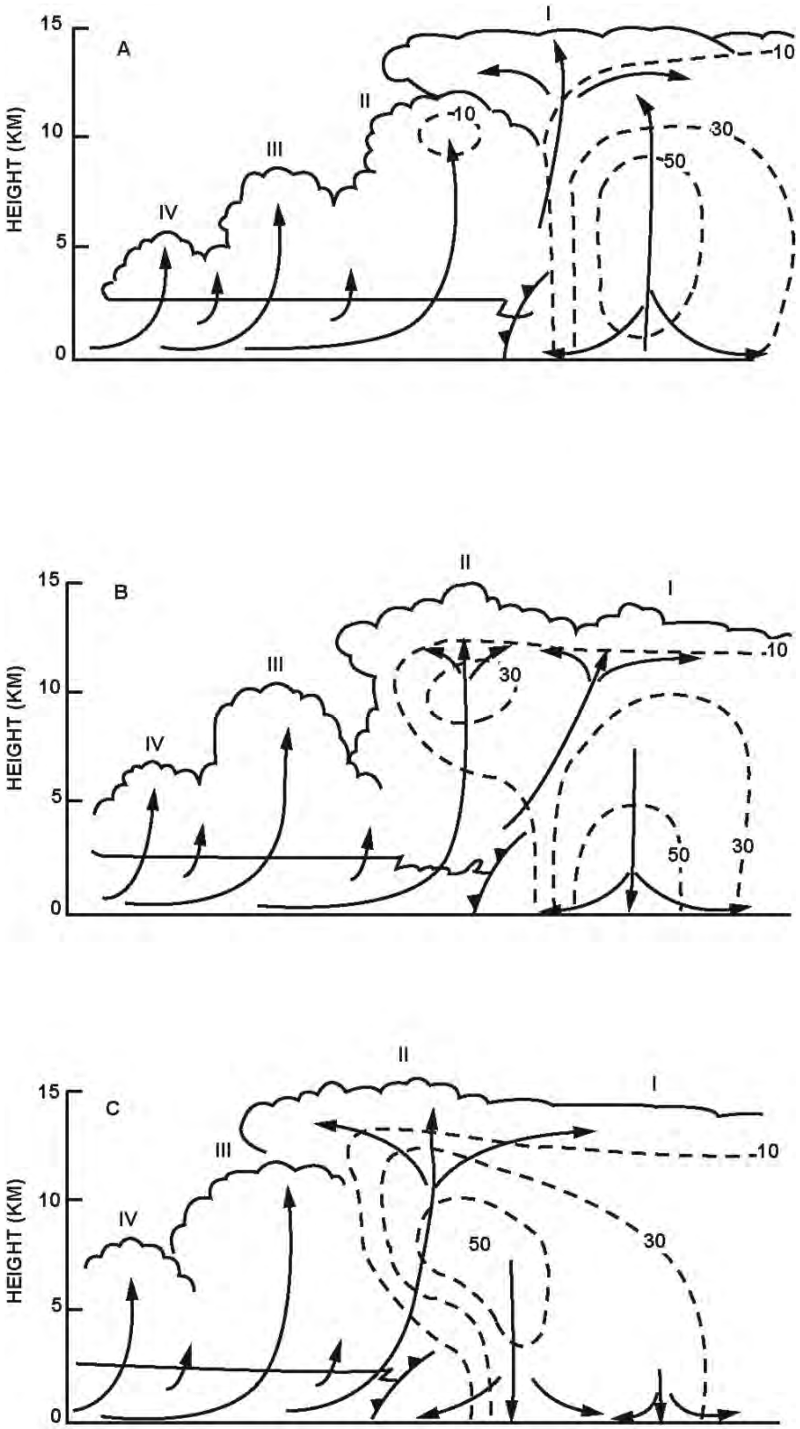


Figure 3–18. Development of multicell thunderstorms.

Several minutes later, the heavy precipitation of cell I has reached the surface and its updraft is weakening (B of figure 3–18). Now the top of cell II has passed through its equilibrium level and is decelerating, which forces it to spread out aloft. This spreading out, combined with the supporting updraft, produces a pronounced WER. Cells III and IV continue to develop.

Now after several minutes, cell I has all but dissipated, while cell II has entered the mature stage (C of figure 3–18). Cell II's high radar reflectivity core has begun its descent with its precipitation reaching the surface. Cell III is on the verge of being indicated on Doppler radar, while cell IV is still a growing towering cumulus.

The regeneration of new cells forming on the inflow flank, while older cells decay, causes the thunderstorm complex to *appear* to move to the right of the mid-level winds. In reality, the cells are moving with the mid-level winds and not to the right. This apparent erratic movement is referred to by radar meteorologist and co-designer of the WSR–88D, Mr. Lesley Lemon as “discrete propagation.”

Figure 3–19 shows cells A-D developing and dissipating through time. Cell A has developed on the left edge of the right diagram while cell B to the south is beginning to develop. The next diagram shows cell A moving northeast with the mid-level steering winds and dissipating. Note that cell B is now well developed and is also moving northeast while cell C is just beginning to develop. As new cells are developed on the inflow flank to the south, older cells are moving to the northeast and dissipating. The new cells are developing on the southern flank with the older cells dissipating to the north. This results in an apparent eastward progression of the line of thunderstorms.

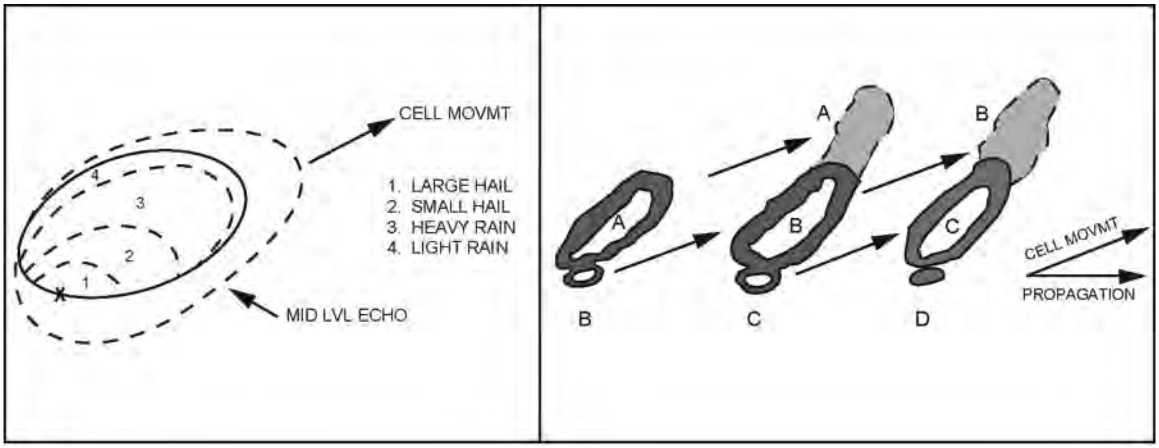


Figure 3–19. Discrete propagation.

Although multicellular events may be the most common type of severe storm, they are *generally* not the most severe. Meteorologists have studied these phenomena and have found that multicell thunderstorms have less probability of producing the kind of severe weather, other than hail, than do supercell thunderstorms.

Supercell severe thunderstorms

If the air mass is very unstable and the vertical wind profile is veering and increasing in speed with height, the updraft of the severe multicell continues to intensify. Strong, low-level inflow into the right flank continues to feed an abundant supply of moisture into the storm. A strong polar-front jet (PFJ) further enhances the updraft and acts as an exhaust mechanism.

In the lower levels, if the atmosphere possesses a horizontal roll due to vertical speed shear, the horizontal roll can be entrained into the updraft. This vertical tilting results in cyclonic rotation of the updraft. During this vertical tilting a conversion from horizontal to vertical vorticity takes place. The rotating updraft becomes so intense that it begins to divert the mid-level flow around it. The updraft is acting as a barrier to the environmental flow. We can visualize this effect by imagining a rock in a stream. The rock acts as the barrier and the water is the environmental flow that is diverted around the rock. There are other weather conditions to consider when discussing supercell severe thunderstorms such as the magnus effect, rear and forward flank downdrafts (FFD), and the phenomena associated with rear-flank downdrafts (RFD).

Magnus effect

The magnus effect is the rightward shift of the storm with respect to the environmental flow. The flow around the cyclonically rotating updraft causes a pressure increase on the left side (resulting in convergent flow) and a pressure deficit on the right (resulting in divergent flow). This causes the updraft to shift to the right toward lower pressure to equalize the pressure imbalance and the entire storm to shift to the right with the updraft (fig. 3–20).

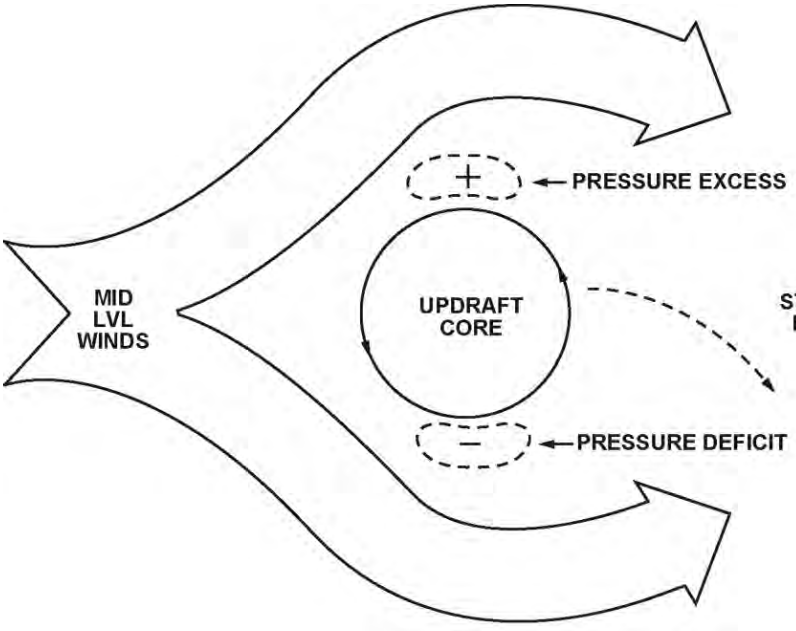


Figure 3–20. Magnus effect.

With the weather condition just described, the updraft is dominating much of the low-level moisture. This causes the surrounding cell's updrafts to weaken and die. This leaves one mammoth cell—the supercell (fig. 3–21). The core of the intense updraft forces the precipitation aloft even higher, digging a hole and creating an overhang in the mid-level precipitation area. This creates the bounded weak echo region (BWER), sometimes referred to as the vault.

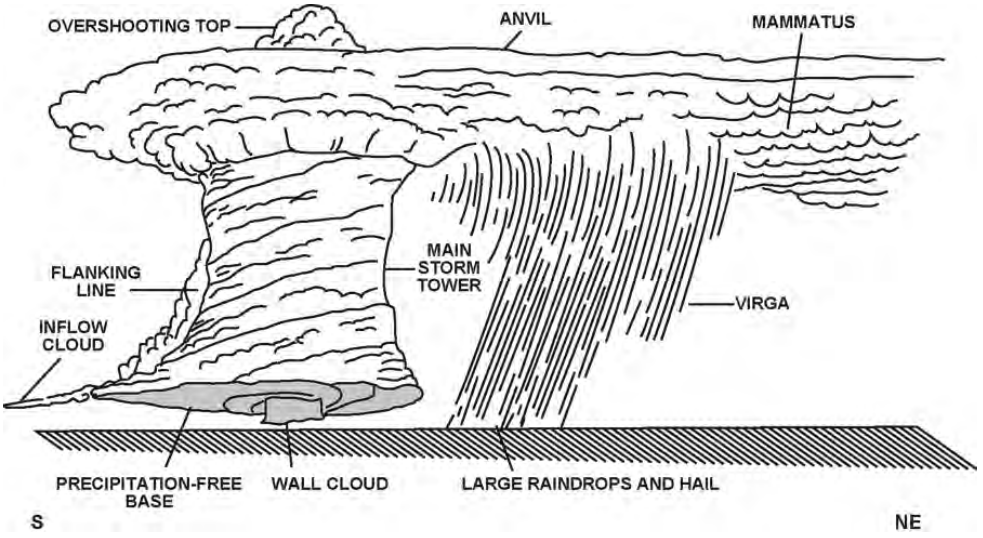


Figure 3–21. Example of a supercell.

These BWERs occur in supercells when the weak echo region in the lower levels actually extends upward into, and is surrounded by, the high radar reflectivities aloft. Such a structure has been understood to represent a high-speed updraft core within the larger region of updraft. Such a strong updraft forces forming precipitation upward rapidly, producing a weak echo region bounded on three sides by higher reflectivities as seen in figure 3–22. Any thunderstorm possessing a BWER is considered severe.

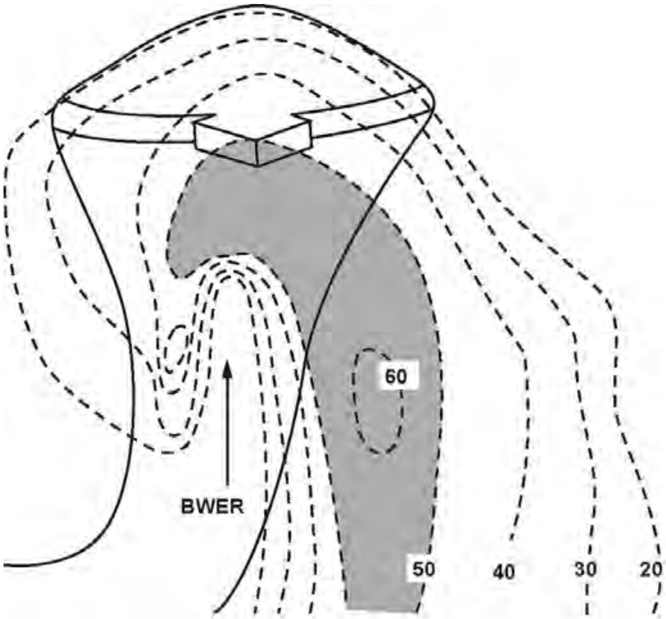


Figure 3–22. Example of a BWER.

Rear and forward flank downdrafts

Hail both at the surface and aloft reaches its maximum size when the BWER is present. Funnel clouds are often observed around the base of the updraft core. Some of the mid-level flow on the upwind side (rear) of the updraft core piles up and is forced downward due to greater density differences. This creates the rear-flank downdraft (RFD). The downdraft associated with the gust front is referred to as the forward-flank downdraft (FFD) (fig. 3–23).

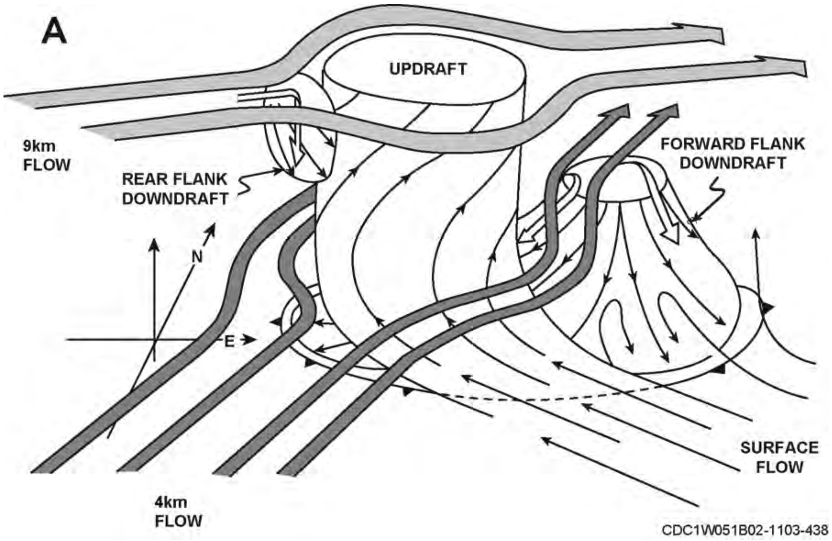


Figure 3–23. Initiation of FFD and RFD.

If the mid-level winds are strong, the RFD can cause severe winds, greater than 50 knots, when it reaches the surface. The RFD is forced around the right side of the updraft core due to the cyclonic rotation of the core. Some precipitation wraps around with the RFD making a prominent, pendant shaped, reflectivity signature on the right-rear quadrant of the storm on the WSR–88D (fig. 3–24).

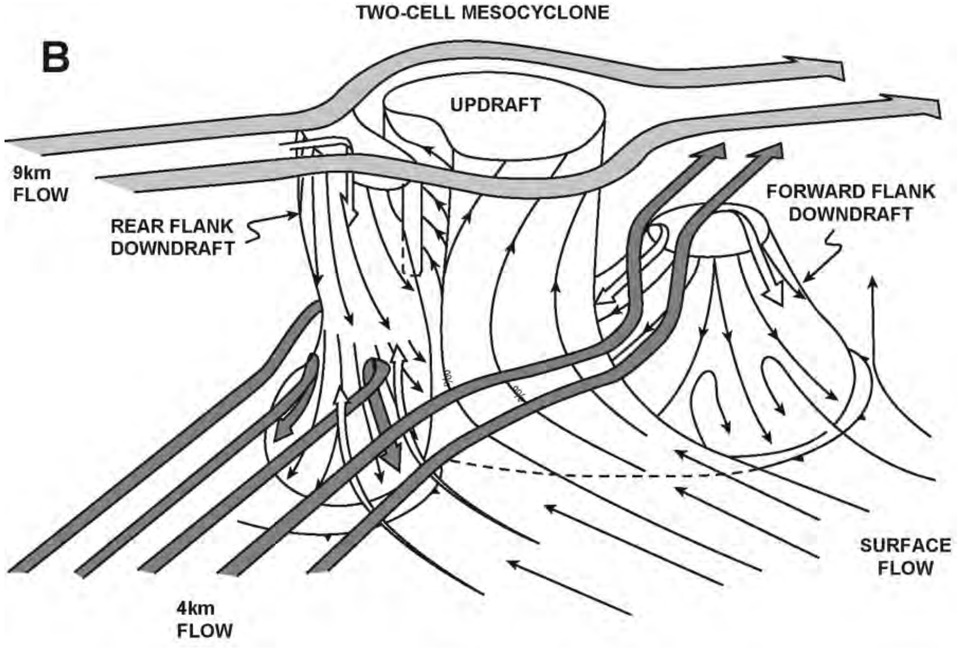


Figure 3–24. RFD continues to develop.

A strong cyclonic shear zone is formed between the outflow from the RFD and the horizontal low-level inflow into the updraft. The updraft begins to weaken due to its interaction with the RFD. The storm's overshooting top and the BWER collapse and the hail reaching the surface increases significantly (fig. 3–25).

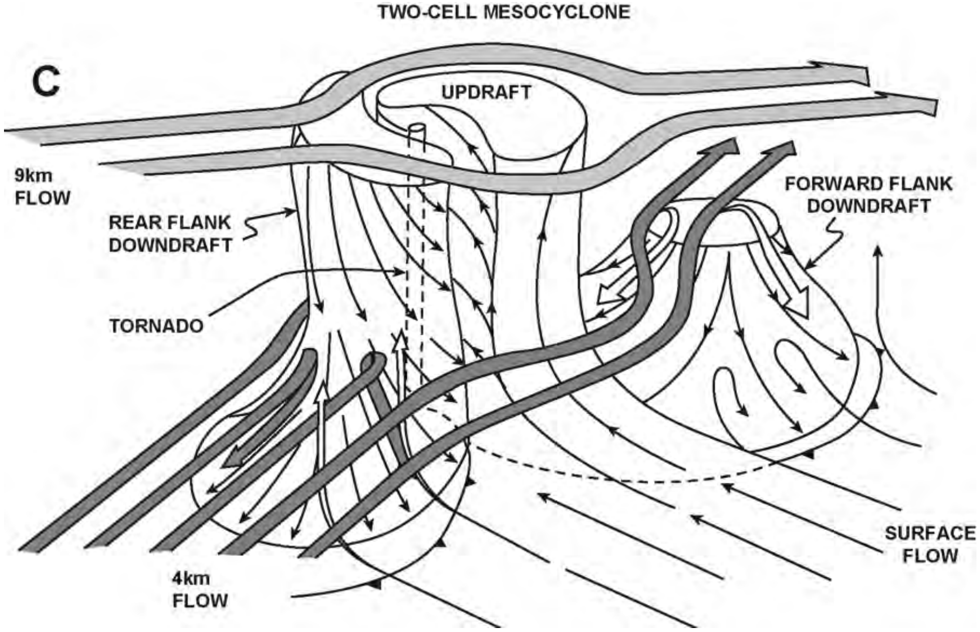


Figure 3–25. Updraft weakens due to RFD interaction.

The RFD begins to dominate at the expense of the updraft. Damaging surface winds are likely in the pendant region. The strongest rotation of the mesocyclone is in the updraft found in the mid levels. *This is the point in time when the mesocyclone can be identified on the WSR–88D.* The RFD drags the rotation downward while stretching the vortex tube or column, thus reducing the radius and increasing the wind speed. If the mesocyclone interacts with the low-level cyclonic shear zone, the vortex reaches the surface in the form of a tornado. The small triangle at the apex of the occlusion represents the approximate location of the tornado.

As the RFD develops, the circulation core comes to lie near the boundary between the updraft and the RFD. The view of the circulation at the lower levels looks remarkably like that of an extratropical cyclone, with gust fronts, downbursts, and microbursts replacing the large-scale fronts. As the storm evolves, the RFD-produced downburst on the rear flank sweeps around the circulation; inducing updrafts ahead of it. This causes the updraft region to look horseshoe shaped, and the circulation “occludes,” much like the extratropical cyclone (fig. 3–26).

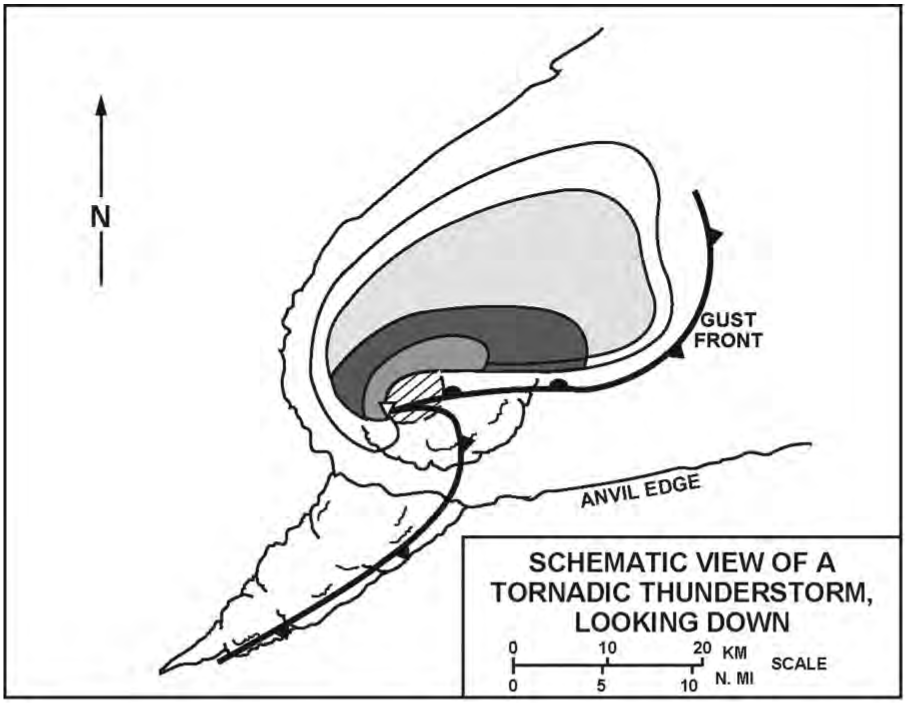


Figure 3–26. Mesocyclone and shear zone development.

The radar pendant wraps up into a hook, identifiable on radar reflectivity products, and will eventually disappear. When the updraft completely weakens, the tornado soon dissipates. A new updraft frequently forms on the inflow, right flank, and the entire cycle may repeat itself, as the system becomes a meso-occlusion.

Phenomena associated with the rear-flank downdraft

We now know that supercells commonly produce tornadoes. However, it also has the capability of producing other damaging phenomena associated with the RFD—downbursts and microbursts.

Downbursts

The downburst is a concentrated downdraft with great potential for producing damage at the surface. It often produces damaging winds that reach at least 35 knots. This phenomenon is dynamically enhanced during a rear-flank downdraft, it is not the usual downdraft associated with the heavy precipitation, gust front and FFD. Downbursts are often found behind the gust front and appear to be a form of a RFD.

Microbursts

Microbursts are concentrated downbursts, 0.2nm to 2.4nm in size, with a greater potential for producing damage at the surface. This potential is primarily caused by straight-line winds. Often, tornado like damage equal to F3 on the Fujita scale occurs. Microbursts are associated with a change in velocity of 50 knots or greater and are often located within the downburst. Tornado-like (shear) vortices can also occur along the leading edge of the microburst and can inflict significant damage to the surrounding area. Both downbursts and microbursts create dangerous low-level wind shear. Catastrophic accidents have occurred to aircraft as they attempted take-off and landings in downburst and microburst environments.

Splitting cells

Occasionally, a supercell may split into two separate cells, one right and one left. This is primarily caused by:

- Environmental wake flow “cutting” into the updraft on the downwind side of the updraft core.
- Precipitation “loading” in the mid levels above the updraft core.

The environmental wake flow situation is similar to the previous discussion about the supercell. For clarification, let’s again use our example of a rock in a stream. As water is diverted around the rock, it creates a wake on the downstream side of the rock. The same situation can be applied to the updraft core as the environmental flow is diverted around it. If the environmental flow is strong, it also creates a wake on the downstream side of the updraft core. The wake flow can eventually “cut” into the updraft core on the downstream side, splitting it into two storms—cyclonic and anticyclonic rotating storms.

The right cell often intensifies into a supercell and takes on a cyclonic rotation. It deviates approximately 30 degrees to the right and slows down. The right splitting cell is a potential tornado producer because as it shifts right, it moves into the warm, moist air that continues to “fuel” and intensify the storm.

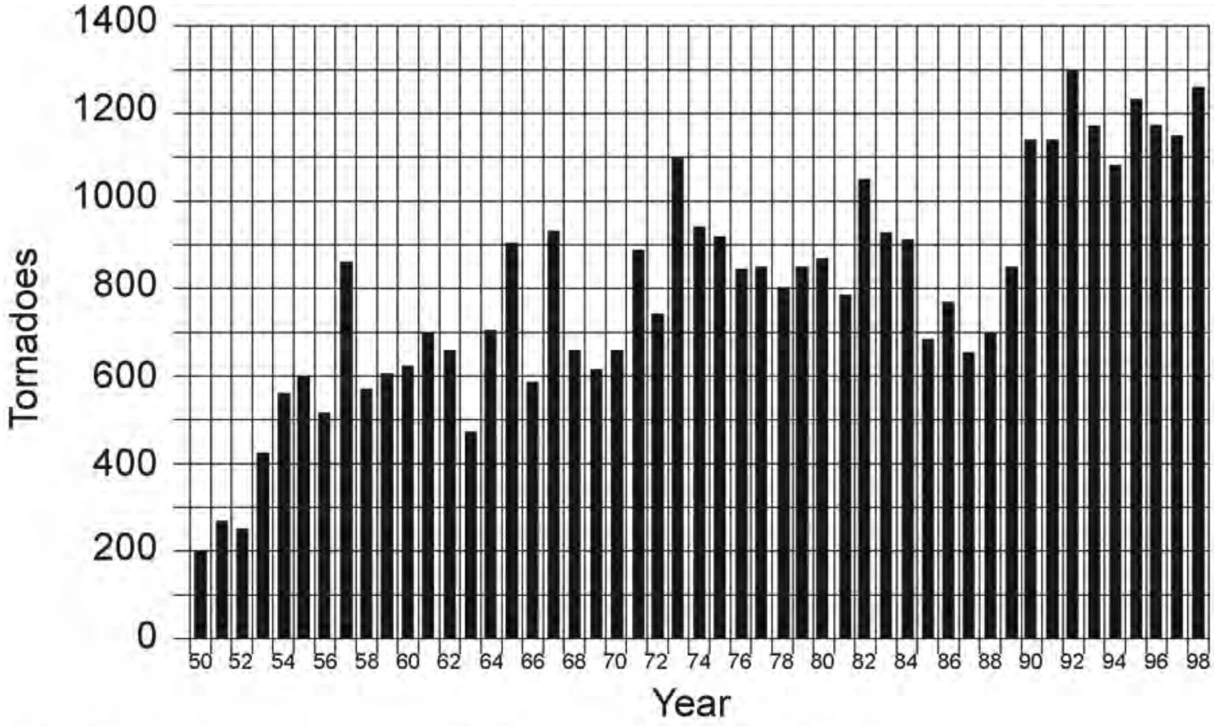
The left cell often weakens and takes on an anticyclonic rotation. It weakens because it moves away from the warm, moist air. It deviates approximately 50 degrees to the left and accelerates. The left splitting cell is a good hail producer because it moves into the colder air with associated lower freezing levels. Tornadoes are rare, but when they do occur they exhibit anticyclonic rotation.

246. Tornado statistics

Tornadoes are the least understood atmospheric phenomena. Only their comparatively small size ranks them as second in the severity of the damage they cause, with tropical cyclones ranking first. For their size, tornadoes are unsurpassed in the damage they can cause.

Tornadoes are violently rotating columns of air extending downward from a cumulonimbus cloud. They are nearly always observed as funnel clouds. They occur only in certain areas of the world and are most frequent in the US in the area bounded by the Rockies on the west and the Appalachians on the east. Tornadoes also occur during certain preferred seasons of the year in the US with their most frequent occurrence during May. The season of occurrence varies with the locality. Eighty percent of the tornadoes in the US have occurred between 1200 and 2100 hours. Figure 3–27 shows the occurrence of tornadoes in the US between 1950 and 1998.

The highest observed wind speed within a tornado was made by research meteorologists from the Storm Prediction Center equipped with a portable Doppler radar on 3 May 1999. A tornado near Moore, Oklahoma, 6 miles southwest of Tinker AFB, was observed with maximum winds of 313 miles per hour (272 knots). This tornado was classified as an intense F5 on the Fujita scale. Thirty eight deaths were attributed to this storm.



1998 total is based on preliminary statistics tabulated by The Storm Prediction Center.

Tornado fatalities dramatically increased in 1998. There were 129 tornado-related deaths which is the most since 1974 when there were more than 350 fatalities. In only one year since 1974 were there more than 100 deaths and that was in 1984 where there were 122 deaths.

Figure 3–27. Number of tornado occurrences in the United States.

The following two tables, Tornado Facts and Synoptic Requirements, summarize most of the information presently available on tornadoes concerning the location of most frequent occurrence, seasonal frequency, diurnal frequency, movement, width of path, synoptic requirements, and synoptic properties.

Tornado Facts	
Specific Statistic	**Fact**
Location of the most frequent occurrence	United States, 145/year; Australia, 140/year.
Areas of most frequent occurrence in the US	From 600 miles east of the Rockies to 100 miles west of the Appalachians. Some occur on the west coast but no state in the continental United States is tornado-free.
Seasonal frequency	Any month; maximum, May; minimum, December.
Diurnal frequency	Any time; 80% between 1200 and 2100.
Speed of movement	From a few miles per hour to 150mph.
Speed of funnel wind	From less than 100mph to over 300mph.
Length of path	From 100 feet to 300 miles; average, 20 to 40 miles.
Funnel diameter	From a few yards to 2 miles.
Direction of movement	Depends on the movement of parent cloud; 90% move from between south and west.

Synoptic Requirements:	
Location and Stability	**Weather Requirement**
a. Surface	Large 1,002mb low with sharp trough or cold front with sharp wind shift, separating mP or cP from mT air. May occur anywhere in low; most frequent 100–600 miles SE of low center in warm sector. Must have a warm, moist tongue.
b. Aloft	Dry air over warm, moist air. Dry tongue at 700mb must intersect moist tongue of SFC. Lifting must occur. Winds aloft between 10,000 to 20,000 feet must have narrow band with 35knots speed or more.
c. Stability	Conditionally unstable; stability index of –4 or less.

Tornadoes *may* be forecast when the synoptic requirements are met and when the properties of the air fall within the ranges as listed under the synoptic properties that are in the table below. Under ordinary circumstances, *all* the synoptic requirements must be met for a large outbreak. Tornadoes can be forecast to appear first on the windward border of the moist tongue near the 55°F isodrosotherm. They are not likely to develop if the moisture pattern at 850mb is diffused or is displaced downstream compared with the surface moisture pattern. Tornadoes may develop where the 700mb dry tongue crosses over the lower moist tongue. A mixing ratio of 2 grams per kilogram (g/kg) at 700mb is usually sufficiently dry.

Synoptic Properties	**Minimum Value**	**Average Value**	**Maximum Value**
Depth of moist layer	2,600ft	5,000ft	8,800ft
Inversion (70%) or isothermal layer (30%)	820mb	—	750mb
LFC at	800mb	660mb	540mb
Humidity of the moist layer	65%	85%	100%
Humidity at top of inversion	—	50% or less	—
Humidity at 500mb	—	50% or less	—

The table below describes and defines the Fujita tornado scale. The late Dr. Theodore “Ted” Fujita, Professor of Meteorology, University of Chicago, developed the scale from his research on tornadoes.

Fujita Scale of Tornado Intensity
F0: Gale tornado (40–72mph): Light damage; some damage to chimneys; break branches off trees; push over shallow-rooted trees; damage sign boards.
F1: Moderate tornado (73–112mph): Moderate damage. The lower limit (73mph) is the beginning of hurricane wind speed; peels surface off roofs; pushes mobile homes off foundations or overturns them; pushes moving autos off roads.
F2: Significant tornado (113–157mph): Considerable damage. Tears roofs off frame houses; demolishes mobile homes; pushes boxcars over; snaps or uproots large trees; generates light-object missiles.
F3: Severe tornado (158–206mph): Severe damage. Tears roofs and some walls off well-constructed houses; overturns trains; uproots most trees in forest; lifts and throws heavy cars.
F4: Devastating tornado (207–260mph): Devastating damage. Levels well constructed houses; blows structures with weak foundations some distance; throws cars and generates large missiles.
F5: Incredible tornado (263–318mph): Incredible damage. Lifts strong frame houses off foundations and carries them considerable distance to disintegrate; automobile-sized missiles fly through the air in excess of 100mph; debarks trees; incredible phenomena occurs
F6: Inconceivable tornado (319–379mph): These winds are very unlikely. The small area of damage they might produce would probably not be recognizable along with the mess produced by F4 and F5 wind that would surround the F6 winds. Missiles, such as cars and refrigerators would do serious secondary damage that could not be directly identified as F6 damage. If this level is ever achieved, evidence for it might only be found in some manner of ground swirl pattern, for it may never be identifiable through engineering studies

NOTES:

1. The majority of reported tornadoes are of the F0 - F2 intensities.
2. F0 and F1: Weak tornadoes.
3. F2 and F3: Strong tornadoes.
4. 4. F4 and F5: Violent tornadoes.

If the moisture content of the lower layer is extremely high, the value at 700mb may be higher than 2g/kg. Tornadoes *may* be forecast to occur near the intersection of the axis of the wind maximum (between 10,000 and 20,000ft) with the windward edge of the moist tongue at low levels. Tornadoes occur *only* when appreciable lifting takes place.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

240. Atmospheric stability

1. Explain the cold adiabatic process.

2. Explain the warm adiabatic process.

3. What are the moist and dry adiabatic lapse rates in °C?

4. How is stability determined?

5. What is the stability of the atmosphere if an air parcel rises from the surface and remains colder than the environmental lapse rate?

6. What is the stability of the atmosphere if the environmental lapse rate is less than the moist adiabatic lapse rate?

7. What effect does cold air advection at the surface and warm air advection in the upper levels of the atmosphere have on atmospheric stability?

8. What is the stability of the atmosphere if an unsaturated air parcel rises at the same lapse rate as the environment?

9. What is the stability of the atmosphere if an air parcel rises from the surface and remains warmer than the environmental lapse rate?

10. What is the stability of the atmosphere if the environmental lapse rate is greater than the dry adiabatic lapse rate?

11. What condition exists if a rising saturated parcel of air has a lapse rate less than the dry adiabatic rate but greater than the moist adiabatic rate?

12. What is the stability of the atmosphere if a saturated air parcel rises and decreases temperature at a rate of 4.5° for every 1,000 meters?

13. When during a day does maximum instability usually occur?

241. United States seasonal variations of convective severe weather

1. What region of the CONUS has the most thunderstorm activity?

2. Where are severe thunderstorms most prevalent in the CONUS?

3. What months in the CONUS appear to be the best for the occurrence of tornadic activity?

4. What are the two air-mass boundaries in the CONUS that are conducive to forming severe weather?

5. How does CONUS seasonal activity shift?

6. What are the categories for non-severe and severe thunderstorms?

7. During the weather briefing, how should you brief an aircrew that has isolated thunderstorms projected along its route?

242. Basic conditions necessary for the development of convective severe weather

1. What are the five conditions necessary for the formation of severe thunderstorms, tornadoes, and their associated phenomena?
2. Explain the parcel theory.

243. Thunderstorm types

1. Why can air-mass thunderstorms result in local heavy rain?
2. What type of thunderstorm is often confused with a frontal passage?
3. When does climatology support prime activity of the dryline?
4. When do the most violent thunderstorms occur with a dryline?

244. Stages of non-severe thunderstorm development

1. What are the stages of non-severe thunderstorm development?
2. What non-severe thunderstorm stage is the most violent?
3. How does a hailstone form and become larger?
4. What causes a meso-high to form at the base of a downdraft?

245. Types and development of severe thunderstorms

1. Why is the presence of strong mid-level winds essential to developing severe thunderstorms?
2. What is “discrete propagation?”

3. How do strong winds aloft contribute to supercell severe thunderstorm development?

4. What is the Magnus effect?

5. What term refers to the supercell downdraft associated with the gust front?

6. What downdrafts are associated with the RFD?

7. What are the characteristics of the right cell of a splitting cell?

246. Tornado statistics

1. What time of day has the greatest frequency of tornado occurrences?

2. What is the highest wind speed ever measured in a tornado?

3. What area of the US has the greatest frequency of tornado occurrences?

4. What is the average length of a tornado path?

5. Where is the most common place for tornadoes to first appear in relation to the low-level moist area?

3–2. Air Masses and Synoptic Patterns Indicative of Severe Weather

Severe weather seems to occur most often in the Great Plains during the summer time, but it also occurs throughout the CONUS as well. As weather journeymen we must be able to forecast the onset of severe weather no matter where we happen to be stationed. Throughout years of studies, certain convective patterns and systems were noted to support severe weather outbreaks. A major step in forecasting these events is recognizing when the right volatile ingredients are present even though there isn’t a cloud in the sky. We begin our discussion with which synoptic patterns and air masses are conducive for possible severe weather.

It is understood that tornadoes form in weather conditions that are favorable to severe thunderstorm development. Additionally, when there is any implied reference to a tornado-producing air structure it also implies the occurrence of hail and destructive winds. Because of the study of these air structures, it has become generally accepted that four air structure types are responsible for most of the tornadic activity within the US. Although no geographical limitations are intended, most of these air structures are named after the regions in which each predominates.

247. Types of severe weather air masses

In this lesson we'll discuss four types of air masses that normally indicate severe weather: Type I, II, III, and IV. Each one is discussed in detail below.

Type I, Great Plains

The description of the Great Plains type of tornado-producing air structure is based on an accumulation of representative soundings. The average Type I air structure in which tornadoes have formed is shown in figure 3–28. The lowest point plotted is well above 1,000mb, several hundreds of feet above sea level, which reflects the frequency of this type of air mass over the Great Plains.

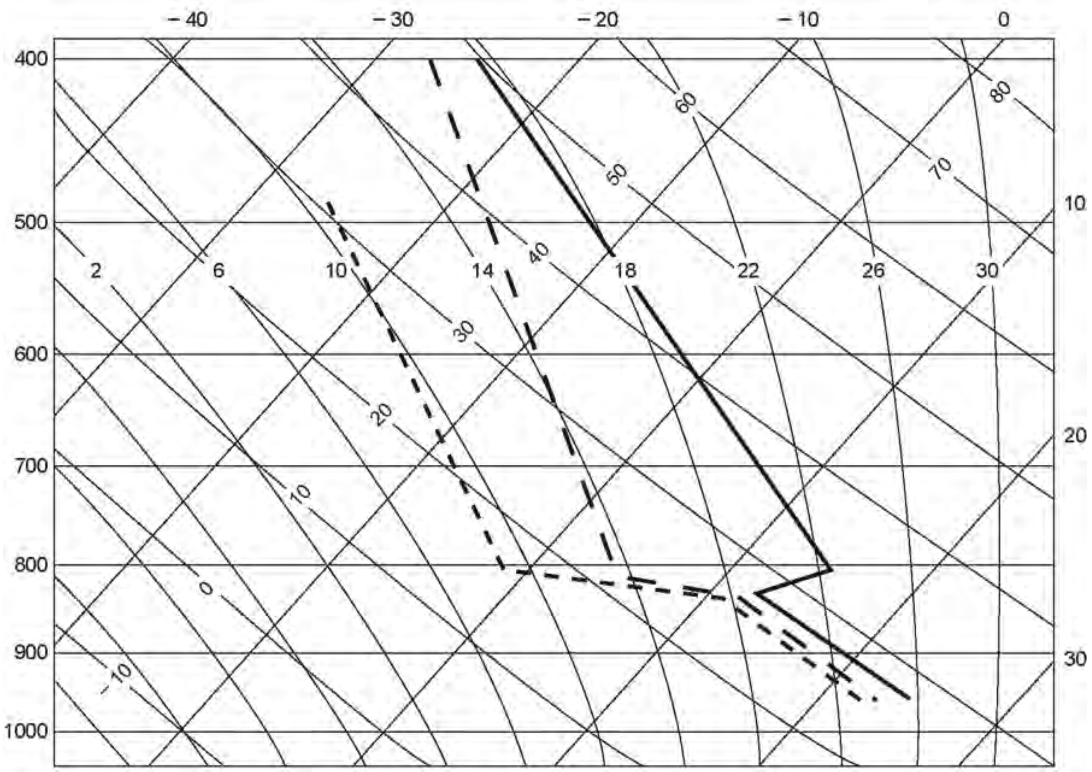


Figure 3–28. Example sounding of a Type I, Great Plains air mass.

NOTE: The following legend is for figures 3–28 and 29.

- The solid line represents the environmental temperature curve.
- The short-dashed line to the left is the environmental dew-point curve.
- The center, long-dashed line is the environmental wet-bulb temperature.

The temperature lapse rate is conditionally unstable in each of the two strata clouds, both of which are in the lower troposphere and separated by an inversion or stable layer. The atmospheric moisture is stratified with the lower layer being moist (i.e., relative humidity (RH) normally over 65 percent and surface dew point normally over 55°F) and very rapid drying is evident through the characteristic inversion. Above this inversion, the RH tends to increase slightly at first, then more rapidly above 550mb.

Winds increase with altitude in the dry air above the inversion, having a component of at least 30 knots perpendicular to the flow in the warm, moist air. The median wind shown by the representative soundings at 850mb is 30 knots from 219° and at 500mb it is 50 knots from 256°.

The air from the surface to 400mb is conditionally unstable and has a negative Showalter stability index. The lifted stability index is about –6 on the mean sounding. The vertical totals index is 28, the cross totals 26, and the total totals is 54.

Tornadoes in this type of air mass most frequently occur in families; their paths are commonly long and wide compared to tornadoes occurring in the other types of air masses. Tornadoes are more numerous in late afternoon, but can occur any time of the day or night. They are usually accompanied by widespread, destructive wind storms and large hail.

Morning stratus, temporary clearing, and middle clouds frequently precede tornadoes in the Type I air mass. Mammatus clouds are common with thunderstorms and are nearly always reported near those with tornadic activity. The temperature is normally high for the season so tornadic activity can occur any time of the day or night. The dew point sometimes rises very rapidly one to four hours before the storm, making the air oppressive. As the storm passes, the temperature drops very rapidly, then returns to normal unless the activity is along a cold front. Preceding winds are usually light to moderate. The pressure drops slowly for some hours and is followed by a brief small rise. Then, with the onset of the storm, the pressure plunges into a deep fall followed by a sharp rise and, in a few minutes, returns to normal following the passage of the thunderstorm cell. Generally, the weather sequence changes rapidly.

Type II, Gulf Coast

In contrast to the Great Plains air mass, tornadoes also form in a tropical type air mass that is moist up to great heights. Such storms are most common on the coast of the Gulf of Mexico. They produce the waterspouts so often reported over the Gulf Coastal waters. The average values of the available soundings representative of the Type II tornado-producing air are shown in figure 3–29. Note the very strong instability below 700mb.

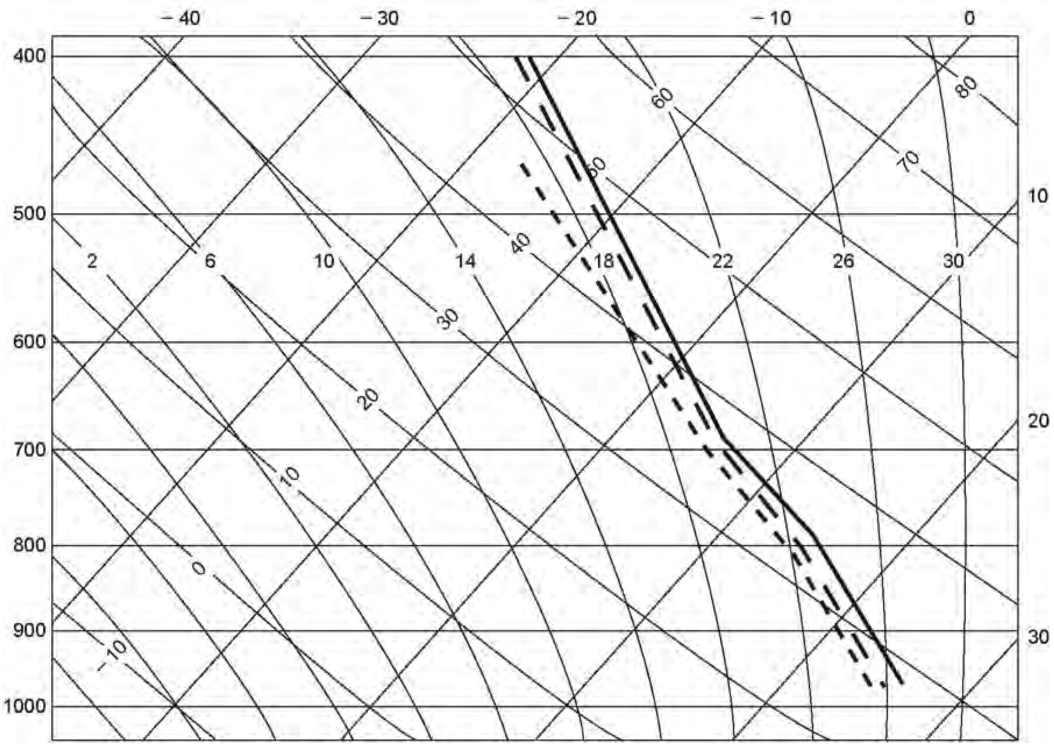


Figure 3–29. Example sounding of a Type II, Gulf Coast air mass.

The temperature lapse rate is conditionally unstable with no significant inversion or stable layer; surface temperatures are usually over 80°F. The moisture content is very high, the RH being over 65 percent in practically all cases from the surface to above 20,000 ft.

The winds normally decrease with altitude. There is no requirement for strong winds at any level in this air structure, though sharp wind shears do appear conducive to tornado formation. The wind at 850mb varies between extremes of 5 to 85 knots and at 500mb from 5 to 55 knots with the average direction veering about 30° between these levels.

The median values of both the lifted stability index and the total totals are the same as in tornado air mass Type I (–6 and 54 respectively), although the absence of an inversion has permitted the extreme cases to reach greater instability values.

Normally, in this type of air mass only one of the many thunderstorms in a given situation produces a tornado or waterspout. When more than one develops tornadic strength to the surface, 30 to 50 miles usually separate them. They are short lived, their paths brief and narrow, and they move more slowly than those in air-mass Type I. Although hail aloft does occur, surface hail and strong thunderstorm gusts are rare, since the wet bulb zero (WBZ) is normally above 11,000 ft.

The weather, both before and after the tornado occurrence, is normally cloudy with showers and scattered thunderstorm activity. There is neither a temperature nor a dew point discontinuity and only the pressure falls rapidly before the tornado. Otherwise, the weather changes slowly with time.

Type III, Pacific Coast

Tornadoes also form in comparatively cold, moist air. This air mass may be called the *Pacific Coast type*. It is responsible for the cold air-mass waterspouts of the West Coast. The average values of the soundings representative of the Type III tornado-producing air mass are shown in figure 3–30. Note the relative coldness of the air at all levels.

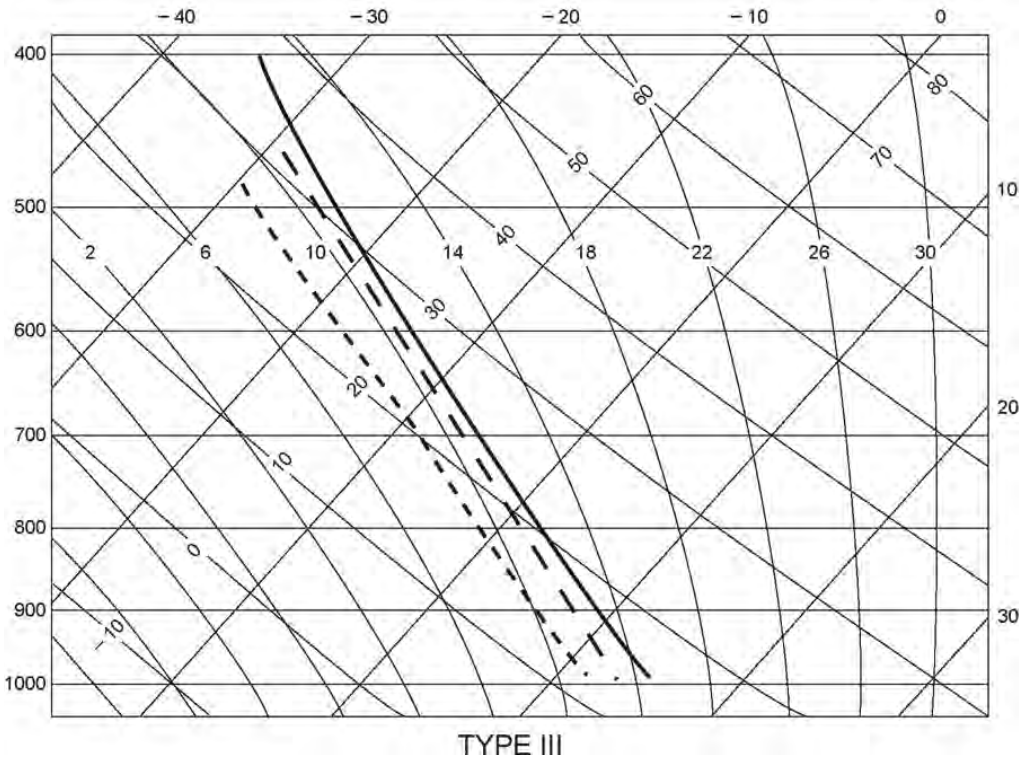


Figure 3–30. Example sounding of a Type III, Pacific Coast air mass.

The temperature lapse rate is conditionally unstable, without significant inversions or stable layers. Compared to Type II (Gulf Coast), the Type III air mass is cold and the surface temperatures range from 50°F to 68°F. Moisture extends to great heights, with the RH commonly exceeding 70 percent at all levels up to at least 500mb. The winds increase and generally veer with altitude, their average speeds being 15 knots at 850mb and 50 knots at 500mb. Apparent instability in this air mass is not as great as in the first two types of air masses discussed—the median values being –3. However, the total totals index is 57.

Generally, tornadoes of this type structure occur by themselves rather than in families or groups, although Mammatus, virga, and funnel clouds often form in the vicinity. The WBZ is usually so low that only small hail is reported and the thunderstorm gusts are often masked by strong gradient winds. Compared to those in tornado air-mass Type I, or even Type II, tornadoes in this situation have only a brief life, with a short and narrow path.

Tornadoes in this type air mass are normally found in a rather extensive cloudy area with scattered rain showers and isolated thunderstorms. The clouds are mostly stratocumulus with buildups embedded in the lowest deck. Also, Mammatus clouds are usually reported. The main cloud base from which the tornado appears usually remains well above the surface. Often, in this type air mass, the only evidence that a tornado existed is found in the apparent rotational pattern observed in surface debris after the storm passes. There are no abrupt or unusual changes in the weather elements except, of course, the pressure within the tornado itself. The Los Angeles basin is the primary hotspot for tornadoes west of the Continental Divide.

Type IV, Inverted “V”

When continental tropical air is overrun by maritime polar air at 5,000 to 8,000 ft above the ground, the inverted “V,” or Type IV, tornado sounding may result. This air structure, when triggered, is favorable for violent straight-line wind storms. The Type IV tornado-producing air structure is confined to the High Plains region to the lee side of the Rockies from western Nebraska, southward into Texas, and westward into the southwestern desert areas. Tornadoes seldom reach the ground with this type of air mass. When they do, the narrow rope-like funnel causes destruction over a comparatively small area. The presence of dry air in this structure, coupled with a favorable WBZ height, makes it a dangerous hail producer.

The lifted index for this sounding is not representative, since the lower layers are quite dry. The representative total totals index for this type sounding is 53.

The median type IV sounding is shown in figure 3–31. A dry lower layer typifies the sounding with cool, moist layers aloft and is conditionally unstable. The WBZ height is near the optimum of 8,000 ft above the ground. The vertical wind profile increases in speed and veers with height. This type tornado is associated with a mid-level trough, which acts as the triggering mechanism. Rapid weather changes and Mammatus associated clouds are characteristic. This radiosonde observation (RAOB) type produces many more occurrences of microburst wind damage than actual tornado damage, especially over and west of the Rockies. When tornadoes occur with this type, they are isolated in nature, usually rapid-moving, short-lived, and their path is short and narrow.

Each of these tornado-producing air structures possesses the capability of spawning severe weather by virtue of thermal structure, humidity distribution, and, therefore, generates stability considerations. Monitor evaporational cooling and upper-level (800/900 - 300mb) lapse rate changes closely. Remember, however, that a sounding representative of an air structure is largely time dependent; that is to say that the structure represented by the sounding may be significantly altered through the influence of the various processes and mechanisms operative in the atmosphere. The sounding reveals the *capability* of an air mass for severe weather production, but this capability is realized only if the structure is made highly unstable. You can *not* base the forecasting of severe weather, therefore, on the sounding alone, but must also depend on an analysis of the conditions and mechanisms that can alter the air structure in question.

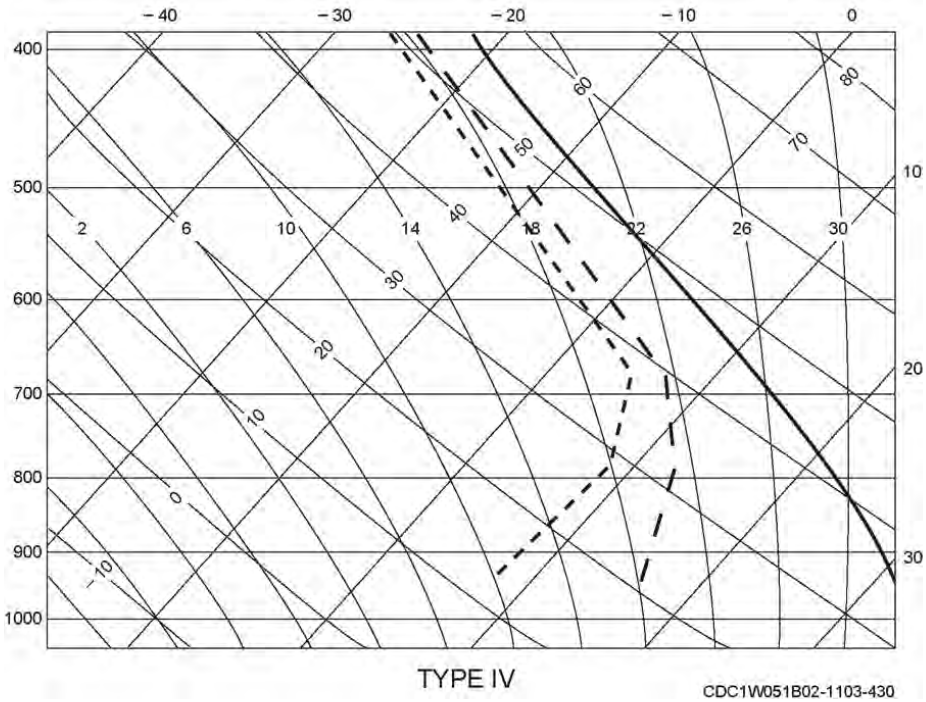


Figure 3–31. Example sounding of a Type IV, High Plains air mass.

248. Types of severe weather producing synoptic patterns

In this lesson, you're introduced to five different synoptic patterns that are favorable for convective severe weather. They are: dryline, frontal, overrunning, cold core, and major cyclone. Each one is discussed in the following paragraphs.

Type A – dryline

The dryline pattern is a narrow, almost-vertical zone, across which a sharp moisture gradient exists. Even though the dryline is often not collocated with the zone of maximum surface convergence, it serves as a focus point for convective activity. The characteristics of the dryline synoptic pattern include a well defined:

1. West to southwest maximum wind band aloft (500 to 300mb).
2. Southeasterly flow of moist air from the Gulf of Mexico at the low levels (surface to 850mb).
3. West to southwesterly flow of dry air from high plateau regions of the southwest US and northern Mexico at the low-and mid-levels (surface to 700mb) that are positioned upstream from the low-level moisture ridge.

There is considerable convergence along the boundary between the moist southeast flow and dry southwest flow of air. Speed convergence is also commonly found upstream from the dryline in the dry air.

The dryline is often associated with a pressure trough and/or wind shift line although neither is necessary for the existence of a dryline pattern. Vertically, the dry air exhibits a nearly dry adiabatic lapse rate, while an inversion "caps" the moist air. The dryline synoptic pattern is normally associated with a Type I air mass.

Location techniques

It is necessary to use a moisture variable to locate a dryline pattern. You can use mixing ratio, dew-point temperature, or equivalent potential temperature to identify the transition zone between the moist and dry air.

For analysis of a dryline, the 55°F isodrosotherm is recommended as the first estimation. This is the lowest value that seems to support tornadic thunderstorms. A dew-point temperature difference greater than 10°F degrees should be present across the dryline for a dryline pattern to develop.

Satellite imagery can also be used to help locate a dryline. Visible images often depict a thin line of convective clouds above a dryline. Infrared imagery often shows a boundary of "black stratus" at the dryline. This nocturnal observation east of the dryline is due to the fact that the moisture content has kept temperatures warmer than those in the dry air to the west.

Movement

A dryline movement is generally much faster than advection would support. One mechanism that accounts for the rapid movement is vertical turbulent mixing of the moist surface with the dry air aloft.

After sunrise, boundary layer mixing starts as surface temperatures rise in response to solar insolation. West of a dryline, a surface layer with an adiabatic lapse rate rapidly replaces a nocturnal radiational inversion. Any moisture trapped beneath the inversion would be freely mixed into the dry air aloft.

The greater the depth of the moist layer, the greater is the degree of heating required to break the capping inversion. Since the general terrain of the southern plains slopes downward from west to east, the depth of the moist layer also increases eastward from the dryline location.

When the needed amount of heat is absorbed, the low-level moist air mixes with the dry air aloft. The surface dew-point temperature drops rapidly and the dryline "leaps" eastward to a position where no appreciable mixing between the air masses has occurred. Dryline bulges often form due to enhanced dry air and the vertical mixing of high-momentum mid-level air to the surface. Enhanced convergence occurs at the bulge. Use the strongest 700mb winds to forecast the probable location of the dryline bulge.

In the late afternoon and evening, the dry air cools rapidly. A nocturnal inversion forms west of the dryline. The inhibition of vertical mixing (due to the inversion) leads to a decrease of the low-level winds in the dry air. East of the dryline, strong easterly flow continues resulting in a net easterly wind across the dryline. The dryline is advected westward by these winds. The entire diurnal process can reoccur the next day.

Activity

Very severe thunderstorm cells or lines form along a dryline. The most violent thunderstorms occur at the intersection points between a dryline and another boundary. Large hail, damaging winds, and tornadoes are common. Activity is mainly limited to the late afternoon and early evening, although severe thunderstorms can develop along a dryline during its nocturnal retreat.

Type B – frontal

Anytime a cold front advances into a warm air regime, some instability results. Of course not all cold fronts have associated severe weather with them, but the frontal type weather pattern is the most predominant severe weather producer in the CONUS.

Characteristics

A frontal pattern features a well-defined flow of moist air at the low levels (usually associated with a low-level jet or maximum wind band) along with a well-developed surface baroclinic low with associated cold and warm fronts. Dry air located behind the cold front is most favorable. A well-defined mid-level (700mb) dry air intrusion from the west to southwest is also favorable. A strong, southwesterly maximum wind band aloft (500–300mb) with a major short-wave trough to the west also exists. A Type I air mass with a capping inversion usually exists.

Activity

The most violent activity occurs where a squall line intersects a warm front or outflow boundary (meso-low and LEWP formation). Activity mainly occurs during the late afternoon and early evening (maximum heating), but can occur at any hour due to the mechanical lifting associated with the fronts.

Large hail, damaging winds, and violent, long-lasting tornadoes are common. The Type B synoptic pattern is responsible for the major tornado outbreaks that occur in the US.

Type C – overrunning

Another synoptic scale severe weather pattern associated with fronts is the warm frontal overrunning pattern. This pattern features warm, moist, unstable air overrunning a stationary or warm frontal boundary.

Characteristics

The strongest overrunning occurs where the 850mb maximum wind band intersects the frontal boundary. The pattern also features a well-defined westerly maximum wind band aloft (500mb to 300mb), north of, and parallel to, the frontal boundary. A major short-wave trough is often embedded in the westerly wind flow and leads to further destabilization and or enhancement of frontal activity. A well-defined mid-level (700mb) dry-air intrusion is located upstream from the area of strongest overrunning.

Activity

Thunderstorms remain below severe limits until the mid-level dry air intrudes into the threat area. A squall line frequently forms along the leading edge of the dry air. The squall line is usually associated with a strong gust front (outflow boundary) that leaves a well-defined meso-high in its wake.

Hail is common due to the low freezing level. Strong, gusty winds occur with the passage of the gust front. Tornadoes are not common due to the cool air at the surface, although they can occur. The severe weather activity is strongest during maximum heating, but can occur at any hour.

Type D – cold core

This pattern features a nearly vertically stacked, occluded system. The threat area is not associated with any fronts, but rather, behind the cold front in the clear air.

Characteristics

Cool mP air is located behind the cold front and under a 500mb cold pocket aloft. Surface heating is abundant with relatively clear skies associated with the dry slot located behind the cold front that aids in the additional destabilization of the atmosphere. The low-levels are dominated by boundary level convergence and, in the strongest cases; a low- and mid-level dry-air intrusion exists upstream. The air mass is typically Type III and shows maximum potential if we use the Total Totals index (the Total Totals reacts to the 500mb cold pocket).

Activity

Intense thunderstorms form shortly after noon and normally dissipate at sunset. Funnel clouds (cold-air funnels) are common with brief, but occasional, tornado touchdowns. Hail is very common due to the low freezing level. Hail normally increases in quantity and size near the cold pocket.

Type E – major cyclone

This pattern is also associated with an occluded system. However, the threat area is ahead of the warm front, much like Type C.

Characteristics

Strong, low-level overrunning of warm, moist, unstable air at 850mb over a surface warm front is a characteristic. Dry-air intrusion exists at the mid levels, along with strong cold-air advection located upstream of the strong overrunning. Stability indices do not show the full potential of the air mass to support convection due to the cool low-level air north of the warm front.

Activity

The activity becomes severe when the dry-air intrusion and cold-air advection enter the threat area. Hail is the most common severe weather phenomenon, although tornadoes do occur with moderate frequency. The activity is the strongest during maximum heating time, but can occur at any hour. This type of severe weather pattern often forms in conjunction with a Type B or Type C outbreak.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

247. Types of severe weather air masses

1. Match the air mass descriptions in column A to the types in column B. Items in column B are used only once.

Column A

____ (1) The lapse rate is conditionally unstable with no significant inversion or stable layer. The surface temperatures are usually over 80°F. The moisture content is very high with the RH being over 65 percent from the surface to above 20,000 ft in practically all cases. The WBZ is normally above 11,000 ft. Tornadoes are short-lived, with narrow paths, and they move slowly.

____ (2) The lapse rate is conditionally unstable. The sounding has a dry lower layer and a cool, moist layer aloft. The WBZ is near the optimum of 8,000 ft. The winds aloft increase in speed and veer with height. Tornadoes are isolated, rapid moving, short-lived, and their path is short and narrow.

____ (3) The lapse rate is conditionally unstable; The air mass is colder than the other types and the surface temperatures range from 50°F to 68°F. Moisture extends to great heights with RH often exceeding 70 percent at all levels up to at least 500mb. Tornadoes in this type air mass have only a brief life, with a short narrow path.

____ (4) This air mass has an inversion or stable layer with a conditionally unstable lapse rate above and below the stable layer. The layer below the stable layer is moist, with a comparatively dry layer above. The moisture increases slowly above the stable layer. The winds increase with altitude in the dry air above the inversion, having a component of at least 30 knots perpendicular to the flow in the warm moist air. The weather changes rapidly, and tornadoes in this type of air mass have the longest life.

Column B

a. Type I, Great Plains.
b. Type II, Gulf Coast.
c. Type III, Pacific Coast.
d. Type IV, Inverted "V".

248. Types of severe weather producing synoptic patterns

1. What kind of air mass is normally associated with the dryline synoptic pattern?

2. What is recommended as a first estimation of the location of the dryline?

3. What dew-point temperature difference should be present across a dryline?

4. Where do the most violent thunderstorms occur with dryline activity?

5. What type of weather pattern is the most predominant severe weather producer in the CONUS?

6. Where does the most violent weather activity occur with a type B pattern?

7. What can lead to enhancement of frontal activity with a type C pattern?

8. Why is hail very common with a type D pattern?

3–3. Convective Patterns and Systems Encountered during Severe Weather Seasons

During the warm season from late spring through early fall, forecasters must place a higher degree of emphasis on the meteorological processes that take place within the lower levels of the atmosphere. Low-level warm advection interacting with surface features such as weak frontal systems and outflow boundaries become, by far, the most prominent features responsible for significant thunderstorm development. This often occurs without strong vorticity maxima and/or upper-level cold air advection to initiate the outbreak. The following lessons cover convective patterns that exhibit these traits.

The following material highlights three types of damaging warm season weather events—northwest flow outbreaks, the derecho, and mesoscale convective complexes (MCC).

249. Northwest flow outbreaks

As the summer season approaches, it is not unusual for outbreaks of severe thunderstorms to occur in areas east of a prominent upper-level ridge where the upper-level flow is fairly weak and northwesterly. We must place a higher degree of importance on the thermodynamic properties that take place within the lower troposphere in terms of conditional instability and low-level warm air advection in the presence of northwest flow (NWF) aloft.

J. C. Galway (1958) and Miller (1972) recognized the NWF scenario. Galway noted that often a short-wave trough that initially appeared minor, but deepened as it approached the long-wave trough position triggered an outbreak. Miller noted that some of the most destructive severe weather outbreaks in the summer are associated with west-northwest to northwest flow in the mid-troposphere.

By 1982, a comprehensive study of severe storms in northwest flow was compiled and documented by Robert Johns of the former National Severe Storms Forecast Center (NSSFC). The investigation, which covered 16 years and 163 outbreaks, showed conclusively that NWF events frequently occur during midsummer from the Plains states to the Mid-Atlantic States. A long-wave ridge west of the outbreak area and a long-wave trough to the east characterizes these events.

An NWF outbreak is characterized by the following three conditions:

1. The average 500mb flow direction near the event is from a direction of 280° or greater.
2. The mean ridge on the 500mb product immediately following the event remains upstream.

3. The mean ridge on the 500mb product immediately preceding the event is upstream of the geographical location of the severe thunderstorm event. The mean trough is downstream.

A 500mb flow of 280° is the most common direction before the beginning of the outbreak. June, July, and August account for 85 percent of the average annual total of NWF outbreaks with July being the peak month.

Geographical frequency

There are two well-defined frequency axes as we see in figure 3–32. The first extends from eastern North Dakota to southwestern Pennsylvania. This corridor is established in June and remains nearly stationary through August. The highest rate of tornado occurrences is associated with this corridor.

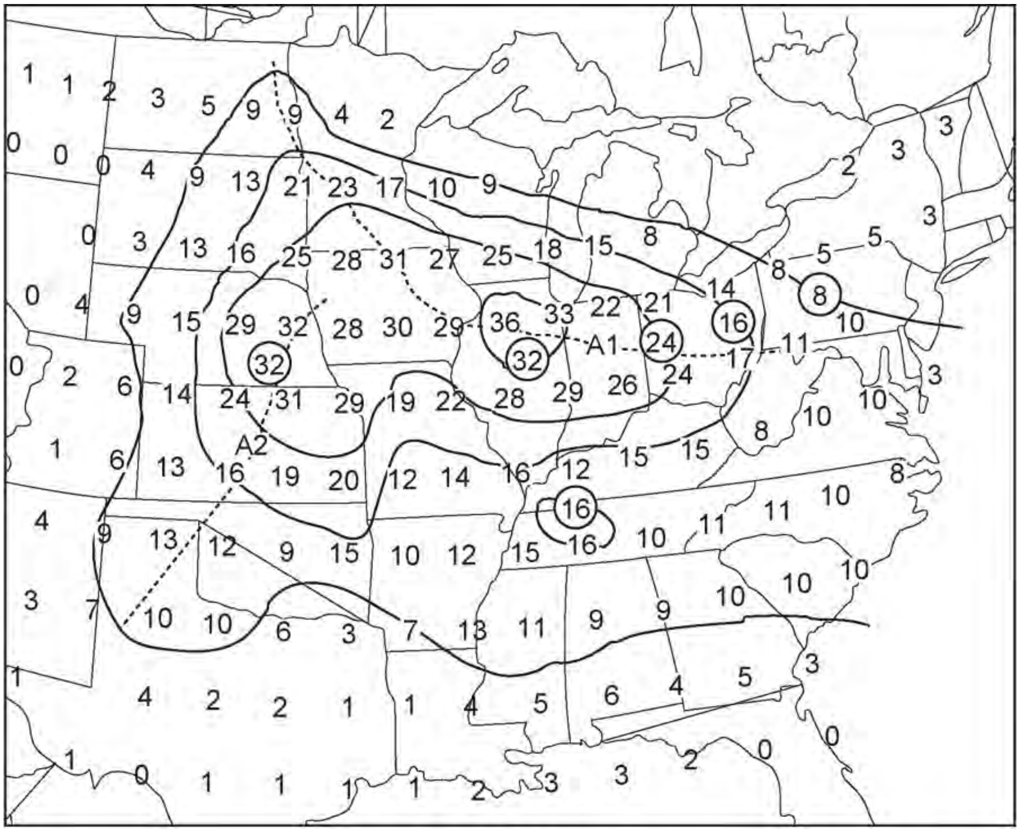


Figure 3–32. Total number of NWF outbreaks for the period of 1962-1977.

The second axis extends from the Texas panhandle into Iowa. This corridor is usually established in June with the highest frequencies shifting northward along the axis as the summer progresses. Small frequencies were found along the Gulf coastal states, central Missouri, and northern New England.

Diurnal distribution and duration

Northwest flow outbreaks typically develop in the afternoon when the effects of solar heating are greatest and then diminish within a few hours after sunset. About one fourth of the NWF outbreaks continue into the early morning hours. Early season outbreaks tend to last longer than those later in the season, with the average duration decreasing from near 10 hours in late May to 8½ hours in August. Major NWF outbreaks are most likely to occur during the months of July and August. It is during this period that the threat of damaging storms is at its highest with tornado occurrences near 30 percent.

Miller noted that as the severe weather season progresses into summer, tornado occurrences decrease while the frequency of straight-line damaging wind storms increases proportionally. Miller also observed that NWF outbreaks are likely to be repeated on several successive days.

The highest threat of repeat outbreaks occurs during the months of June and July. One explanation behind successive outbreaks is that the surface and 500mb pattern remains quasi-stationary with a series of short waves triggering individual outbreaks. These short waves induce backing of the low-level wind field, which, in turn, increases the warm air advection pattern required for a substantial outbreak.

Upper-air patterns

Data extracted from the 500mb, 850mb, and surface products combined with stability indices and synoptic scale wind shears provide strong guidance to alert the forecasters of potential NWF outbreaks.

500-millibar parameters

An average flow angle of 280° is the most common direction before the start of an NWF outbreak. Most outbreaks have an average flow angle ranging from 280° to 300°. When the average flow angle exceeds 310°, the chance of an outbreak is greatly reduced. NWF outbreaks, for the most part, are located to the south (on the anticyclonic shear side) of the 500mb jet. There is usually little 500mb temperature or height change over an outbreak area. Sometimes, the temperature or height may rise due to the release of latent heat of condensation.

Over 2/3 of all NWF outbreaks are associated with 500mb wind maxima exceeding 50 knots. When the 500mb temperature exceeds –6°C, the threat of any organized outbreak is small to nonexistent. Regional variations of 500mb parameters associated with NWF outbreaks are shown in figure 3–33.

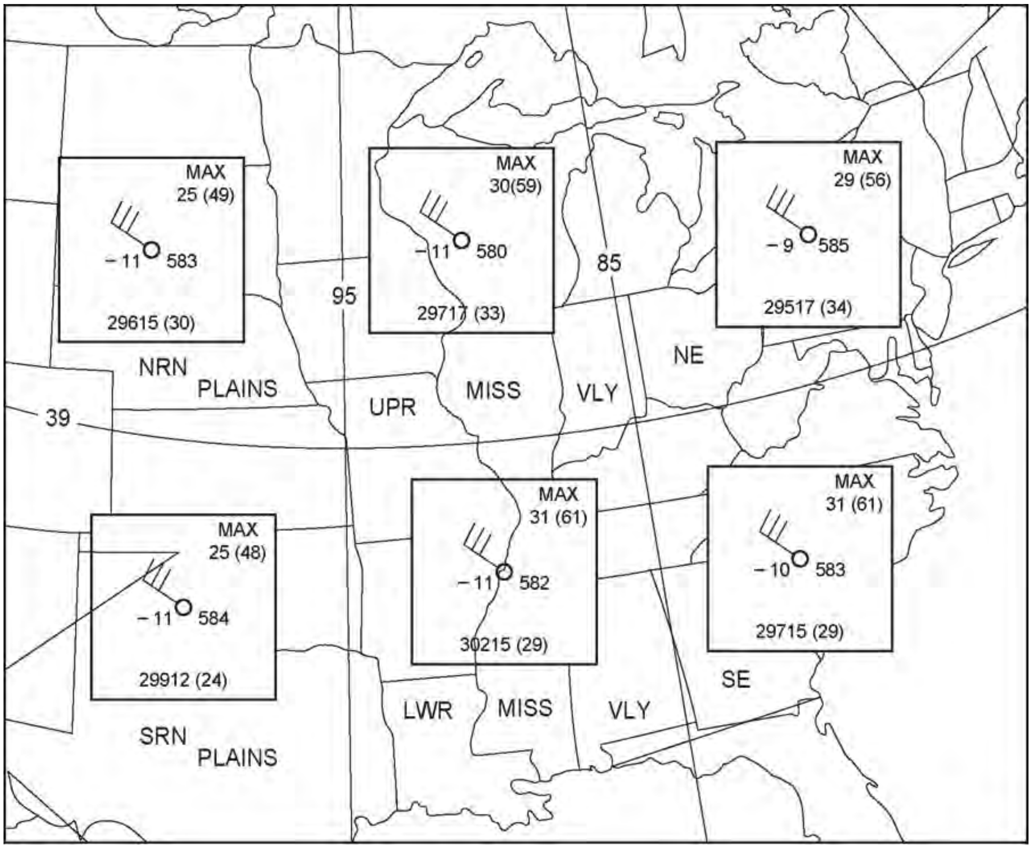


Figure 3–33. Regional variations of the 500mb parameters associated with NWF outbreaks.

850mb parameters

Areas that display the highest rate of warm air and moisture advection are susceptible to outbreaks. In the presence of weak upper-level dynamics, all NWF thunderstorms require high rates of low-level warm air advection and a high degree of conditional instability in the genesis area. Most of the NWF outbreaks are associated with 850mb temperatures exceeding 20°C and dew points greater than 12°C. The 850mb wind direction is typically south-to-southeast over the Plains states west of 95°W and west-to-northwest east of 95°W. Regional variations in the 850mb parameters associated with NWF outbreaks appear in fig 3–34.

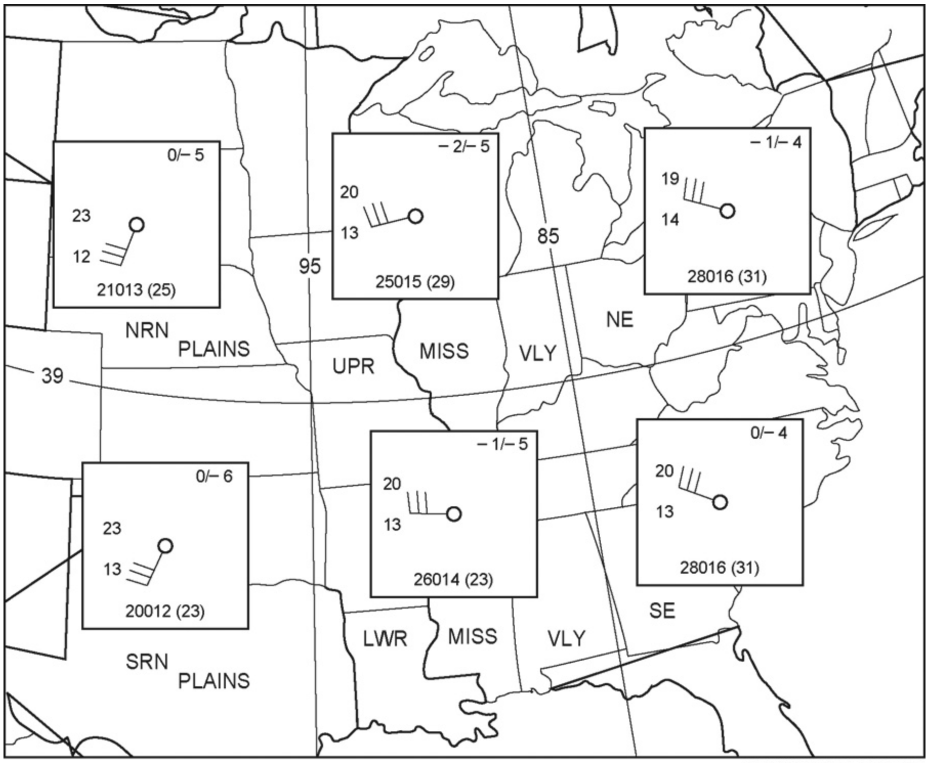


Figure 3–34. Regional variations in the 850mb parameters associated with NWF outbreaks.

Stability

Look for the following stability thresholds which will indicate an increased chance of a NWF outbreak:

- The average total totals index is greater than 54.
- The average Showalter index is lower than –3°C.
- The average surface-based lifted index (SBLI) is lower than –6°C.
- High degree of conditional instability resulting from moisture pooling and warm air advection.

Surface parameters

Consider the amount of warm air and moisture advection. Also consider the amount of solar heating and, to a lower extent, evapotranspiration from vegetation. Low-level convergence fields, along with the knowledge of where the highest rate of warm air and moisture advection is taking place, allow forecasters to know where to look for deep convection. Early in the season, look for surface temperatures around 28°C with dew points exceeding 19°C. By July, the average surface temperature rises to 30°C with the associated dew points rising to 21°C. Look for sea-level pressures from 1,007mb to 1,015mb and southeasterly flow over the Plains states and southerly-to-southwesterly flow east of 95°W. Figure 3–35 shows regional variations in surface parameters associated with NWF outbreaks.

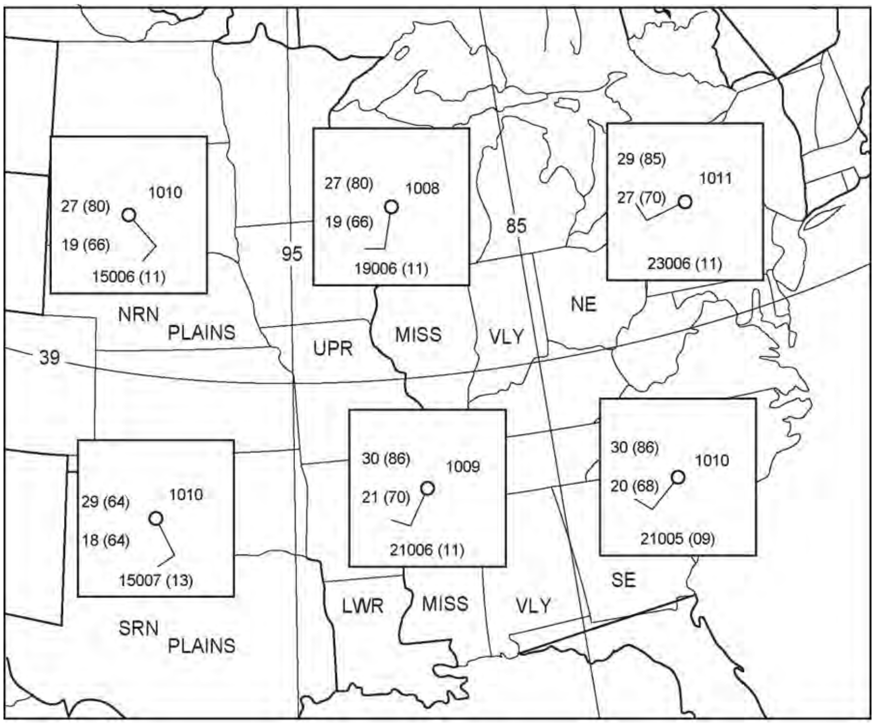


Figure 3–35. Regional variations in surface parameters associated with NWF outbreaks.

Surface patterns

Two basic patterns are associated with NWF outbreaks. The first is a northwest-to-southwest orientated pressure trough or cold front. Most of the NWF outbreaks that form in the Plains states are associated with drylines and cold fronts. The second involves a quasi-stationary front oriented in a west-northwest-to-east-southeast direction. NWF outbreaks that form with this pattern often affect the upper Mississippi valley, and to a lesser degree, the southern CONUS and the lower Mississippi valley.

As the summer season approaches, the frequency of this type of outbreak decreases to near zero over the southeastern CONUS and the lower Mississippi valley while the frequency increases over the upper Mississippi valley. A southerly low-level jet often bisects outbreaks that affect portions of the central CONUS in the presence of a west-northwest-to-east-southeast quasi-stationary frontal boundary. During the summer months, this pattern can lead to rapid and explosive thunderstorm development for several reasons:

1. There is a strong tendency for low-level moisture to pool along and north of the frontal boundary. As enhanced low-level warm air advection interacts with the moisture axis, a high degree of conditional instability results.
2. When a low-level jet bisects this type of frontal boundary enhanced upward vertical velocities and moisture convergence develops along and to the north of where the intersection took place.
3. When the upper-level flow has a northwest-to-west component above a southeast-to-southerly low-level wind field, high values of directional shear result.

Most severe summertime thunderstorms, especially those associated with NWF, have large values of directional shear and a small amount of speed shear. Severe thunderstorms are usually self-perpetuating storms. For this to happen, there must be significant environmental directional shear to allow for a continuous and strong low-level inflow of conditionally unstable air.

250. The derecho

A type of destructive summertime mesoscale system associated with NWF outbreaks is known as a derecho. The term, pronounced day-ray-cho, is Spanish for “straight” or “direct.” A derecho is a rapidly moving extratropical convective system known to produce widespread significant straight-line wind damage via a family of downburst clusters.

There are two primary types of derechos. The first is associated with a rapidly propagating segment of an extensive squall line that has developed ahead of a strong cold front that was driven eastward by a deep and fast moving low-pressure system. Outbreaks associated with this type typically occur in late winter and spring.

The second type develops in the late spring and summer in regions where the upper-level flow is westerly-to-northwesterly. Storms of this type typically develop in association with weak frontal systems, especially those that are quasi-stationary and are oriented in a general west-to-east direction.

Warm season derechos can resemble the size and appearance of Mesoscale Convective Complexes or squall lines. Often, the infrared (IR) satellite image depicts a large convective canopy with a very tight leading edge gradient. Severe downburst winds are closely associated with storms that get a leading edge gradient. The derecho often resembles a LEWP with the crest of the wave at the north end of a bulging squall line (fig. 3–36).

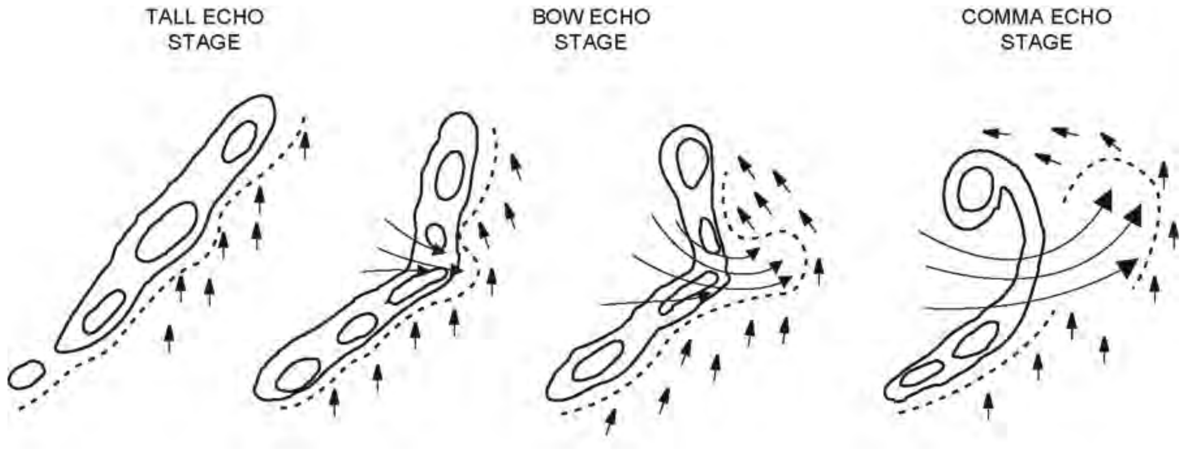


Figure 3–36. A model of the evolution of the bow echo.

The squall line itself follows the late Dr. Theodore Fujita’s observation and definition of a bow echo. The development of a strong low-level reflectivity gradient near the leading edge of a bow echo is often observed. Maximum echo tops are usually displaced over and occasionally ahead of the low-level reflectivity gradient (fig. 3–37).

The two most distinctive common physical characteristics of a derecho are the radar configurations and the rapid rate of propagation in a direction slightly to the right of the mean wind vector. It is this rapid rate of propagation (derechos can move at speeds exceeding 50 knots) that catches forecasters by surprise, and sometimes, results in a severe weather corridor spanning over 1,000 nm.

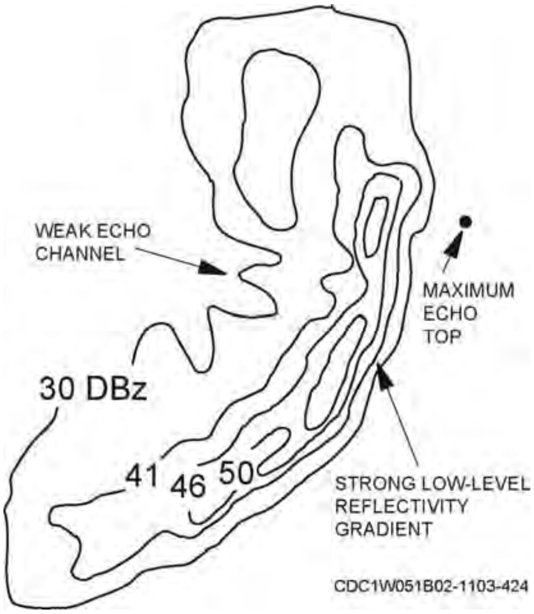


Figure 3–37. An example of a distinctive bow echo.

Surface and upper-air parameters

With derechos moving in excess of 50 knots, it is important to identify them as quickly as possible. To identify derechos, you need to look at several different parameters in both the lower and upper levels.

Surface and lower levels

Derechos usually develop along and north of, an east-to-west oriented quasi-stationary boundary, where there is significant low-level warm air advection and convergence. Once they develop, derechos move eastward or southeastward with the bulging squall line moving rapidly along the southern edge of a pronounced thermal ribbon and eventually into the warmer air mass. Derechos require high values of low-level moisture and result in very high values of conditional instability. This is an absolute requirement for the maintenance or life span of a derecho.

Typically, the surface-based lifted indexes (SBLI) range from –7 to –13. Occasionally, the surface dew points reach or exceed 80°F at various locations along the derecho’s track, which is indicative of significant moisture pooling. Frequently, the surface and 850mb analysis shows this moisture pooling extending through the 850mb level and in a convergence zone associated with the quasi-stationary boundary. Moisture pooling serves to increase buoyancy and lower the LFC; thus, less lifting is required for a parcel to become positively buoyant. This is one very important parameter that allows for the rapid movement and/or propagation of fast moving squall lines.

On the rare occasions when buoyancy values are small, there must be substantial vertical wind shear to support the rapid movement of a derecho. If there is significant moisture pooling in the presence of strong low-level to mid-level vertical wind shear, the threat of tornadic supercell development increases substantially.

Upper levels (500 to 200mb)

A rather flat 500mb ridge situated very near or just west of the derecho genesis region characterizes the upper levels. The primary polar jet is located well to the north with anticyclonic shear over the outbreak region. There are occasions when a separate wind maximum ranging from 40 to 55 knots, in a typical summertime scenario, is in fairly close proximity to the genesis region. In these situations, stronger upper-level kinematic processes interacting with high values of instability and enhanced low-level upward forcing can result in a very intense and long-lived derecho episode.

The one upper-air feature involved in all outbreaks is a 500mb short-wave trough. More often than not, the short wave is fairly weak with little or no cold air advection at 500mb. In all events, the upper-level pattern over the outbreak area indicates high-level anticyclonic shear and, frequently, pronounced difluence.

251. Mesoscale convective complex

The mesoscale convective complex (MCC) as defined by R. Maddox (1983) is characterized by:

1. Cloud shield temperatures at least as cold as –32°C.
2. Horizontal cloud tops extent of at least 100,000km^2 (about 70 percent of the size of the state of Iowa).
3. An interior cloud top region of temperatures colder than –52°C and covering at least 50,000km^2 (fig. 3–38).

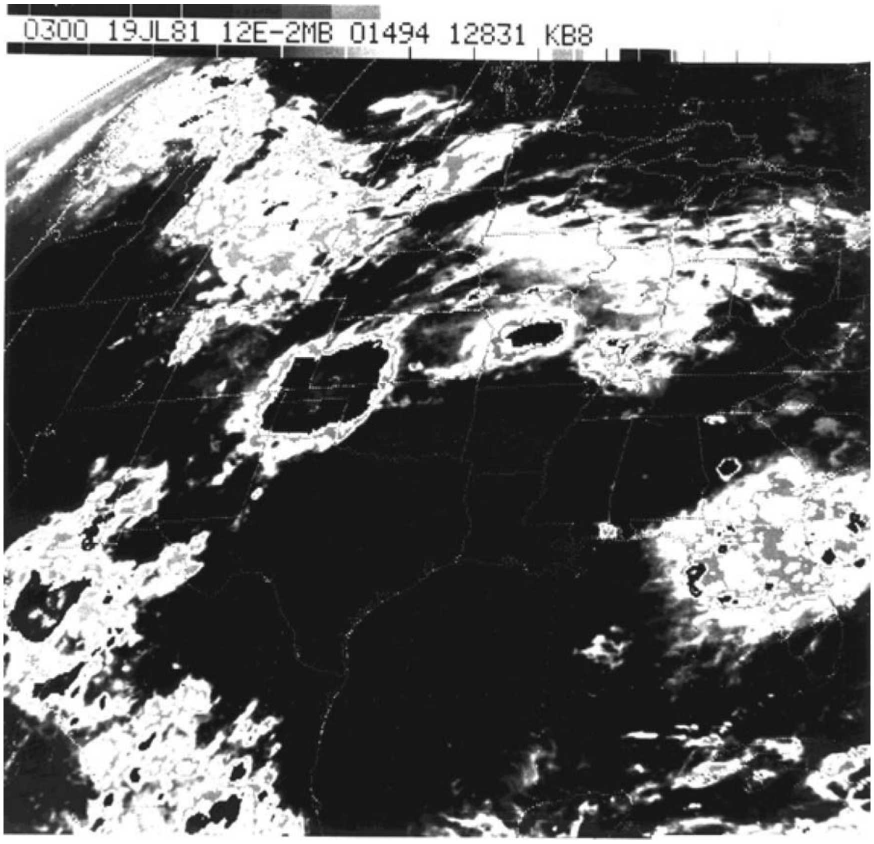


Figure 3–38. Example of an MCC over TX and OK panhandles, southwest KS, and southeast CO at 300Z 19 July 1981.

From late spring to early fall, MCCs pose a substantial operational forecast problem because they produce significant widespread weather. A widespread area of very heavy rainfall replaces severe thunderstorms common during the MCC genesis phase; this results in flash flooding as these complexes mature. Its long-lived nature and tremendous aerial extent make the MCC the largest and most problematic convective phenomena faced by the forecaster during the warm season.

Most MCCs have been found developing ahead of very weak short-wave troughs that display little 500mb thermal or vorticity advection. During the summer months, PVA at 500mb is sometimes not closely related to significant thunderstorm outbreaks. Synoptic scale forcing in MCCs is often very weak in the mid- and upper-troposphere. The primary upward forcing and maintenance mechanism is dominated by a strong, low-level, warm air advection pattern often associated with a low-level jet.

This process provides the growing thunderstorm complex with a persistent influx of very moist unstable air, along with the lifting needed to generate new convection.

Forecasters must also be concerned with the decay of an MCC because it can strongly control the next afternoon's convective activity by leaving behind widespread regions of cloudiness, rain-cooled air, and pronounced outflow boundaries. These outflow boundaries, along with mid-level vorticity centers (weak as they may be) and upper-level jet streaks are a few of the key features to look for as you forecast post-MCC thunderstorms.

The mesoscale convective complex life cycle

The MCC has four stages in its life cycle: genesis, development, mature, and dissipation. Once we discuss each one you'll understand the differences between them and be able to describe each of them.

Genesis stage

In the genesis stage, several individual thunderstorms develop within a region where atmospheric conditions are favorable for convection. Small-scale effects, such as topography and localized heat sources, play an important role in initial storm development. As thunderstorms continue developing, the latent heat released begins to develop a region of anomalous warming in the mid-levels. Thunderstorms often produce severe weather during this phase.

At mid-levels, entrainment of potentially cool environmental air produces strong evaporationally-driven downdrafts. This allows for the development of an outflow boundary along with intensification of the associated bubble high.

Development stage

In response to thunderstorm-induced warming, the larger scale environment begins to respond and a layer of mid-tropospheric (750mb to 400mb) convergence develops with inflow at the surface. Gust fronts and outflows from the individual storms merge to produce a large cold air outflow boundary driven by the meso-high. Strong low-level inflow of moist, unstable air continues and the system grows rapidly.

The most intense convective elements occur along the convergence zone produced by the interaction of the outflow boundary and the low-level inflow. Eventually, this region becomes saturated and exhibits a moist adiabatic, warm core structure compared with the surrounding environment.

Mature stage

Intense convective cells continue to form in a region where low-level inflow provides very unstable air to fuel the system. As the mature stage progresses, the threat of severe thunderstorms diminishes. The primary threat becomes heavy rain and wind gusts from convective cells occurring within a moist environment characterized by weak vertical wind shear. These cells become very efficient precipitators causing flash flooding to be common with this stage.

Dissipation stage

This stage is marked by a rapid change in the character of the MCC. It begins when intense convective elements no longer develop. This occurs when the system's energy supply is cut off or modified because of ingesting dry stable air. As this process continues, the MCC starts showing a less organized appearance in IR imagery. Even when the dissipation process is completed and the complex loses its organization, the cool air and attending outflow boundary at the surface must still be metwatched for redevelopment during peak heating. In conditionally unstable environments, the redevelopment of convection at the outflow boundary can produce severe thunderstorms.

Lightning activity associated with mesoscale convective complexes

Lightning flash rates are comparable to the highest observed rates within other mesoscale systems, four times those observed in severe or multicell storms in Florida, and more than 20 times the rates previously observed in isolated thunderstorms. Lightning damage occurs with half the MCCs and is most frequent between the development and the mature phase of their life cycle.

The average peak of ground discharge rates occurs more than 1½ hours after the first storms take shape and a little over 2½ hours before maximum cloud shield extent. Usually, peak ground discharge rates occur when cloud temperatures are the coldest. More specifically, increasing flash rates occur with the rapid expansion of the anvil cloud with a temperature colder than –52°C. It is during this period that the rainfall rates are the highest. The typical MCC produces cloud-to-ground lightning flash rates exceeding 1,000 per hour for nine consecutive hours.

Mesoscale convective complex facts

MCCs can develop in a variety of large-scale environments and are responsible for much of the beneficial rains and most of the severe weather that occurs in the CONUS during the spring and summer months. Model guidance usually fails to diagnose heavy rainfall areas generated by MCCs; reevaluate temperature and cloud forecasts for areas affected by MCCs. MCCs interact with and modify their large-scale environment by modifying wind fields and temperature patterns from the lower troposphere to the lower stratosphere enough to affect future meteorological events over much of the central and eastern CONUS. Here is a list of statements that describe the characteristics of MCCs:

- Some summertime MCCs produce high winds during much of their lifetime.
- Most systems grow to maximum size after midnight and persist into the morning hours.
- The transition to a large highly organized mesosystem usually does not occur until late evening.
- Lightning damage occurs with half the MCCs and is most frequent during the development and mature phases.
- The nocturnal increase in low-level wind speed and significant veering enhances both the warm air advection and influx of moist unstable air into the MCC.
- The large-scale synoptic pattern provides for a conditionally unstable thermodynamic structure over a large regihead and to the right of the advancing MCC.
- An average of 16½ hours elapses between the first thunderstorm development and the time the MCC begins to decay. MCCs should exist for at least 6 hours to be classified (time wise) as an MCC.

Large, long-lived canopies of cold clouds signal persistent mesoscale upward forcing. The longer the very cold cirrus shields persist, and the more circular they become, the greater the relative strength and influence of upward vertical motion fields.

Forecasting mesoscale convective complexes

Under favorable meteorological conditions, convective systems develop and reach maximum MCC size around midnight. This is about the time that the low-level jet reaches maximum strength. The following parameters are good indicators of either potential MCC development or maintainability once MCCs are established:

1. Standard indices indicate the following:

   SBLI < –8
   TT > 54
   LI < –4
   K > 30

2. Upstream short waves are usually quite weak or not apparent.
3. The thermal analysis from the surface to 850mb always shows a strong warm air advection pattern into the southern and western flank of the MCC.

4. In the low-levels, high equivalent-potential temperatures (θ^e) are located to the front and right front of the system (boundary layer to approximately 850mb).
5. Surface dew points exceeding 65°F, mixing ratios exceeding 16g/kg, and 850mb dew points exceeding 12°C. Typical surface temperatures during the organizing phase are usually greater than 90°F.
6. A pronounced low-level southerly jet should be present in the genesis area. It then shifts eastward and veers during the night so that, by morning, it impinges directly on the southwest flank of the mature MCC (veering implies increasing warm air advection).
7. At 700mb, relative flow enters the MCC from the front and is characterized by low θ^e, which helps drive the strong cold downdrafts and the resulting mesoscale outflows. In other words, MCCs propagate “faster” than the 700mb winds. However, conventional RAOB data at 700mb usually shows neutral to warm air advection.

Upper-level patterns (500mb to 200mb)

Identifying the upper-level difluence zone over the MCC genesis region is extremely important. It generates an area of mass outflow and divergence into the top of the MCC. That, in turn, helps sustain the low-level convergence and the resultant mass inflow from the boundary layer through 700mb.

Typically, a broad ridge associated with a subtropical high is centered over the southern CONUS and extends northward into the central Plains. As the developing MCC moves east of the subtropical ridge axis, the rate and degree of upper-level divergence and difluence increases, consequently enhancing MCC maturation.

Organized thunderstorm generation develops just ahead of, and moves eastward with, a very weak short-wave trough that is translating within a fairly weak west-to-northwest flow more closely associated with the polar jet. The thermal and vorticity advection patterns are not well defined and become strongly affected by the evolution of the MCC itself. The identification of jet streaks is critical for diagnosing regions of enhanced upward vertical velocities. To help indentifying jet streaks look for:

- Entrance areas associated with the polar speed maxima (right rear).
- Exit areas associated with the subtropical speed maxima (left front effects).

An intense jet streak may develop along the northern periphery of the cirrus shield. Also, the upper-level flow becomes strongly divergent, as the convective complex continues to gain mesoscale organization and grows in size. Look for the height and temperature gradients to increase.

Regional forecasting hints

Here we divide the US into three MCC event regions. Do not just read about the area in which you may be stationed. Remember, we brief pilots who may fly into various MCC regions.

Northern Plains

Northern plains MCC events occur in Iowa, Nebraska, South Dakota, and southern Minnesota. They are often characterized by the following upper-air patterns from 500mb to 300mb.

A broad west-to-west-southwest flow is present over the central and northern Plains. The strongest winds are usually at the 500mb level across northern South Dakota into central Minnesota with speeds of 40 to 50 knots. A closed low-or long-wave trough that remains nearly stationary for two to three days should be present over southern Alberta or southern Saskatchewan. Weak, short-wave troughs and/or associated thermal troughs rotate east-northeast out of the long-wave trough position into an area of strong thermodynamics situated over the central and northern plains.

As the thunderstorms develop and move eastward into a region of increasing difluence, you’ll notice a marked increase in thunderstorm development taking place. Sometimes, there may not be any evidence of PVA or cold air advection aloft at these levels. However, a difluence zone should coincide with an area of low-level convergence within an area of upward vertical motion generated by warm air advection extending from the surface through 850mb.

Gulf Coast mesoscale convective complex

Gulf Coast MCCs occur predominantly during the late winter into the early spring months and can produce both widespread and extended severe weather along with very heavy rainfall. MCCs affecting this region tend to occur well to the east of a major trough, but with rapidly moving, weak, short waves and strong, persistent, low-level warm air advection.

The synoptic surface pattern usually depicts an east-to-west stationary or warm frontal zone that is the focusing mechanism for development. Once they are formed, these complexes tend to move slowly to the right of strong mid-level flow as new cells repeatedly develop along its southern flank. In these MCCs, individual thunderstorm cells move very rapidly with the mean flow, yet the MCC often remains stationary with extreme flooding developing.

Southern Plains

The surface features depict a slow moving or quasi-stationary synoptic scale cold front that is oriented in a south-southwest-to-north-northeast direction. Some MCCs form from squall lines developing on this front and gradually acquire MCC characteristics as they persist and grow in size. This often happens on the southern, or trailing, portion of the squall lines where the cold front trigger weakens and becomes slow moving under parallel flow aloft.

Often associated with a closed low, a major 500mb trough is located well to the west of the MCC genesis region. In ideal situations, a strong low-level moist southerly flow associated with the low-level jet begins to back into an approaching short-wave trough. As a result, convergence into the cold front increases, as does the low-level warm air advection pattern. In the upper-levels, difluence occurs at 200mb with a polar and/or subtropical jet located from 200 to 400 nm west of the genesis area.

252. Haboob dust storms

Water covers about 70 percent of the Earth’s surface. However, there are some regions of the globe that are less influenced by the abundance of moisture on Earth than others. Because of the lack of moisture in these regions, weather forecasters face unique challenges. For example, forecasters must be aware of the atmospheric differences between moist air and dry air such as the difference in a moist versus a dry adiabatic lapse rate. Dust storms can be a common occurrence in desert regions. The two most common dust storms are the haboob and the shamal. In this lesson we’ll discuss the haboob.

One of the most dynamic weather occurrences in the desert is the haboob, an intense sandstorm or dust storm found in arid regions such as the desert southwest area of the United States and the Sahara Desert in Asia. Based on climatology these storms are most frequent May through September, with June being the peak month. Haboobs are typically associated with downdrafts or downbursts from thunderstorms. Essentially a haboob is a mesoscale gust front or microburst occurring over a favorable dry source region. When the downburst interacts with the dry surface below, dry, loose sand is blown up, creating a wall of debris. Meteorological satellite imagery has shown this “wall” can extend up to 18,000 feet in height. However, the average height of these dust storms is usually between three and six thousand feet. Moreover, the storm can extend to a horizontal width of up to 90 miles.

Weather forecasters should be familiar with the atmospheric conditions favoring haboobs. Because forecasting the occurrence of an outflow boundary is difficult, haboobs can be very challenging to forecast. Initially the source region for the haboob’s formation must be considered.

The two primary concerns for forecasting haboobs are visibility and wind speed. Visibilities in the wall of the storm vary based on proximity to its source region and the amount of blowing debris. When the storm is in or near its source region the visibility can decrease from one half to zero nautical miles. The further the storm is from its source region the higher the visibility. Although visibilities can decrease to zero, drastic improvement can occur within an hour. When the dust and sand settles visibilities are observed between two to five nautical miles. In addition to visibility concerns with haboobs, wind speed forecasts are of great importance. Peak winds can be up to 95 percent greater than the speed of the haboob's movement. For example, if the haboob is moving at 25 knots, multiply 25 by .95, this equals 24 knots. Add the speed of the haboob (25 knots) to 95 percent of the haboob's speed (24 knots). Using this method the haboob would produce a max gust potential of 49 knots. Remember, these are only general guidelines an in-depth analysis of the atmosphere is required to make an accurate forecast.

Forecasting a haboob will require an understanding of desert thunderstorms. The conditions necessary for thunderstorm development as described earlier in this unit remain important. However, in the desert the dynamics are slightly different. While moisture is necessary for thunderstorm development, instability brought about by intense surface heating or an upper level trough can trigger convective activity. Because of the lack of moisture in the desert, thunderstorms have high bases. Additionally, precipitation associated with the thunderstorm often does not reach the ground. This phenomenon is known as virga, precipitation evaporating before reaching the Earth's surface. We learned earlier in this volume how heat is absorbed from the surrounding atmosphere through evaporation, resulting in a cooler atmosphere. This cool dense air rapidly sinks, contributing to the formation of a haboob.

Haboob detection using satellite imagery

Because haboobs occur in data sparse desert regions, the use of meteorological satellite (METSAT) imagery can be a very helpful tool for identifying dust storms. When dust is viewed using METSAT imagery the upstream edge is usually not well defined. Dust has a filmy, diffuse appearance with a medium to light grey shade on visible and near infrared imagery. Dust does not usually show up on distant infrared imagery due to the lack of thermal difference between the ground and the dust. However, if it does show up it is a dark to a medium grey shade. Use of an enhancement curve specifically designed to identify dust can be very helpful.

Haboob detection using radar

When dust particles are lifted by wind it makes it possible for weather radar to detect a haboob. Usually the detection is limited to the lower elevation slices of the radar. Due to the limited back scattering associated with dust versus precipitation, reflectivity values are usually low. In the desert southwest region of the United States, the WSR–88D radar can best detect dust using 'clear air' mode. Because of the increased sensitivity of 'clear air' mode, dust is easier to detect. Unfortunately due to the nature of the convective environment, the WSR–88D in most cases will be in the less sensitive "precipitation' mode, making dust difficult to identify. Tactical radars used in Southwest Asia usually are less sensitive and do not identify sand as well as the WSR–88D.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

249. Northwest flow outbreaks

1. What characterizes an NWF event?

2. What are the two most frequent axes for NWF events?

3. What is the average 500mb wind flow angle for an NWF event?

4. What is the average Showalter index, the total totals index, and the surface-based lifted index associated with NWF activity?

250. The derecho

1. What is a derecho?

2. What are the two most distinctive, common physical characteristics of a derecho?

3. What is an absolute requirement for the maintenance or life span of a derecho?

4. What upper-level feature is involved in all derecho events?

251. Mesoscale convective complex

1. What characterizes an MCC?

2. What are the stages in the lifecycle of a MCC?

3. What are the standard indices for an MCC event?

4. Why is it extremely important to identify the upper-level difluence zone over an MCC genesis region?

252. Haboob dust storms

1. Define a Haboob?

2. What are the two primary forecast concerns when forecasting a haboob?

3. What is the max wind speed you would forecast for a haboob moving at 30 knots?

4. How does dust appear on visible satellite imagery?

3–4. Severe Weather Associated with Tropical Weather Systems

As a weather journeyman at an operational weather squadron a tropical weather system will present a unique challenge. A hurricane and/or tropical storm has the potential to affect the operations at many DOD installations for a period of days to as long as a week. Past hurricanes that have made the US gulf coast, weakened to a tropical depression stage, moved northeast off the eastern seaboard, and restrengthened to become a hurricane again. These weather systems are difficult to forecast and it's important to realize that even as a tropical depression can produce heavy rainfall over a large geographical area. Let's learn more about these systems.

253. Classifying tropical cyclones

Tropical cyclones form over all the tropical oceans except the South Atlantic and the South Pacific east of about 130°W. Regional differences in terminology are listed in the table below.

Region	**Range of Maximum Wind Speeds (Knots)**	
	34–63	64–165
	Classification	
Western North Pacific	Tropical Storm	Typhoon
Bay of Bengal and Arabian Sea	Cyclone	Severe Cyclone
South Indian Ocean	Tropical Depression	Tropical Cyclone
South Pacific	Tropical Depression	Tropical Cyclone
North Atlantic and Eastern North Pacific	Tropical Storm	Hurricane
NOTE: In some regions, "Tropical Storm" is subdivided into Tropical Storm – 34 to 47 knots, and Severe Tropical Storm – 48 to 63 knots.		

Besides the classifications in the table above, the US meteorological services define a tropical depression as a weak tropical cyclone with a surface circulation incorporating one or more closed isobars, with the highest sustained winds (averaged over one-minute or longer periods) of less than 34 knots. Extreme surface wind gusts in tropical cyclones may be 30 to 50 percent higher than the reported sustained surface wind. US meteorologists also use the term "tropical disturbance," defining it as a separate system of apparently organized convection, generally 75 to 250nm (150 to 500 km) across, originating in the tropics or subtropics, having non-frontal migratory characteristics and having maintained its identity for a day or more. Tropical disturbances may subsequently intensify into tropical cyclones. Development of a tropical cyclone can take several days as shown in figure 3–39. A meteorological satellite is the tool used to monitor tropical cyclone development over oceans where conventional surface data is sparse.

The hurricane season starts June 1 and runs to the end of November. Early and late in the Atlantic season, cyclones tend to form in the southwestern Caribbean when the monsoon trough extends eastward from the eastern Pacific. These storms pose a severe weather threat to eastern and gulf coastal areas of the US.

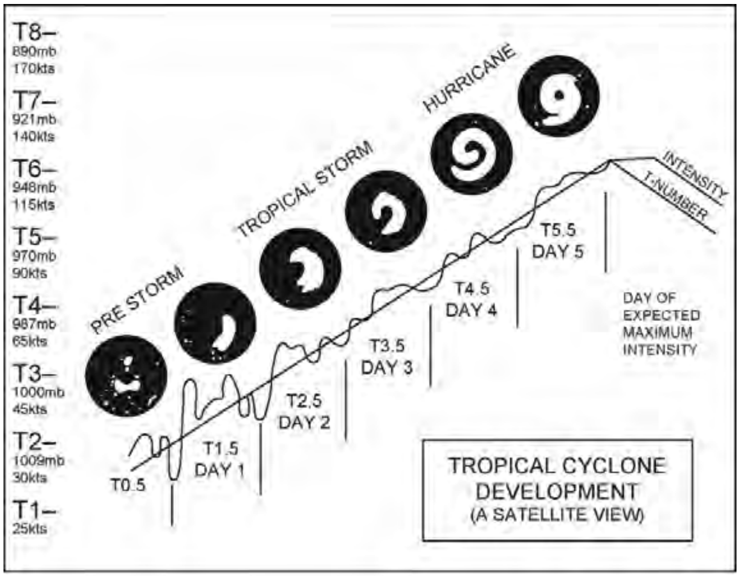


Figure 3–39. Tropical cyclone development.

254. Tropical cyclone weather elements

Understanding the weather elements associated with tropical cyclones enhances your ability to support specific operations. You'll not specifically forecast the tropical cyclones' movement, but you'll tailor either the Tropical Prediction Center's (TPC) or the Joint Typhoon Warning Center's (JTWC) warnings to your local area briefings.

Remember these two agencies are the only ones authorized to issue warnings on tropical cyclones. You can never issue a warning for movement and intensities on tropical cyclones.

Surface pressure

Since the central pressure of a mature storm is well below average, sea-level pressures furnish an excellent tool for analysis. Isobaric patterns may be almost symmetrical or elliptical. The strongest pressure gradient is to the right of the storm's path.

Barometric tendencies (isallobars) are not a particularly good indication of the storm's movement outside its sphere of influence. Usually, the pressure falls extremely quickly within 3 hours before the storm's arrival, and rises at an equal rate after passage. Central pressures of 950 to 960mb are not uncommon.

Temperatures

Tropical cyclones can change the sea-surface temperatures (SST) behind them because of the strong winds mixing the water layers. The temperature is usually 9°F (5°C) cooler behind the storm versus the temperature before the storm moved through. Due to the latent heat released into the storm aloft, the temperature is 5°C warmer in the tropical cyclone than the surrounding environment.

Clouds

Almost all the cloud forms are present in mature tropical cyclones. The most significant clouds associated with mature storms are heavy cumulus and cumulonimbus clouds spiraling inward toward the edge of the storm's eye. They become more massive and closely spaced as the clouds approach the eye. These spiral bands are known as rain bands.

Cirrus (CI) and cirrostratus clouds occupy the largest portion of the sky over the storm's area. Cirrus clouds become denser and changes to cirrostratus (CS) clouds while lowering, often giving the first clue to the arrival of a storm system. The appearance of the sky is very similar to an approaching warm front. The cloud pattern and aircraft data show there is very little vertical tilt between the low-level and upper-level centers.

Precipitation

Very heavy rainfall accompanies mature cyclones, but measurements are subject to error because of the high winds. Precipitation concentrates in the inner core of the spiral bands and totals of 20 inches are not uncommon. Over water, the ceilings are lower and the visibility is restricted by heavy precipitation. Over land, topographical features produce widespread flood conditions. Low ceilings, poor visibility, and high winds are additional threats to operations over land. High winds force moisture-laden tropical air up steep mountain ranges, thereby producing heavy rainfall. A total of 88 inches was recorded during one storm in the Philippines. At the other extreme, only a trace was reported at a station in Florida, which had 120-knot winds during the storm's passage.

Even after cyclones make landfall and weaken they are capable of producing large quantities of rainfall. Hurricane Agnes in 1972 made landfall on the Florida gulf coast then moved northeast as a tropical storm then as a depression, off the North Carolina coast over the Atlantic Ocean. It strengthened to tropical storm stage and paralleled the East Coast as it moved northward. As it moved over cooler waters it weakened becoming an extratropical low but not before dropping 6 to 15 inches of rain on New York, New Jersey, Pennsylvania, Maryland, Delaware, and Virginia. Flooding occurred as a result of the heavy precipitation and was responsible for 122 deaths. Agnes had maximum winds of only 75 miles per hour during her lifetime but caused death and hundreds of millions of dollars in property damage.

Eye of the storm

The eye of the storm is one of the oldest phenomena known to meteorologists. Precipitation ceases abruptly at the boundary of a well-developed eye. Inside the "eye," the sky partly clears and the wind subsides to less than 15 knots, and, sometimes, a calm area exists.

In mature storms, the diameter of the eye ranges from 3 to 100nm. The eye is not always circular; sometimes it becomes elongated and even diffuses, with the appearance of a double structure. Radar data has shown the eye of a storm constantly undergoes change.

Sea state

In coastal areas where rescue operations may be required, sea state is critical. One of the first signs of an approaching cyclone is the sea swell. It is a series of waves with crest intervals considerably longer than usual. The high winds with the storm create waves that move outward from the storm center more rapidly than the storm's progression and they outrun it. As the waves move outward from the storm center, the height from crest to trough diminishes. The length between crests is reduced and a low, rolling wave, known as a swell, develops. The size and speed of the swell depend on the velocity of the winds, the size of the water surface the winds flow over, and the slope and terrain of the ocean floor.

Storm surge

A storm surge is an abnormal rise of the sea along a shore and is the result of storm winds. It's measured as the difference between the actual sea-surface elevation and the elevation expected without a storm (predicted astronomical elevation).

Along low coasts, the storm surge is the chief destructive agent of tropical cyclones. With intense cyclones the sea may penetrate 10 to 20nm (20 to 40 km) inland. Since the surge generated at the coast must move inland as a gravity wave, the peak rise at the end of a long channel may lag the peak storm winds by many hours. All waves of consequence on the ocean surface are gravity waves. Interaction with the tide may shift the time of the worst flooding several hours away from the time of greatest storm intensity. Besides the winds and tides, the extent of coastal flooding also depends on the pressure difference between the storm's center and the surrounding environment, the size and speed of the cyclone, the bottom topography near the cyclone's land fall, and the topography around the land fall area. The most important factor in determining storm surge is the relationship of the maximum wind to the coast. The maximum wind is in the right-front quadrant of the cyclone along its direction of movement. Essentially, if the maximum wind is perpendicular to the coast, the storm surge is higher than if the maximum wind is parallel to the coast.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

253. Classifying tropical cyclones

1. How do US meteorological services define a tropical depression? Tropical disturbance.?

254. Tropical cyclone weather elements

1. What are the only two agencies authorized to issue warnings on tropical cyclones?

2. How would the surface pressure react with the approach and passage of a tropical cyclone?

3. What are the most significant clouds associated with a mature tropical cyclone?

4. What cloud type is often the first clue to the arrival of a storm system?

5. What is the first sign of an approaching cyclone?

6. What determines the size and speed of the sea swell?

7. Define storm surge.

8. Sometimes the worst flooding associated with a tropical storm may occur several hours away from the time of the storm’s greatest intensity. Why?

9. What is the most important factor in determining the storm surge?

Answers to Self-Test Questions

240

1. At the earth's surface, an air parcel has the same temperature as the air surrounding it. As we lift the air parcel up into the atmosphere the air pressure decreases with height. Consequently, the air pressure surrounding the parcel is relatively lower. The lower pressure outside allows the air molecules inside to push the parcel walls outward, expanding the parcel, lowering the average speed of the molecules resulting in a lower parcel temperature. Hence, a rising parcel of air expands and cools. During this process there is no interchange of heat with the environment.
2. If the parcel is lowered to the surface, air pressure becomes higher. The higher pressure squeezes (compresses) the parcel back into its original (smaller) volume. This squeezing increases the average speed of the air molecules and the parcel temperature rises. Hence, a sinking parcel is compressed and warms. During this process there is no interchange of heat with the environment.
3. The moist adiabatic lapse rate is 6°C per 1,000 meters. The dry adiabatic lapse rate is 10°C per 1,000 meters.
4. By comparing the temperature of a rising air parcel to that of its surrounding environment.
5. Absolutely stable.
6. Absolutely stable.
7. Increase stability.
8. Neutral.
9. Absolutely unstable.
10. Absolute instability.
11. Conditional instability.
12. Absolutely stable.
13. During the hottest part of the day.

241

1. Southeast United States
2. From the Continental Divide to the Appalachian Mountains and from the Gulf of Mexico to Canada. The greatest occurrence of severe activity is along "Tornado Alley."
3. March through June.
4. cP and mT, mT and cT.
5. With the position of the PFJ. From the Gulf Coast states in late winter, to the Southern Plains states in early spring, to the Central Plains in Midwest states in late spring, and finally, to the Northern Plains and Great Lakes region in the summer.
6. Non-severe: winds up to 49 knots and hail up to 3/4 inches. Severe: winds equal to or greater than 50 knots and hail equal to or greater than 3/4 inches.
7. Brief the crew to consider all thunderstorms in flight as severe. Also brief that hail, severe turbulence and icing, heavy precipitation, lightning, and wind shear are to be expected in and near thunderstorms.

242

1. The thermal air structure must be conditionally unstable. Large quantities of moisture must be available. Strong mid-level winds are required. There must be a lifting mechanism. The height of the WBZ must be favorable.
2. If the lapse rate of the surrounding air mass exceeds the rate of cooling by an ascending parcel, the parcel becomes warmer than its environment and rises due to positive buoyancy. Conversely, the parcel sinks if it is colder than the environment due to negative buoyancy.

243

1. Because weak winds aloft result in little cell movement.
2. The squall line thunderstorm.
3. Spring and early summer.

4. When the dryline is associated with an upper-level atmospheric disturbance along with intersections between the dryline and other boundaries.

244

1. Cumulus, mature, and dissipating.
2. Mature stage.
3. Hail forms in the shear zone between the updrafts and downdrafts near the freezing level in a convective cloud. The developing hailstone grows rapidly due to the accretion of super-cooled water droplets on ice crystals.
4. The meso-high results from the increased density of subsiding evaporatively cooled air striking the surface.

245

1. Because they can carry precipitation downstream to fall ahead of the updraft; keeping the updraft unobstructed.
2. New cells forming on the inflow flank (while older cells decay) causing the thunderstorm complex appear to move to the right of the mid-level winds.
3. They act as an exhaust mechanism, enhancing the updraft.
4. When the updraft shifts to the right (lower pressure) due to pressure imbalances.
5. The forward flank downdraft (FFD).
6. Downbursts and microbursts.
7. Cyclonic rotation; deviates approximately 30° to the right and slows down; potential tornado producer.

246

1. Between 1200 and 2100 hours local time.
2. 313 miles per hour or 272 knots.
3. In an area bounded by 600 miles east of the Rockies to 100 miles west of the Appalachians.
4. 20 to 40 miles.
5. At the windward border of the moist tongue near the 55°F isodrosotherm.

247

1. (1) b.
   (2) d.
   (3) c.
   (4) a.

248

1. Type I air mass.
2. The 55°F isodrosotherm.
3. Greater than 10°F degrees.
4. At intersection points between the dryline and another boundary.
5. The type B – frontal pattern.
6. Where a squall line intersects a warm front or an outflow boundary.
7. A major short-wave trough embedded in the westerly wind flow.
8. Due to the low freezing level.

249

1. The mean ridge on the 500mb product immediately preceding the event is upstream of the geographical location of the severe thunderstorm event. The mean trough is downstream. The average 500mb flow direction near the event is from a direction of 280° or greater. The mean ridge on the 500mb product immediately following the event remains upstream.
2. The first extends from eastern North Dakota east-southeast to southwestern Pennsylvania. The second axis extends from the Texas panhandle into Iowa.
3. 280°.

4. Less than –3°C, greater than 54, and less than –6°C, respectively.

250

1. A rapidly moving extratropical convective system known to produce widespread significant straight-line wind damage via a family of downburst clusters.
2. The radar configuration and the rapid rate of propagation in a direction slightly to the right of the mean wind vector.
3. High values of low-level moisture which results in very high levels of conditional instability.
4. A 500mb short-wave trough.

251

1. Cloud shield temperatures at least as cold as –32°C. Horizontal cloud tops extent to at least 100,000km^2 (about 70 percent of the size of the state of Iowa). An interior cloud top region of temperatures colder than –52°C and covering at least 50,000km^2.
2. The genesis stage, the development stage, the mature stage, and the dissipation stage.
3. A surface-based lifted index less than –8, a total totals index greater than 54, a lifted index less than –4, and a K-value greater than 30.
4. It generates an area of mass outflow and divergence into the top of the MCC that helps sustain the low-level convergence and the resultant mass inflow from the boundary layer through 700mb.

252

1. An intense sandstorm or dust storm resulting from a convective downdraft or downburst.
2. Visibility and wind speed.
3. 58.5 knots.
4. Diffuse with a medium to light gray shade.

253

1. A tropical depression is a weak tropical cyclone with a surface circulation with one or more closed isobars, with the highest sustained winds of less than 34 knots. A tropical disturbance is a separate system of apparently organized convection, generally 75 to 250nm (150 to 500 km) across, originating in the Tropics or subtropics, having a non-frontal migratory character and having maintained its identity for a day or more.

254

1. The Tropical Prediction Center and the Joint Typhoon Warning Center.
2. The pressure falls very rapidly 3 hours before the storm arrives and rises at an equal rate after passage.
3. Heavy cumulus and cumulonimbus clouds that spiral inward toward the edge of the storm’s eye.
4. Cirrus that is becoming denser.
5. The sea swell.
6. The velocity of the winds, the size of the water surface the winds are flowing over, and the slope and terrain of the ocean floor.
7. An abnormal rise of the sea along a shore and is the result of storm winds. It’s measured as the difference between the actual sea-surface elevation and the elevation expected without a storm.
8. An interaction between the tide and the storm surge.
9. The relationship of the maximum wind to the coast. Essentially, if the maximum wind is perpendicular to the coast, the storm surge will be higher than if the maximum wind is parallel to the coast.

Do the unit review exercises before going to the next unit.

Unit Review Exercises

Note to Student: Consider all choices carefully, select the *best* answer to each question, and *circle* the corresponding letter.

49. (240) The combination of cold air aloft and warm surface air can produce a
a. sharp lapse rate and a stable atmosphere.
b. gradual lapse rate and a stable atmosphere.
c. sharp lapse rate and an unstable atmosphere.
d. gradual lapse rate and an unstable atmosphere.

50. (241) In the continental United States, what months of the year are the *best* for severe thunderstorm activity?
a. April through September.
b. March through September.
c. April through July.
d. March through July.

51. (242) What are the conditions necessary for the development of tornadoes, severe thunderstorms, and their associated destructive phenomena?
a. Stable atmosphere, large quantities of moisture, weak mid-level winds, surface-based lifting mechanism, and a favorable wet-bulb-zero (WBZ).
b. Unstable atmosphere, small quantities of moisture, weak mid-level winds, surface-based lifting mechanism, and a favorable WBZ.
c. Unstable atmosphere, large quantities of moisture, strong mid-level winds, surface-based lifting mechanism, and a favorable WBZ.
d. Stable atmosphere, small quantities of moisture, strong mid-level winds, surface-based lifting mechanism, and a favorable WBZ.

52. (243) What kind of thunderstorm is the closest to a three-stage, non-severe thunderstorm?
a. Air-mass.
b. Squall-line.
c. Cold-frontal.
d. Warm-frontal.

53. (244) What stage of a non-severe thunderstorm occurs when precipitation is suspended aloft?
a. Mature.
b. Cumulus.
c. Super cell.
d. Dissipating.

54. (244) Hail size is dependent on the
a. size of the precipitation and the amount of moist-air advection.
b. size of the precipitation and the amount of dry-air entrainment.
c. strength of the updraft core and the length of fall from the freezing level to the surface.
d. strength of the downdraft core and the length of fall from the freezing level to the surface.

55. (245) Thunderstorms that appear to move to the right of mid-level winds are called
a. right movers.
b. splitting cells.
c. cell propagation.
d. discrete propagation.

56. (245) Which action is associated with the gust front?
a. Microburst.
b. Downburst.
c. Rear flank downdraft.
d. Forward flank downdraft.

57. (245) Super cells splitting into two separate cells is caused by the environmental wake flow "cutting" into the updraft on the
a. downwind side of the updraft core and precipitation "loading" in the mid levels above the updraft core.
b. upwind side of the updraft core and precipitation "loading" in the mid levels above the updraft core.
c. downwind side of the updraft core and precipitation "loading" in the low levels below the updraft core.
d. upwind side of the updraft core and precipitation "loading" in the low levels below the updraft core.

58. (246) What is the one requirement that must *always* be present in order for tornadoes to form?
a. Abundant moisture.
b. Appreciable lifting.
c. A low-level dry tongue.
d. A diffuse 850-millibars temperature ridge.

59. (247) In what tornado-producing air structure do tornadoes *most* frequently occur in families, with paths that are commonly long and wide?
a. Type I, Great Plains.
b. Type II, Gulf Coast.
c. Type III, Pacific Coast.
d. Type IV, Inverted "V".

60. (248) What is the *most common* severe weather phenomenon associated with the Type E, major cyclone severe weather producing synoptic pattern?
a. Hail.
b. Tornadoes.
c. Microbursts.
d. Downbursts.

61. (249) Where does one of the two well-defined northwest flow frequency axes extend?
a. Central Missouri to New England.
b. Along the entire Gulf coastal states area.
c. Eastern Texas to central portions of Illinois.
d. Eastern North Dakota to southwestern Pennsylvania.

62. (250) What type of destructive summertime mesoscale system is associated with northwest flow outbreaks?
a. Derecho.
b. Mesocyclone.
c. Vortex signature.
d. Mesoscale convective system.

63. (251) On the average, how many hours elapse between the first thunderstorm development of a mesoscale convective complex and the time it begins to decay?
a. 6.
b. 8 1/2.
c. 10.
d. 16 1/2.

64. (251) What upper-level (500 millibars [mb] to 200mb) zone pattern is extremely important to identify over a mesoscale convection complex genesis region?
a. Difluence.
b. Confluence.
c. Cirrus shield.
d. Anticyclonic shear.

65. (252) What is the *peak* month for haboob dust storms to occur based on climatology?
a. May.
b. June.
c. July.
d. August.

66. (252) What are the two primary concerns when forecasting a haboob dust storm?
a. Visibility and wind speed.
b. Visibility and wind direction.
c. Temperature and wind direction.
d. Temperature and wind speed.

67. (253) The typical hurricane season extends from
a. July through October.
b. June through November.
c. August through October.
d. January through March.

68. (254) A fairly accurate sign of an approaching tropical cyclone is
a. a strong temperature gradient ahead of the storm.
b. the barometric tendency up to 48 hours ahead of the storm.
c. an occurrence of steady precipitation up to 48 hours ahead of the storm.
d. a swell that comes in a series of waves with a long interval between crests.

69. (254) The difference between the actual sea-surface elevation and the predicted astronomical sea-surface elevation is known as
a. sea state.
b. storm surge.
c. storm swell.
d. cyclone tide.

Please read the unit menu for unit 4 and continue ➔

Student Notes

Unit 4. Non-convective Severe Weather

4–1. Precipitation 4–1
255. Heavy snow—upper level and surface parameters 4–1
256. Heavy snow producing synoptic patterns 4–2
257. Identify precipitation type using surface and upper-air temperatures and freezing level 4–5
4–2. Wind 4–10
258. Strong nonconvective winds 4–10
259. Nonconvective dust storms 4–14

NON-CONVECTIVE SEVERE WEATHER is just as much a threat to lives and property as is convective severe weather. For instance, an intense ice storm (freezing rain) can impact an Air Force (AF) base or a metropolis to the same degree as an intense thunderstorm. Ice-covered roads and runways can bring traffic to a halt or even worse, result in a mishap. Communication is disrupted as telephone lines fall due to the increased weight of the ice. Similarly, power lines can also come down resulting in people without electricity that is needed for heat. A case in point is the ice storm of January 1983, which left more than 250,000 people from Mississippi to New England without power and caused an estimated 25 deaths. Considering the life-threatening situations that can arise from nonconvective severe weather, it’s imperative that you as a weather journeyman take an active role in learning the principles for forecasting this related phenomena.

In this unit, we cover five key topics related to nonconvective severe weather:

- Heavy snow—upper level and surface parameters.
- Synoptic pattern that produce heavy snow.
- Identify precipitation type using surface and upper air temperatures and the freezing level.
- Severe nonconvective winds.
- Nonconvective dustsorms.

4–1. Precipitation

As you know, precipitation can occur in convective and nonconvective environments. Although convective storms are usually the first thought that comes to mind when thinking about severe weather nonconvective storms can be equally devastating. One of the greatest factors associated with nonconvective severe weather is the excessive precipitation. Excessive rainfall can lead to dam failure, mudslides, and flooding.

In this section we’ll discuss weather patterns associated with nonconvective severe weather. Additionally, we’ll discuss the techniques and general guidelines used to forecast nonconvective weather events. Whether its rain, snow, or freezing precipitation a good forecaster needs to be aware of the nonconvective precipitation forecast challenge that may affect the mission you are supporting.

255. Heavy snow—upper level and surface parameters

The 500mb absolute vorticity maximum, the 500mb height contours, the 1,000mb to 500mb thickness contours, the surface low-pressure center, and dew points at 850mb and 700mb have been studied to establish a model for forecasting heavy snowfalls. It was found that 6.5° to 7.0°of latitude downstream and 2.5° to the left of the track, with respect to the 500mb vorticity maximum, identifies the most favorable area for heavy snow.

The most favored area in relation to the surface low-pressure center was found about 5° along and 2.5° left of the path of the low. Based on the positions of the 500mb low, the most favored area for heavy snow was found along the path of the low and a little downstream from the point where the contour curvature changes from cyclonic to anticyclonic. You can also expect heavy snow near the 1,000mb to 500mb thickness ridge within the contour interval of 5,310 to 5,370 meters.

Be aware of advection since warm advection in the lower levels during the snow-producing situations can suggest increased snow potential. At the 850mb level, the –5°C dewpoint lines, and at 700mb level, the –10°C dewpoint lines are used as the basic defining lines (fig. 4–1). The area at 850mb that lies within the overlap of the 0°C isotherm and the –5°C dewpoint lines are the first approximation of the maximum. This area is then refined by superimposing the –10°C dewpoint lines at 700mb on the area. The final area of forecast heavy snow is where the three lines overlap.

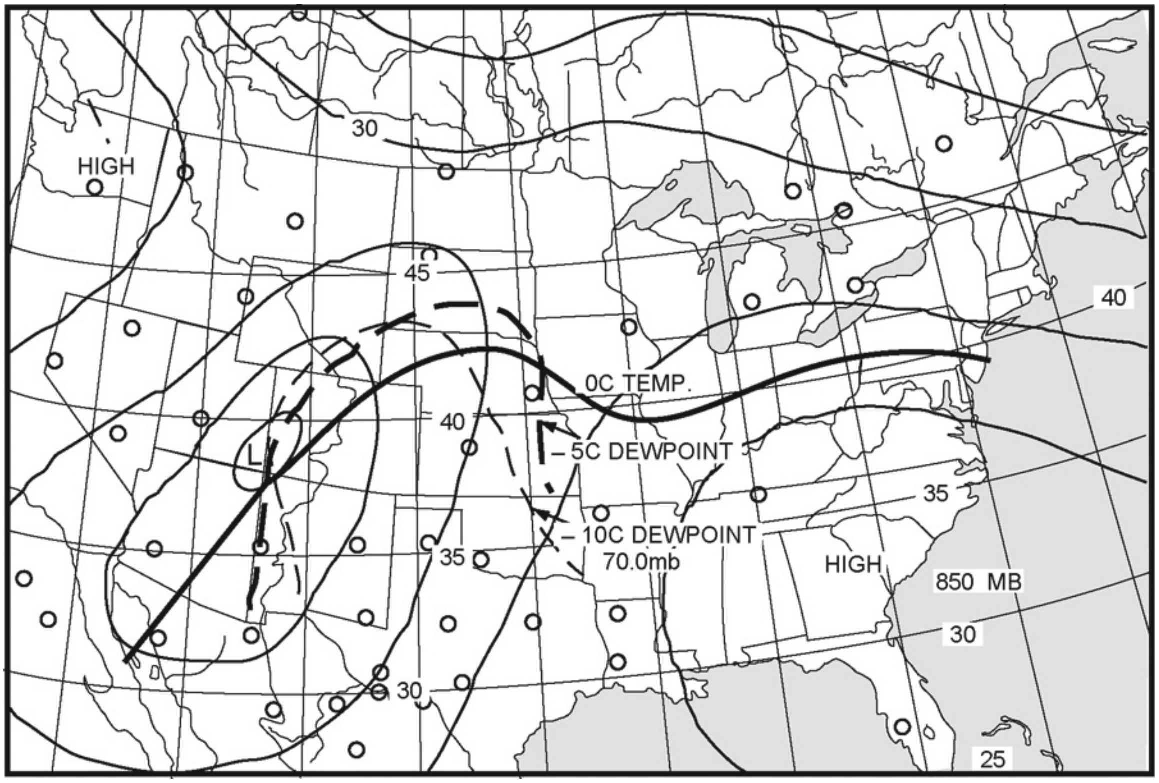


Figure 4–1. Location of maximum snow area.

Snow usually begins with the passage of the 700mb ridge and heavy snow usually ends with the passage of the 700mb trough or low center. All snowfall usually ends with the passage of the 500mb trough or low center.

256. Heavy snow producing synoptic patterns

There are a number of synoptic situations conducive to producing heavy snow. Let’s take a look at some of them.

Baroclinic non-occluding low

This system usually moves to the east-northeast at approximately 25 knots or greater. It is usually associated with a fast-moving trough at 500mb. The area of maximum snowfall:

- Has a snow band approximately 100–200 nautical miles wide.
- Lies to the left of the surface low oriented parallel to the track of the low.
- Has a snowfall rate of greater than or equal to 1 inch per hour and lasts from four to eight hours.

- Depicts the southern edge of the maximum snowfall to be located about 60 nautical miles (1° latitude) to the left of the low track.

Baroclinic deep-occluding low (blizzard type)

This is the most dangerous type of winter storm that includes winds greater than or equal to 40 knots, temperatures less than or equal to 20°F, and drastically reduced visibility caused by blowing and/or falling snow. A severe blizzard contains winds greater than or equal to 50 knots, and temperatures less than 10°F.

This system features a deep occluding low with a downstream, wraparound ridge that is oriented northwest to southeast (a type B occlusion). It is associated with a closed, cold-core low at 500mb. The system usually moves to the north-northeast, slowing to approximately 5 to 10 knots.

The area of maximum snowfall:

- Lies to the left and is parallel to the movement of the surface low.
- Depicts the western edge of the heavy snow area to be the 700mb low and/or trough axis, while most of the snow usually ends with the passage of the 500mb trough.

Snowfall rates of one to two inches per hour lasting up to 10 hours are quite normal. However, the length of duration is highly variable based on the speed of the system, and other factors. If short waves continue to develop and propagate around the upper-level low, the system becomes highly susceptible to secondary development, especially along the east coasts of continents.

Post-cold frontal trough

This system is associated with a steep cold front oriented in a deep long-wave trough. This synoptic situation exists when strong cold-air advection exists from the surface to 850mb west of the front. Sharp amplitude short-wave troughs at 700 and 500mb are located about 200 to 300 nautical miles to the west of the surface front. The area of maximum snowfall is located between the 850 and 700mb short-wave troughs. This area usually generates and dissipates in 12 to 18 hours with the heavy snow lasting only two to four hours.

Warm advection synoptic pattern

This synoptic pattern does not have an active low-pressure center in the vicinity of the maximum snow area. A persistent high-pressure ridge is situated to the north to northeast of a quasi-stationary front. A well-developed, low-level flow (850mb) of moist air overruns the frontal surface. The area of maximum snowfall is located in a band parallel to and just north of the quasi-stationary surface front. The moderate to heavy snow usually lasts for six to 12 hours. The front transitions into a warm front and begins to move, expect a change from snow to freezing rain and then to rain due to the advection of warmer, low-level air.

Inverted trough

This heavy snow situation is associated with an inverted trough that extends northward from a surface low-pressure system. The easterly component of the low-level flow around the low produces overrunning of the inverted trough as well as an upslope condition associated with an upper-level disturbance. All snowfall usually ends with the passage of the 700mb trough. Heavy snow is produced when the 500mb flow is parallel to the surface inverted trough. The area of maximum snowfall is located within and on either side of the inverted surface trough. The amount of snowfall depends on the moisture supply, winds, and speed of movement of the troughs.

Typical heavy snow track patterns

Figures 4–2 through 4–4 show the three most common tracks of both the surface lows and accompanying 500mb support that cause heavy snow accumulations in the CONUS. Keep in mind that the tracks shown in these figures can deviate from their indicated location to some degree.

Heavy snow track pattern 1

In figure 4–2, the alignment and movement of the surface low, the 500mb low, and the 500 height fall center tracks are shown. The 500mb level and short wave trough low is moving towards the Midwest and bottoms out over the southern Rockies and western and northern Texas before turning northeastward. The main frontal low would likely be along an mP frontal system approaching from the west.

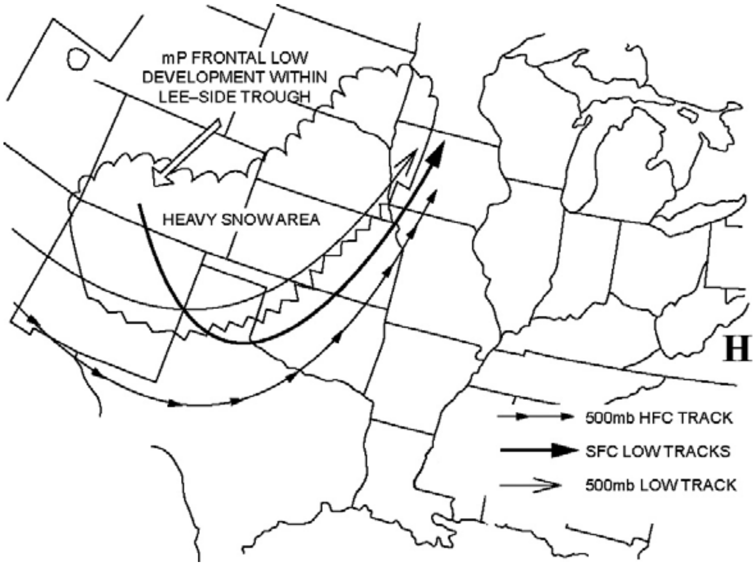


Figure 4–2. Heavy snow track pattern 1.

Heavy snow track pattern 2

In figure 4–3, the main surface low is likely to be a frontal low with maritime polar (mP) or continental polar (cP) air. The low is located some distance to the southwest of the upper low-pressure system due to the presence of a strong high-pressure area or ridge over the Midwest. There are many variations to the pattern shown in figure 4–3 depending upon the paths of the upper and surface low-pressure systems. This is an excellent overrunning situation, and considerable precipitation occurs southward to the surface low-pressure center. The division line between rain and snow lies along and to the northwest of the height fall center track.

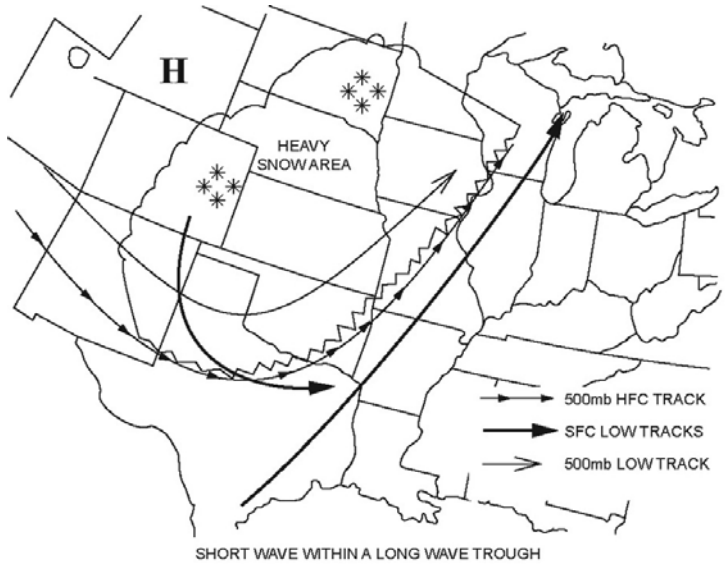


Figure 4–3. Heavy snow track pattern 2.

Heavy snow track pattern 3

The alignment of tracks shown in figure 4–4 is similar to figure 4–3. The pattern is presented because it occurs quite frequently over the central and western US and accounts for the majority of missed snow forecasts. In this pattern, a deepening trough over the western US exists as a long-wave feature. Short waves move through the long wave, bottom out over the Colorado Plateau, and swing northeastward across the Western Plains. At the surface, there is usually a stationary mP or modified cP front lying northeast-southwest across the Midwest. The main low development usually occurs along the front. The snowfall path is usually found along and to the northwest of the 500mb height fall track, within the colder air of the surface ridge rather than along the main surface low track.

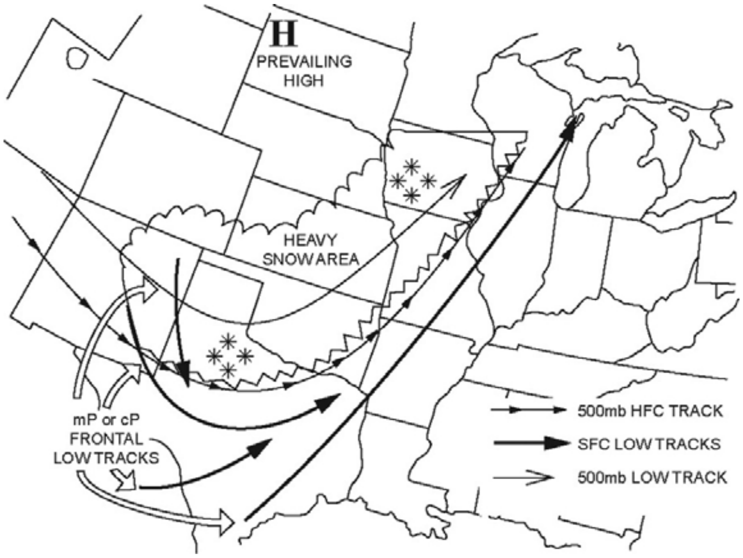


Figure 4–4. Heavy snow track pattern 3.

Each storm system track is different; therefore, carefully evaluate each situation and focus on where the 500mb height fall center bottoms out, where the main surface low develops, and the subsequent path of both features.

257. Identify precipitation type using surface and upper-air temperatures and freezing level

After you consider synoptic-scale situations, look at the surface temperature, the upper-air temperatures, and the height of the freezing level when you forecast precipitation types.

Surface temperature

The surface temperature by itself is not an effective criterion to predict the type of precipitation. Its use in the prediction of snow or rain has generally been in combination with other thermal parameters. One study for the northeastern US found that at 35°F, snow and rain occurred with equal frequency. By using 35°F as the critical value (predict snow at 35°F and below, rain above 35°F), 85 percent of the original cases could be classified. The table below shows additional study that gives guidance for predicting snow and rain in the US.

Surface temperature and precipitation type relationship	
When the temperature is	Forecast
Less than 22°F	Snow
23 to 33°F	Freezing rain (ZR), Ice pellets (IP), or mixed
34 to 39°F	Mixed rain and snow
Greater than 39°F	Rain

A third study suggests that snow never occurs when the surface dew point is equal to or greater than 3°C and that rain never occurs with a dew point equal to or less than –3°C. It suggests a marked change occurs at 0°C.

However, it is obvious from these studies that although surface temperature is of some general use in separating rain from snow, it is inadequate as a discriminator in crucial cases. Thus, most weather journeymen have looked to upper-level temperatures as a further aid to predict snow and rain.

Upper-level temperatures

The temperature-moisture distribution between the surface and the 700mb surface appears the determining factor in forecasting the type of precipitation. However, prediction of this distribution with any degree of accuracy is extremely difficult. Therefore, the medium level (850mb) is used with the precipitation area and the location of the 32°F isotherm on the surface synoptic product. When the areas defined by the 0°C isotherm at 850mb and the 32°F isotherm on the surface are superimposed on precipitation areas, they are found to separate precipitation types in most cases. Most of the rain-only cases are found on the warm side of the 32°F surface isotherm and most of the snow-only cases are found on the cold side of the 850mb 0°C isotherm. The intermediate types generally occur within the area enclosed by these two isotherms.

This is a rough delineation at best because of the time difference between the synoptic 850mb and the surface product. Furthermore, in many situations, evaporation and condensation affect the 850mb and the surface temperature significantly. Thus, to apply the above observations objectively, we should use the wet-bulb temperature because it is conservative with respect to evaporation and condensation, and because it is easily computed directly from the temperature and dew point.

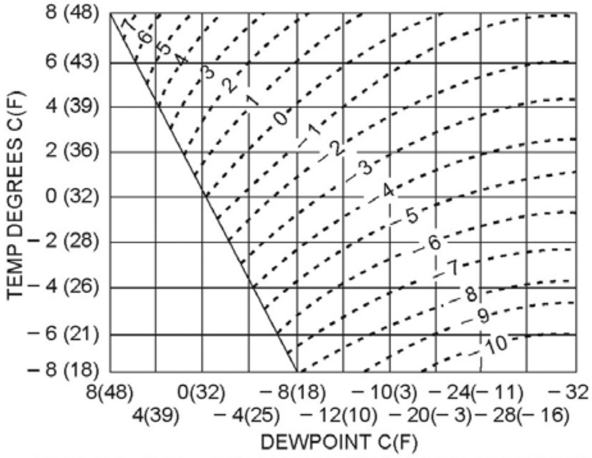
This technique uses three graphs (figure 4–5). Graphs I and II quickly compute the wet-bulb temperature on the 850mb and 1,000mb surface. Looking at the forecast air temperature and dew point, read the wet-bulb temperature in Celsius from the dotted lines. Since this technique confines itself to the area of the Rocky Mountain plateau, the surface is assumed at the 1,000mb surface.

The final graph is entered by plotting the 850mb wet-bulb temperature against the 1,000mb wet-bulb temperature. The type of precipitation is then read at the intersection. While the forecast use of graph III is obvious, the graph is not intended to be used as a sole predictor but, by its use, the forecast can be placed on a more objective basis. In other words, the use of this graph takes some of the subjectivity out of forecasting precipitation type.

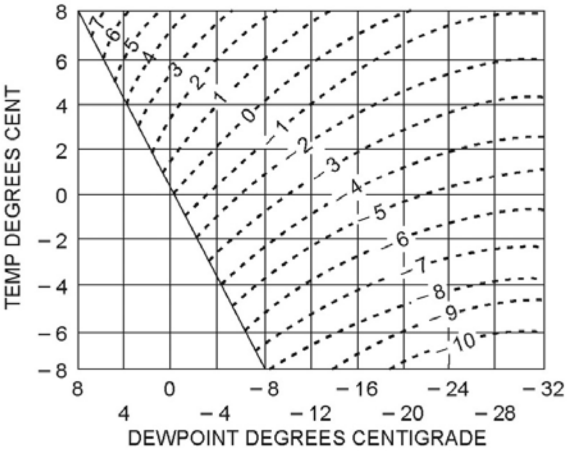
Carefully consider the factors affecting the wet-bulb temperature at a location before using the graph. For example, elevation, proximity to a warm body of water, known layers of warm air between the surface, the 850mb surface, and so forth are all factors affecting wet-bulb temperature. Area “A” on the graph calls for a forecast of rain; area “B,” freezing rain; and area “C,” snow. Area “D” is not clearly defined since it is an overlap portion of the graph; however, wet snow or mixed rain and snow predominates in this area.

The areas marked with an asterisk show no data were available in the original study. In that study, ice pellets (sleet) occurring for more than one or two hours was rare and should be forecast with caution.

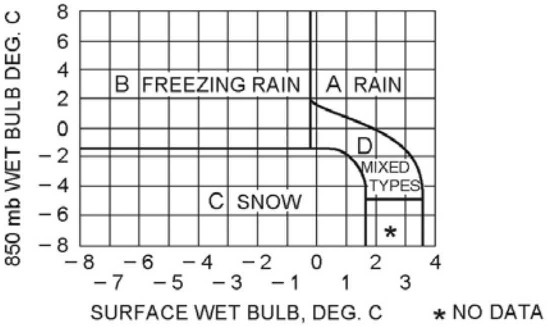
Admittedly, the wet-bulb temperature at the 1,000mb surface and 850mb surface does not give a complete picture of the actual temperature-moisture distribution since errors are introduced in estimating the 850mb wet-bulb temperature, particularly when values are changing. However, they are quantities that you can work with and predict with some degree of accuracy.



GRAPH I. 1000 mb/SURFACE WET BULB TEMPERATURE



GRAPH II. 850 mb WET BULB TEMPERATURE



GRAPH III. TYPE OF PRECIPITATION

Figure 4–5. Objective method of forecasting precipitation types

Another study uses the 850mb temperature as a guide for rain versus snow discrimination for stations close to sea level:

- Forecast snow if the 850mb temperature is less than –2°C.
- Forecast mixed precipitation if the 850mb temperature is between –2° and 0°C.
- Forecast rain if the 850mb temperature is warmer than 0°C.

Results compiled from various studies using upper air temperatures from geographical areas are shown in the table below. Surface temperature should be solely used as a predictor of precipitation type. Other thermodynamic parameters need to be considered as well.

Level Sno	w	Rain
Surface	Less than or equal to +0.7°C	Greater than 2.2°C
Dew Point	Less than or equal to –3°C	Greater than or equal to +3°C
850mb	Less than or equal to 0°C (–2°C East Coast US)	Greater than 0°C
700mb	Less than or equal to –6°C	Greater than –6°C
500mb: N of 40 N & mountains	Less than or equal to –30°C	
S of 40 N	Lees than or equal to –20°C	

Freezing level

The freezing level has been used to determine the type of precipitation. Generally, it is believed that the freezing level must be less than 1,200 feet above the surface to ensure that most of the precipitation reaching the ground is snow. One study presented conclusions on the height of the freezing level as follows:

- 12mb above the surface = 90 percent probability of snow.
- 25mb above the surface = 70 percent probability of snow.
- 35mb above the surface = 50 percent probability of snow.

You must understand that using freezing levels to determine the type of precipitation is somewhat subjective. Although there are many different methods used to forecast precipitation type, our intent is only to introduce you to some of them. Consequently, these methods use different freezing level thickness and height values to determine precipitation type. The method you use should be compared with the Local Analysis and Forecast Program (LAFP) at your base. In many instances, the LAFP states which method of forecasting precipitation type works best for that region.

In some cases the synoptic situation and corresponding Skew-T may depict either a single or multiple freezing level. The following methods are used to forecast expected surface precipitation based on single and multiple freezing levels. They consider the change of state of precipitation from liquid to solid or solid to liquid as it falls through the atmosphere.

Single freezing level

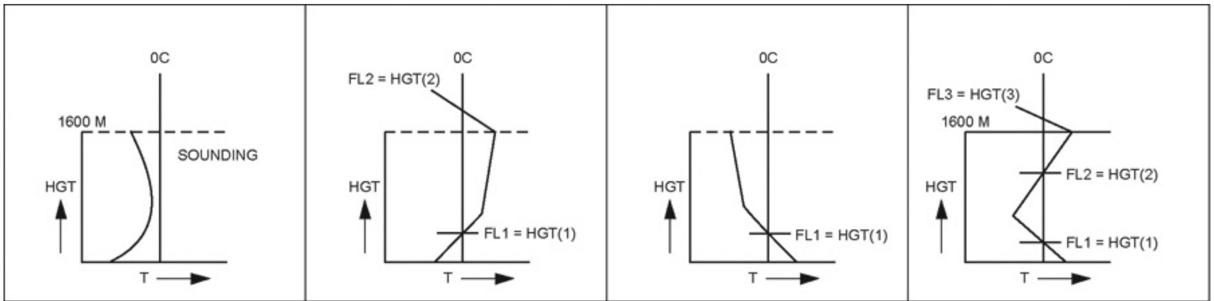
If the freezing level is greater than or equal to 1,200 feet above ground level (AGL) forecast liquid precipitation. If the freezing level is less than or equal to 600 feet AGL, forecast solid precipitation. And finally, if the freezing level is between 600–1,200 feet AGL, forecast mixed precipitation.

Multiple freezing levels

When there are multiple freezing levels, the Skew-T shows a warm layer where the temperature is above freezing. The following rules explain the effects of multiple freezing levels based on the thickness of the warm layer and the thickness of the cold air at the surface. If the warm layer is greater than 1,200 feet thick and the cold layer closest to the surface is less than or equal to 1,500 feet thick, forecast freezing rain. Conversely, if the warm layer is greater than 1,200 feet thick and the cold layer closest to the surface is greater than 1500 feet thick, forecast ice pellets. Finally, if the warm layer is between 600–1,200 feet thick forecast ice pellets regardless of the height of the lower freezing level.

National Weather Service study

A study completed by NWS developed sample soundings capable of being used to forecast the precipitation type. Figure 4–6 shows these soundings. Simply find the sounding that fits the sounding in your area and forecast precipitation according to the directives in figure 4–6. For example, if the sounding in your area shows a double freezing level, with the first freezing level less than 1,000 feet above the surface and less than 500 feet between the two freezing levels, you should forecast freezing rain.



FREEZING LEVEL SFC Below Freezing (BF)	FREEZING LEVEL 2 = FL2 = HGT(2) FREEZING LEVEL 1 = FL1 = HGT(1) SFC BF	FREEZING LEVEL 1 = FL1 = HGT(1)	FREEZING LEVEL 3 = FL3 = HGT(3) FREEZING LEVEL 2 = FL2 = HGT(2) FREEZING LEVEL 1 = FL1 = HGT(1)

PRECIP TYPE Snow (S)	If FL1 > 1000' and (FL2 - FL1) > 500' = Ice Pellets (IP)
	If FL1 <1000' and (FL2 - FL1) >500' = Freezing Rain (Z)
	If (FL2 -FL1) <500' = Snow (S)
		If FL1 <2000 FT = Mixed Rain/Snow
		If FL1 >2000FT = Rain (R)
			If (FL3 - FL2)<500' and FL1 <1200 FT and SFC Temp <36F = Snow (S)
			If (FL3 - FL2)>1000' and (FL2 - FL1)>500' and FL1 <1000' = Ice Pellets (IP)
			If (FL3 - FL2)>500' and <1000' and (FL2 - FL1) <500' and FL1 >1000' = Mixed Rain/Snow (M)

Figure 4–6. Using freezing levels to predict precipitation type.

Cloud-top temperatures

The thickness of the cloud layer aloft and the temperatures in the upper-levels of clouds are usually closely related to the type and intensity of precipitation observed at the surface, particularly in the mid-latitudes. Infrared meteorological satellite imagery with enhancement curves can be used to measure cloud-top temperatures. Climatology reveals the following:

- In 87 percent of the cases where drizzle was reported at the surface, the cloud-top temperatures were colder than –5°C.
- In 95 percent of the cases during continuous rain or snow, the cloud-top temperatures were colder than –12°C.
- In 81 percent of the cases, intermittent rain or snow fell from the clouds with cloud-top temperatures colder than –12°C; in 63 percent of cases, with cloud-top temperatures colder than –20°C.

The table below shows the relationship between cloud-top temperatures and the occurrence of showery precipitation in the United Kingdom.

Cloud-top Temperature	Shower Probability
0° to –12°C	Slight possibility
–13° to –40°C	Likely
Below –40°C	Almost certain

Weather Surveillance Radar 1998-Doppler (WSR–88D) indicators

You can also utilize WSR–88D products to determine where the freezing level is. We'll avoid going in-depth on WSR–88D products since they are discussed in a different CDC volume. The base reflectivity and composite reflectivity products can indicate where the melting level is. The melting level is that level where frozen precipitation particles melt during their descent to the surface. When frozen particles begin to melt, they become coated with water and provide a much stronger return than either the frozen particles above or liquid droplets below. This level may be detected on WSR–

88D reflectivity displays. It appears as a ring, or partial ring, of enhanced reflectivity around the radar data acquisition (RDA) unit. On the composite reflectivity product, you can also detect the melting level from the distinctive circles of higher reflectivities (fig. 4–7).

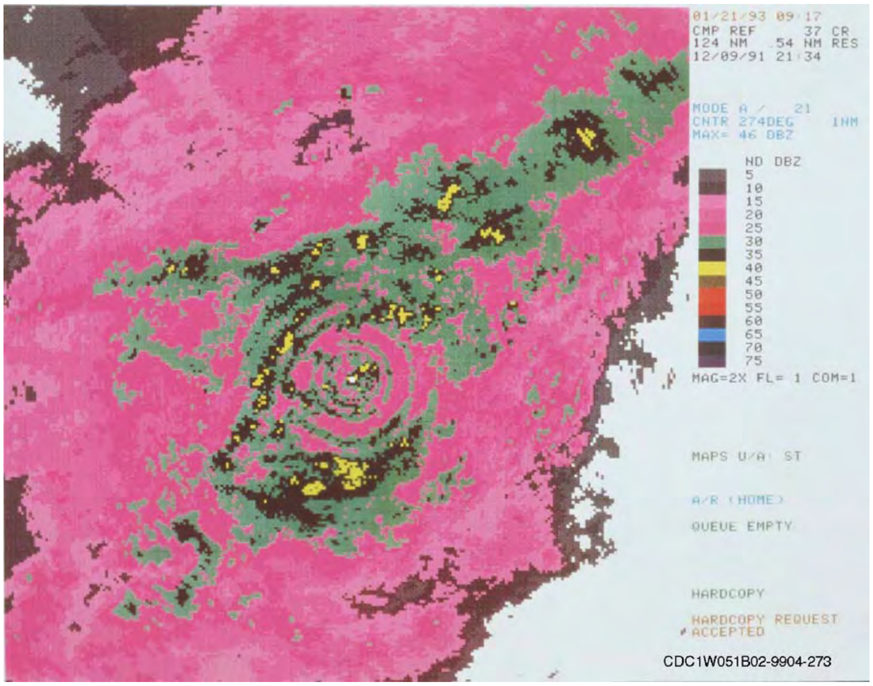


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Figure 4–7. Example of composite reflectivity (melting level).

As you can see in figure 4–7, each elevation slice that detects the melting level is depicted at the range or distance from the RDA where the melting level is encountered (the green concentric circles). This results in multiple rings of enhanced reflectivity around the RDA. You cannot determine the height of the melting level using composite reflectivity alone since it is a volumetric product. However, the composite reflectivity quickly shows the melting level in a given situation. You can then use an appropriate slice of a base reflectivity product to further investigate.

The base reflectivity product can easily determine the height of an enhanced ring. At any given elevation angle, the WSR–88D can compute the range and cursor height almost instantaneously. This is especially helpful when you are forecasting a change in precipitation type as a system approaches or moves across a region.

4–2. Wind

The previous unit we discussed the impact precipitation can have in a nonconvective environment. In this section will we shift our focus to another concern—nonconvective wind events. Strong nonconvective winds can present themselves in a variety of ways. This section will help you understand how to identify these events and equip you with the forecast techniques necessary to produce an accurate wind forecast.

258. Strong nonconvective winds

Strong nonconvective winds are winds that reach speeds in excess of 34 knots and are *not* associated with thunderstorm activity. Although not associated with convection; these winds have the same dangerous impact. They can blow down trees and utility poles, propel foreign objects like missiles, tear the roofs off buildings, and damage aircraft. If all these things sound bad, it’s because they are.

When wind speeds are forecasted to exceed 34 knots, weather forecasters are required to issue a weather warning. To better understand your role in this process, you must:

- Recognize the impact that strong nonconvective winds have on your customer.
- Learn the meteorological techniques or tools used at your location to forecast strong nonconvective winds.

Recognizing the impact of strong nonconvective winds on your customer begins after you arrive at your unit. Each operational weather squadron (OWS) has a unique mission, which also means a unique set of customer requirements. The Regional Area Forecast Program (RAFP) at the OWS documents the weather's impact on the various customers that the OWS supports. Part of your on-the-job training (OJT) is examining the RAFP to understand the impact of strong nonconvective winds on your customers. This information could be covered in the CDC, but with nine OWSs in AFW the information would be extremely lengthy and potentially redundant.

This lesson concentrates on illustrating some of the meteorological techniques used for strong nonconvective wind forecasting and is broken down into three topics:

- Strong nonconvective-wind forecasting rules of thumb.
- High wind warning decision flowcharts.
- Wind boxes.

NOTE: This lesson does not repeat the information covered in unit 1 or general tools for forecasting winds. Instead, the material in this lesson is biased towards wind speeds in excess of 34 knots.

However, it's also important to note that the wind forecasting techniques presented in the previous lessons can lead to a forecast in excess of 34 knots. For example, you could be using the techniques to forecast gusts equal to 80 percent of the strongest wind from the surface to 5,000 ft. If this technique concluded with a forecast greater than 34knots, use it, however, under the advisement of the RAFP.

Strong nonconvective wind forecasting rules of thumb

Over the years, weather units have completed forecast studies on high-wind events. These studies have revealed wind-forecasting techniques that the weather community calls a "rule of thumb" or ROT. To be considered a ROT is no small feat. Before a procedure or technique is declared a ROT, there must be overwhelming evidence proving that it can be verified consistently.

ROTs are compiled and documented in the RAFP for all weather forecasters to use. In some cases, these ROTs are even included on the forecast worksheet. Whatever the case, it's your job to use these ROTs on the job to make the best wind speed forecast for your customers. The table below depicts ROTs used for forecasting strong nonconvective winds around the Great Lakes. Use this table when cold fronts or storm centers are passing near the Great Lakes.

ROTs For Great Lakes Strong Winds Maximum Gust With Strong Cold Fronts
North-northwesterly, easterly, or southeast isobaric pattern
Compute pressure difference along a 400nm axis. Each millibar (mb) of pressure equals 3 knots of wind speed. A difference of 11 mb equates to 35 knots.
Northwest winds, north-south isobaric pattern
Compute the pressure difference along a 240nm axis. Each mb of pressure difference equals 3 knots of wind speed. Winds speeds are normally less than 45 knots, except for when an exceptionally strong storm passes the Great Lakes.

High-wind warning decision flowcharts

Another tool for forecasting nonconvective winds is the decision tree or flowchart. A flowchart is a pictorial representation of a process. Flowcharts use simple and easily recognized symbols to steer you through a process. The three most often used symbols in flowcharts for weather are the activity box, decision diamond, and direction arrows. Figure 4–8 illustrates these three symbols.

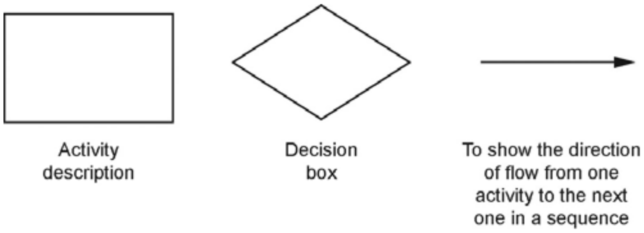

Figure 4–8. Flowchart symbols.

When experienced forecasters recognize that a particular procedure consistently provides favorable results, they'll standardize the process in a flowchart for the entire organization to use. Figures 4–9 and 4–10 are example flowcharts developed for nonconvective wind forecasts.

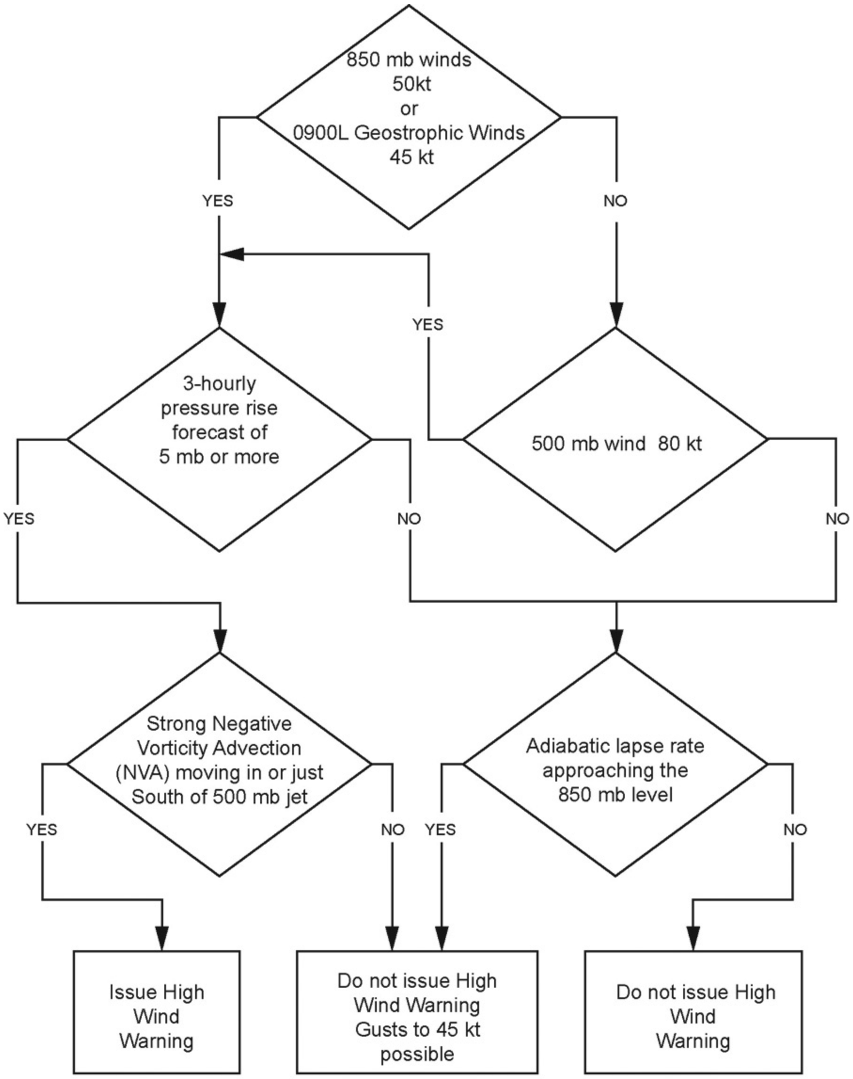

Figure 4–9. High-wind flowchart for the Northeast United States.

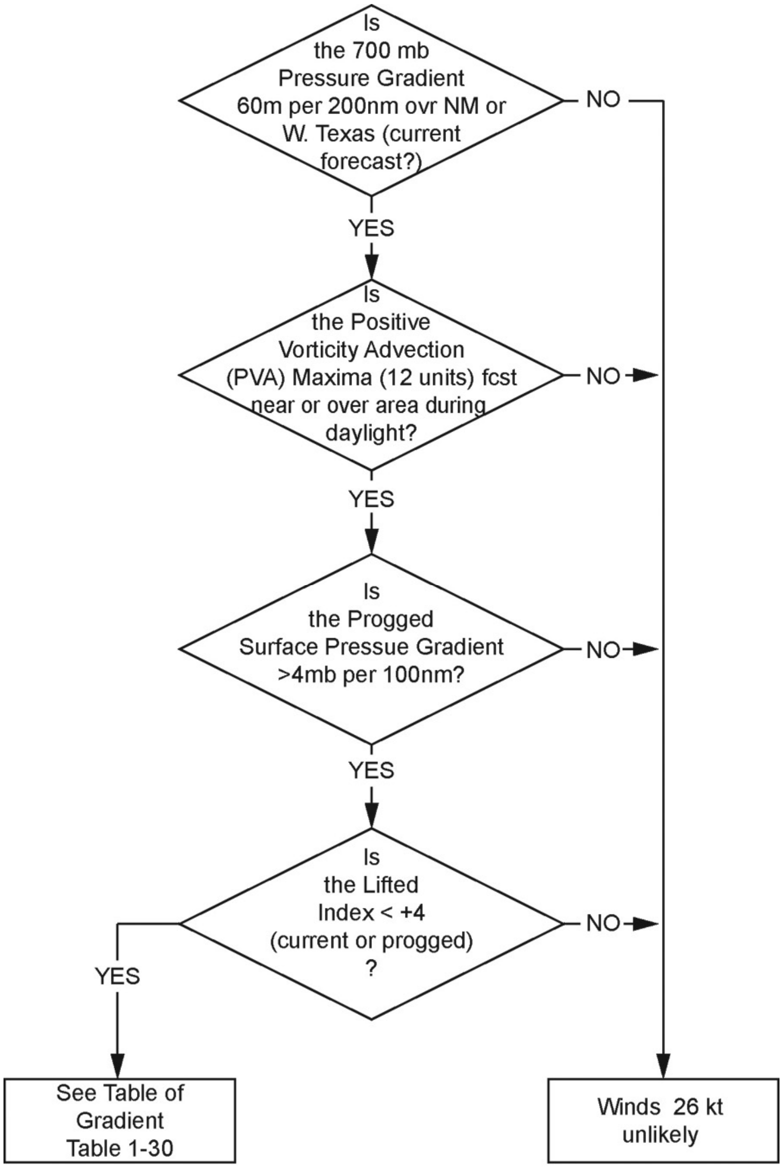

Figure 4–10. High-wind warning flowchart for North Texas.

Wind boxes

Wind boxes are notorious areas of strong wind gust patterns, over 35 knots. Of course, winds over 35 knots may occur in areas not included in these boxes. However, these areas have a history of experiencing strong nonconvective winds. Due to their location, topography, and migratory air mass patterns, boxed areas frequently encounter situations where all the ingredients necessary for strong nonconvective winds are present. Features such as tight thermal gradients, strong isobaric patterns, down-sloping winds, funneling, and so forth, commonly occur in these boxed areas. Figure 4–11 is a map depicting high-wind areas for the CONUS.

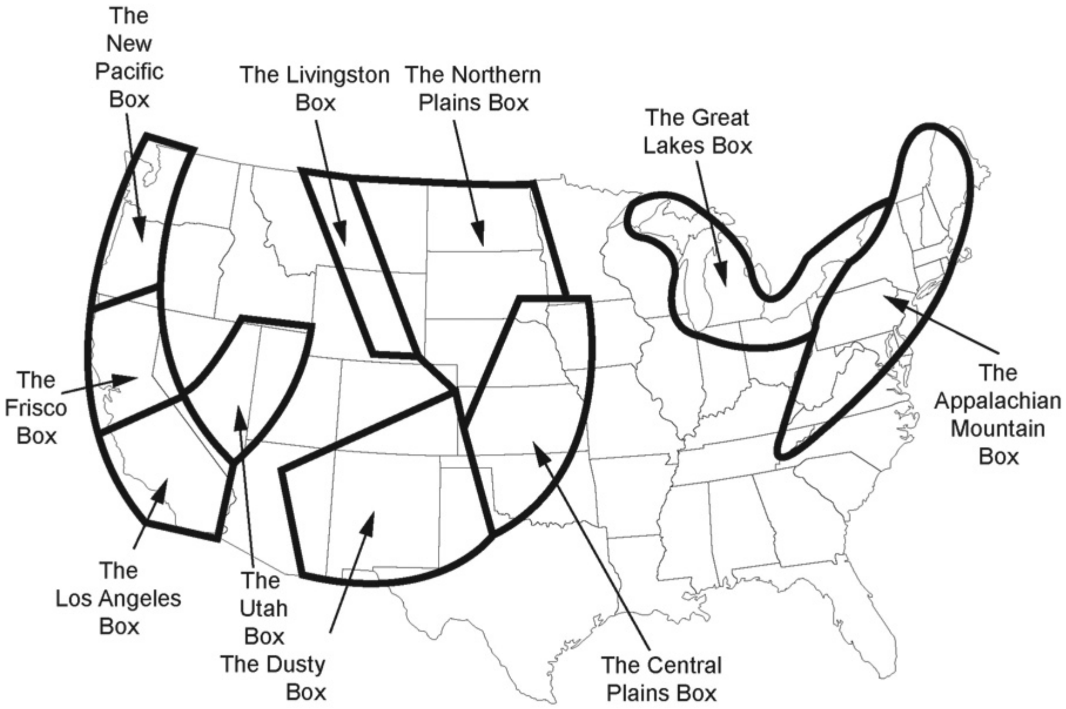

Figure 4–11. Location of CONUS "notorious wind boxes."

259. Nonconvective dust storms

In the previous unit we discussed one of the two common dust storms, the haboob. In this lesson we'll discuss the other dust storm, the shamal, as well as other nonconvective dust storms.

Shamals

The word shamal is an Arabic term meaning "north wind." Shamal's can be categorized by season—the summer shamal and the winter shamal. In the summer, shamals can be seen as a northwesterly wind blowing over Iraq, as well as other countries along the Persian Gulf. The summer shamal has also been called the "Wind of 120 days" as it occurs from early June through late September. The northwesterly flow only averages 10–15 knots. However, if the conditions are favorable the winds can be up to 40 knots for three to five days. In extreme cases these winds can last up to 10 days. The dust averages 3,000 to 8,000 feet in height but can reach up to 15,000 feet. Visibilities deteriorate to near zero as the storm moves over a location. Shamals are stronger during the day due to surface heating and increased instability. At night, radiational cooling stabilizes that atmosphere and settles the dust.

When forecasting shamals, the source region is of primary concern. There are many source regions for dust storms in Southwest Asia. The type of soil at the source region can vary based on particle size, which is an important factor when considering how far the debris will travel. Particles capable of traveling great distances usually have diameters of less than 20 microns. The three main soil types in the desert are clay, silt, and sand. Clay has a particle size of less than two microns. Silt (fine dust) is between two and 50 microns and sand usually measures greater than 50 microns. Figure 4–12 illustrates soil types in Southwest Asia.

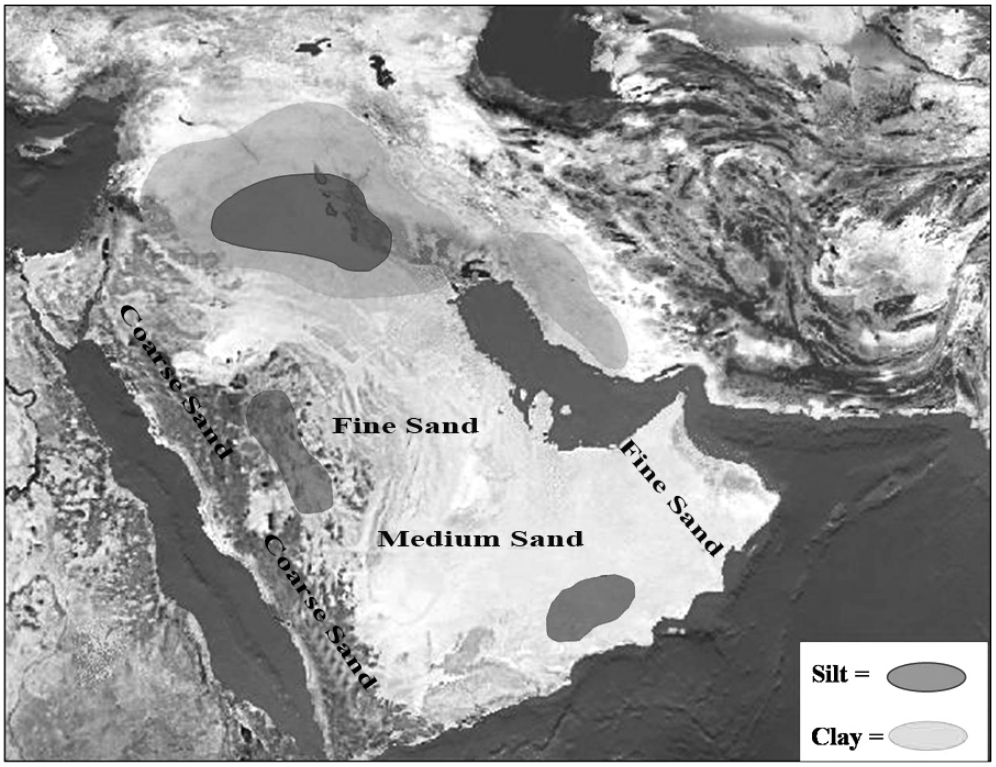

Figure 4–12. Soil types in Southwest Asia.

The synoptic conditions associated with a shamal has surface high pressure in place over the northern portion of the Saudi Arabian Peninsula with the monsoon trough draped over the southern portion of the peninsula. The trough creates a convergent zone between the sub-tropical ridge over the Mediterranean Sea and ridging in the southern Arabian Peninsula.

The winter shamal is associated with the funneling of very cold air masses from Turkey or Syria southward towards Iraq and into the Persian Gulf. The winter shamal is most often associated with the southward extension of the Siberian High. In this case, a tongue of dense cold air from the north lifts the warmer air to the south also lifting the dust and debris.

Pre-frontal dust storm

Pre-frontal dust storms frequently occur across Southwest Asia. Some of these storms are locally known as the Sharki in Iraq, the Kaus in Saudi Arabia, the Shlour in Syria, and the Khamsin in Egypt. Wind directions are usually southeasterly or southerly with wind speeds averaging 10–20 knots. Primarily dust is lifted ahead of a fast moving cold front due to a band of pre-frontal surface winds. Turbulent air caused by frontal lifting aids in the sand and dust displacement. Also, when the polar front jet core behind the surface front overlaps the subtropical jet core, upper vertical velocities are enhanced, lifting the dust and sand from the surface. Consider the techniques similar to what you've learned about forecasting surface winds with fronts.

Post-frontal dust storms

Post-frontal dust storms can last up to three to five days as a front stalls over a region. The dust storm is very active with strong cold air advection resulting in gusty surface winds. The dust storm usually extends from 8,000 to 15,000 feet high. Surface winds are typically 15–30 knots with visibilities commonly dropping down to zero. While forecasting for any single location, a post-frontal dust event typically lasts from 24 to 36 hours. If the front stalls the event could last up to five days with the potential for cyclogenesis to occur along the frontal boundary.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

255. Heavy snow—upper level and surface parameters

1. Where should you forecast heavy snow in relation to the 500mb vorticity center track?

2. Where would you forecast heavy snow in relation to the path of the surface low?

3. Where would you forecast heavy snow in relation to the 1,000-to–500mb thickness ridge?

4. What type of temperature advection in the lower levels increases the potential of heavy snow?

256. Heavy snow producing synoptic patterns.

1. Where does heavy snow fall that's associated with a non-occluded low? What is the snowfall rate?

2. What are the winds, temperatures, and visibilities associated with a baroclinic deep occluding low?

3. When does the heavy snowfall end that is associated with an inverted trough?

257. Identify precipitation type using using surface and upper-air temperatures and freezing level

Indicate whether rain or snow should be forecast:

1. Surface temperature 20°F.

2. Surface dew point of 5°C.

3. 850mb temperature of –4°C.

4. Freezing level at 2,000 feet.

5. Freezing level 10mb above the surface.

6. Entire sounding below freezing (fig. 4–6).

258. Strong nonconvective winds

1. What wind speed do strong nonconvective winds exceed?

2. What knowledge is beneficial to weather journeymen in better understanding their role as forecasters of nonconvective winds?

3. In terms of nonconvective wind forecasting, describe a ROT.

4. Explain how flowcharts assist forecasting nonconvective winds.

5. What are wind boxes?

259. Nonconvective dust storms

1. During which months does the summer shamal typically occur?

2. Explain the synoptic conditions associated with a shamal during the summer?

3. What are the three main soils in the desert?

4. How long can a post-frontal dust storm last?

Answers to Self-Test Questions

255

1. 6.5° to 7.0° downstream and 2.5° to the left.
2. 5° downstream and 2.5° to the left.
3. Within the contour interval of 5,310 and 5,370 meters.
4. Warm advection increases the potential.

256

1. It lies to the left of the surface low oriented parallel to the track of the low. The southern edge of the maximum snowfall is located about 60 nautical miles to the left of the low approximately 100–200 nautical miles wide.
2. Blizzard: winds ≥ 40 knots, temperatures ≤ 20°F; Severe blizzard: winds ≥ 50 knots, temperatures ≤ 10°F. Visibilities are drastically reduced in both instances.
3. *All* snowfall usually ends with the passage of the 700mb trough.

257

1. Snow.
2. Rain.
3. Snow.
4. Rain.
5. Snow.
6. Snow.

258

1. 34 knots.
2. (a) They must know how strong nonconvective winds impact their customers. (b) They must know the meteorological techniques used by their weather unit for forecasting nonconvective winds.
3. It is a meteorological technique that has been proven successful in forecasting nonconvective wind speeds.
4. A flowchart assists journeymen forecasters by taking them through a planned sequence of activity and decision making steps to arrive at a forecasting solution for nonconvective winds.
5. Geographical areas that often experience wind speeds in excess of 35 knots.

259

1. June through September.
2. Surface high pressure in place over the northern portion of the Saudi Arabian Peninsula with the monsoon trough draped over the southern portion of the peninsula.
3. Clay, silt, and sand.
4. Up to five days.

Do the unit review exercises before going to the next unit.

Unit Review Exercises

Note to Student: Consider all choices carefully, select the *best* answer to each question, and *circle* the corresponding letter.

70. (255) Warm advection in the lower levels during snow-producing situations
a. can suggest decreased snow potential.
b. can suggest increased snow potential.
c. shows where the snow/rain line occurs.
d. shows the most favorable area for blizzard conditions.

71. (256) What direction and speed does a non-occluding baroclinic low that produces heavy snow usually move?
a. East-northeast at approximately 5 to 10 knots (kt).
b. North-northwest as approximately 5 to 10 kts.
c. East-northeast at approximately 25 kts or greater.
d. North-northwest at approximately 25 kts or greater.

72. (256) A *severe* blizzard contains winds
a. greater than or equal to (≥) 40 knots (kt) and temperatures less than or equal to (≤) 20 degrees (°) Fahrenheit (F).
b. ≥40 kts and temperatures ≤ 10°F.
c. ≥50 kts and temperatures ≤ 20°F.
d. ≥50 kts and temperatures ≤ 10°F.

73. (257) When forecasting precipitation type, the layer between the surface and what millibar level is the determining factor in terms of temperature-moisture distribution?
a. 850.
b. 700.
c. 500.
d. 300.

74. (258) What is the threshold value, in knots, for strong nonconvective winds?
a. greater than (>) 29.
b. greater than or equal to (≥) 29.
c. > 34.
d. ≥ 34.

75. (259) What is the *primary* concern when forecasting shamals?
a. Time of day.
b. Time of year.
c. Source region.
d. Atmospheric stability.

Please read the unit menu for unit 5 and continue ➔

Student Notes

Unit 5. Evaluate Air Mass Soundings

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WHEN DECIDING how stable or unstable the atmosphere is over a location, you must do an analysis of the air mass. One means of describing the properties of the atmospheric air mass in a vertical column over an area is to use the Skew-T, Log P diagram. The accuracy of your forecast depends on your understanding of the indices derived from your Skew-T air mass sounding analysis.

5–1. Preliminary Analysis

The Skew-T diagram is one of your most useful tools. You can use it to analyze an air mass or front. It represents a realistic vertical cross section of the atmosphere over a given area and offers a virtual instantaneous snapshot of the atmosphere from the surface to about the 100 millibar (mb) level. It is also used to analyze atmospheric stability and the likelihood of thunderstorms, freezing rain, fog, icing, and turbulence. The table below lists some of the advantages for using the Skew-T diagram.

Advantages of Using Skew-T Diagram
You can better evaluate the character of severe weather should it develop.
Most importantly is that it’s the best tool to evaluate the stability of the atmosphere.
You can also see weather elements at every level of the column of air over a location.

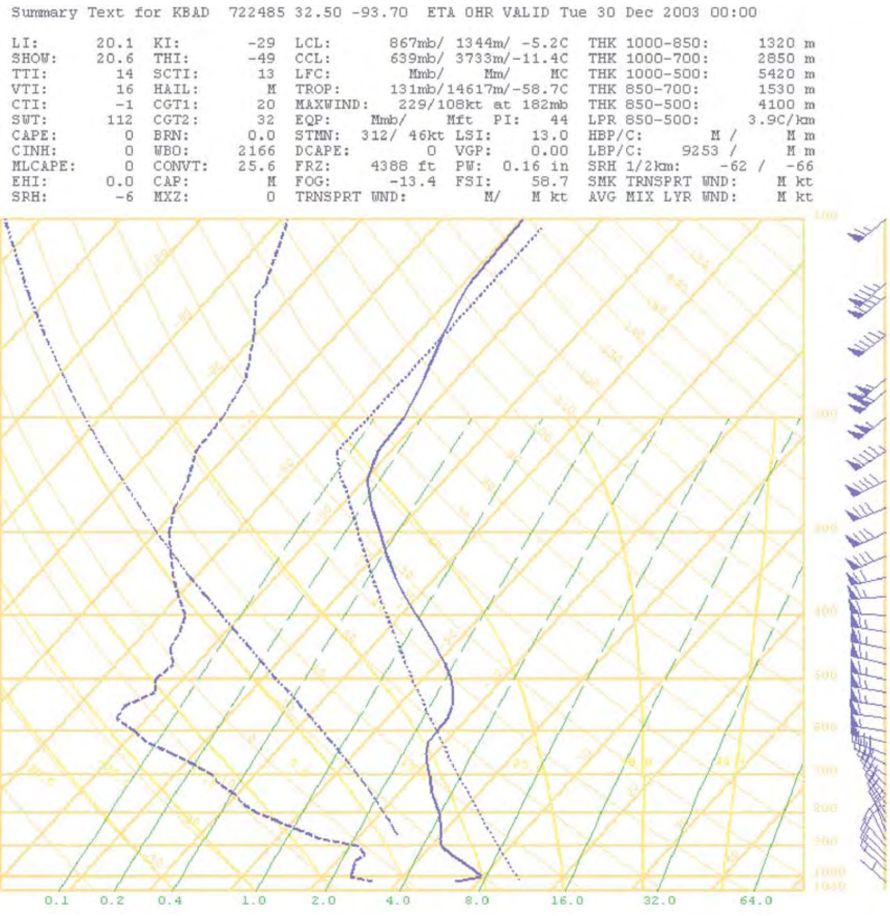
Unfortunately, there are some disadvantages to using the Skew-T diagram. The table below lists the disadvantages of using the Skew-T diagram.

Disadvantages of Using the Skew-T Diagram
It can take several hours for the data to become available.
It is only available twice a day and the elements can change dramatically in the interim.
The sounding does not always give a true vertical representation because the upper level winds blow the balloon downstream.

In this section, we dissect the chart and discuss each line separately. We recommend that you use a copy of a blank color Skew-T to assist in your understanding. You can download and print out one from various sources over the Internet.

260. Basic components, terms, and definitions

At first glance, the lines on a Skew-T product seem utterly complex and confusing (figs. 5–1 and 5–2). In actuality, they are only vertical representations of atmospheric parameters. These lines are spaced according to logarithms of pressure levels. The skewing of the lines presents a sufficiently large angle between the adiabats from which you can accurately deduce stability indices. It is constructed so that a ratio of area on the product to energy is the same over the whole diagram. Thus, a negative energy area at 850mb can be quantitatively compared to a positive energy area at 300mb. Because the atmosphere is constantly changing, accomplish a Skew-T analysis every 12 hours so you can assess the changes occurring in the atmosphere.



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SHOW:     20.6  THI:        -49  CCL:      639mb/ 3733m/-11.4C  THK 1000-700:      2850 m
TTI:        14  SCTI:        13  LFC:        Mmb/     Mm/   MC  THK 1000-500:      5420 m
VTI:        16  HAIL:         M  TROP:     131mb/14617m/-58.7C  THK 850-700:       1530 m
CTI:        -1  CGT1:        20  MAXWIND:   229/108kt at 182mb  THK 850-500:       4100 m
SWT:       112  CGT2:        32  EQP:     Mmb/    Mft  PI:   44  LPR 850-500:     3.9C/km
CAPE:        0  BRN:        0.0  STMN:  312/ 46kt LSI:    13.0  HBP/C:          M /     M m
CINH:        0  WBO:       2166  DCAPE:          0 VGP:    0.00  LBP/C:     9253 /     M m
MLCAPE:      0  CONVT:     25.6  FRZ:     4388 ft  PW:  0.16 in  SRH 1/2km:     -62 /  -66
EHI:       0.0  CAP:          M  FOG:      -13.4  FSI:    58.7  SMK TRNSPRT WND:     M kt
SRH:        -6  MXZ:          0  TRNSPRT WND:         M/   M kt  AVG MIX LYR WND:     M kt
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Figure 5–1. Skew-T Graph (26th OWS).

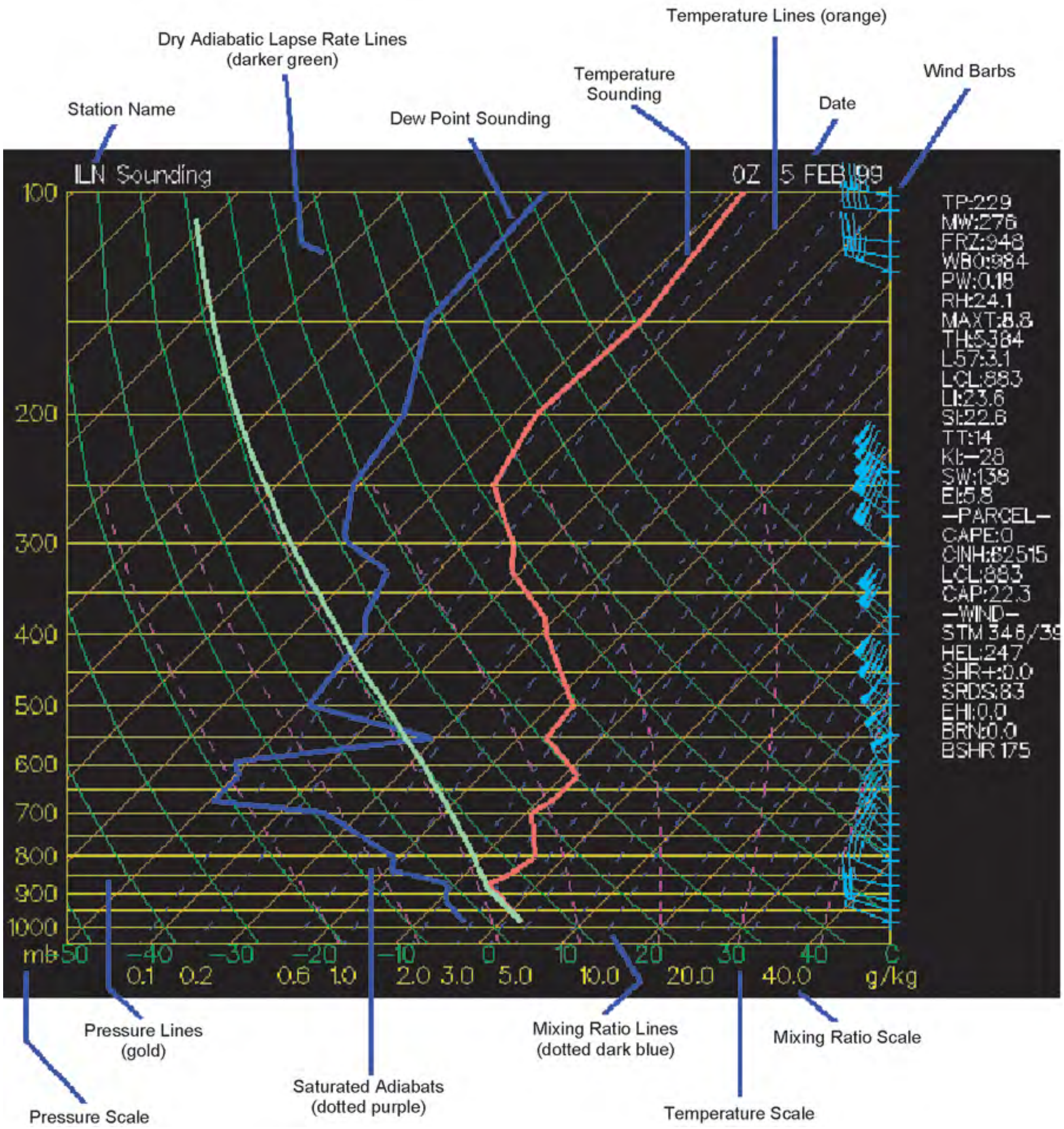

Figure 5–2. Skew-T Graph (Internet).

Standard heights

The following chart reviews the standard Skew-T pressure levels and the corresponding heights:

Pressure Level	Heights
850mb	1500 meters (m)
700mb	3100m
500mb	5500m
300mb	9300m

International Civil Aeronautical Organization standard atmosphere lapse rate

The International Civil Aeronautical Organization (ICAO) standard atmosphere lapse rate is shown by a thick line in the center of the diagram. It is used to compare the actual atmospheric conditions to the standard atmosphere.

Geopotential heights

Geopotential measurements in the atmosphere compensate for the effect gravity has on the column of air. The actual height is slightly greater than the geopotential height and increases with an increase in height (altitude). The small difference between geopotential height and actual height has no consequence in day-to-day weather forecasting and geopotential height is normally used as the actual height.

Isobars

Isobars are horizontal lines spaced at 10 millibar intervals from 1,000mb to 100mb and labeled in intervals of 100mb. The spacing between isobars increases from the bottom to the top of the chart. This accounts for the decrease in atmospheric density with height in the atmosphere.

Wind scale

Wind barbs are printed along the right side of the diagram for use in displaying wind data. For direction, north is the top of the diagram. A triangle denotes 50 knots (kt) of wind, a full barb is 10 kts and a half barb is 5 kts of wind. Cold air advection (CAA) and warm air advection (WAA) can be inferred by looking for backing (counterclockwise) and veering (clockwise) in the winds with height.

Isotherms

These are straight, equally spaced, lines that slope from the lower left to the upper right of the diagram. Each isotherm represents 1°C and is labeled in intervals of 10°C. A Celsius temperature scale appears at the bottom of the chart with values that coincide with the appropriate isotherm.

Dry adiabats

The dry adiabats are slightly curved lines that slope from the lower right to the upper left of the chart. They indicate the rate of temperature change in a parcel of dry air rising or descending adiabatically. They are labeled in 10°C intervals. Spacing between the dry adiabats decreases as their value increases. If the dry adiabats were continued beyond the chart, they would converge at a single point where the temperature reached absolute zero.

Saturation adiabats

Saturation adiabats, also called moist adiabats, are slightly curved lines. They slope from the lower right to upper left of the chart, but extend only to the 200mb isobar because humidity readings are not normally available above this level. Saturation adiabats represent the temperature change undergone by a saturated parcel of air as it rises or descends through the atmosphere. The moist adiabatic lapse rate defines the amount the temperature changes with height. Moist adiabats are printed and labeled every 2°C at the 530mb and 200mb levels. Their spacing increases as their values increase. Near 200mb, the moist adiabats are nearly parallel to the dry adiabats on the left side of the chart at temperatures and pressures where the amount of moisture in the air is small and temperatures are cold.

Saturation mixing-ratio lines

The saturation mixing-ratio lines are the slightly curved dashed lines sloping from lower left to upper right. These lines, just as the saturation adiabats, extend only to the 200mb level. The label for each line is in grams of water vapor per kilogram of dry air. These values are printed at the bottom of the chart from 0.1 to either 40.0 or 64.0 grams per kilogram (g/kg). The values represent the amount of water vapor necessary to saturate a volume of air at a given temperature and pressure. Notice the values increase toward the right side of the chart where the temperatures are higher because warm air can hold more water vapor than cold air.

Thickness scales

Thickness is defined as the distance between two pressure levels and is a function of temperature and moisture content.

Saturation mixing ratio

The saturation mixing ratio (w_s) is the moisture content air would have at a given temperature and pressure if the air were saturated. Saturation mixing ratio is expressed in grams of water vapor per kilogram (g/kg) of dry air. Saturation mixing ratio gives a maximum value to the amount of water vapor a parcel can hold and is one of the best measures of actual moisture present in the atmosphere.

The w_s at any level is found by using the temperature (T) of that level as one coordinate and pressure as the other coordinate. A mixing ratio line passes through the point where the isotherm and the isobar cross. The value attached to this mixing ratio line is the $w_{s.}$ We can summarize computing saturation mixing ratio by completing the following steps:

1. At the pressure level, find the temperature.
2. Read the value of the saturation mixing ratio line.
3. Label this value in g/kg.

For example, let's locate the point on a Skew-T where there is a temperature of –5°C and a pressure of 700mb. The 3.8 mixing ratio line passes through this intersection. So, for our example, the w_S is 3.8 grams of water vapor per 1,000 grams of dry air and is labeled in grams per kilogram (g/kg).

Mixing ratio

The mixing ratio (w) is the actual moisture content of the air.The actual mixing ratio can be equal to or less than the w_S but never more. Mixing ratio we can find in a similar way to the way we found w_S. The difference is in one coordinate. Use the dewpoint temperature (T_d) value as one coordinate and pressure as the other coordinate.

At the intersection of the dewpoint isotherm and the isobar, we can compute a mixing ratio. We can summarize computing mixing ratio by completing the following steps:

1. At the pressure level, find the T_d.
2. Read the value of the saturation mixing ratio line.
3. Label the value in g/kg.

For example, using the T_d of –13°C and pressure of 700mb, you'll find the moisture content or the mixing ratio is 2 grams of water vapor per 1,000 grams of dry air or 2g/kg.

Saturation vapor pressure

The saturation vapor pressure (e_s) is the partial pressure that the water vapor would contribute to the total atmospheric pressure, if the air were holding all the moisture possible for its temperature. Saturation vapor pressure is related to the w_s by the following empirical equation based on thermodynamics:

$$w_s = \frac{622e_s}{P}$$

You can see that if the pressure were 622mb, w_s would be approximately equal to e_s. This relationship allows us to find the e_s of a sample of air. You can do this by following the isotherm of the *free-air* temperature to the 622mb level and reading the mixing ratio shown at this point. The numerical value of the mixing ratio is equivalent to the e_s in millibars. We showed this by using a free-air temperature of 17.2°C and a pressure of 963mb. Following the 17.2°C isotherm to the

622mb pressure level, we read a mixing ratio of 20.5 g/kg (fig. 5–3). The e_s is 20.5mb where the temperature is 17.2°C and the atmospheric pressure is 963mb.

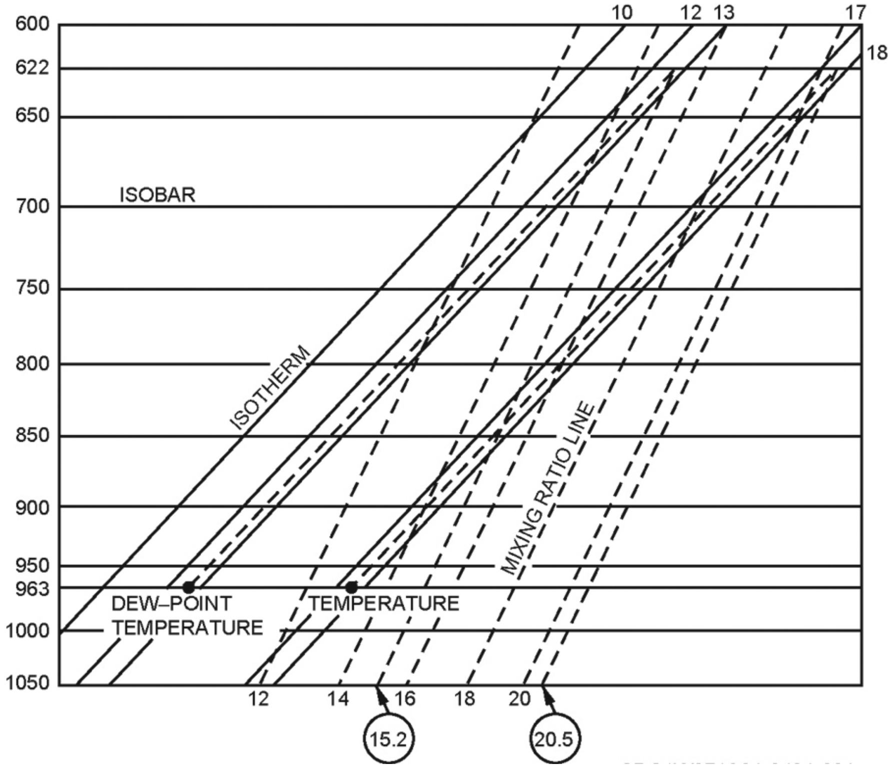

Figure 5–3. Saturation vapor pressure and vapor pressure.

Vapor pressure

The vapor pressure (e) is the actual partial pressure contribution that water vapor makes to the total atmospheric pressure. You find the actual vapor pressure the same way you found the e_s, except that you use the T_d instead of the free-air temperature. In other words, from a *dewpoint* temperature for a given pressure, follow the isotherms to the 622mb isobar. The value of the saturation mixing ratio through this point gives the vapor pressure in millibars at the given pressure.For example, you can see this on a Skew-T by taking a dew point of 12.8°C, with a pressure of 963mb (fig. 5–3). By following the 12.8°C isotherm to the 622mb level and reading the mixing ratio of 15.2 g/kg, we can say that the actual vapor pressure is 15.2mb. You must remember that this vapor pressure applies to the original conditions, which were a dew point of 12.8°C and a pressure of 963mb. The partial pressure exerted by the air is then 947.8mb (963 - 15.2).

Relative humidity

Relative humidity (RH) is the comparison of the actual mixing ratio to the w_S. RH is also known as the ratio of the amount of water vapor in a given parcel to the amount the parcel would hold if it were saturated. This ratio has no units since the units cancel. The result is strictly a percentage figure.

To compute RH, first find the mixing ratio (w) and then find the saturation mixing ratio. Finally, solve the following formula:

$$RH = \frac{w}{w_s} \bullet 100$$

where again: w = mixing ratio and w_s = saturation mixing ratio.

You can compute the RH by knowing the same radiosonde data as before—that is, the pressure, temperature, and T_d. This method uses the Skew-T diagram; however, no plotting is necessary. From the temperature and pressure, you obtain a value for the saturation mixing ratio (w_s) from the dewpoint temperature and pressure, you obtain a value for the w. Let’s look at the previous example, where you used the value 963mb for pressure, 17.2°C for temperature, and 12.8°C for dewpoint temperature (fig. 5–4). The w value at 12.8°C and 963mb is 9.8 g/kg, while the w_s value at 17.2°C is 13.0 g/kg. Using the formula for RH from the previous page:

$$\mathrm{RH} = \frac{9.8\frac{\mathrm{g}}{\mathrm{kg}}}{13.0\frac{\mathrm{g}}{\mathrm{kg}}} \bullet 100$$

we can produce a calculated relative humidity of 75.4 percent.

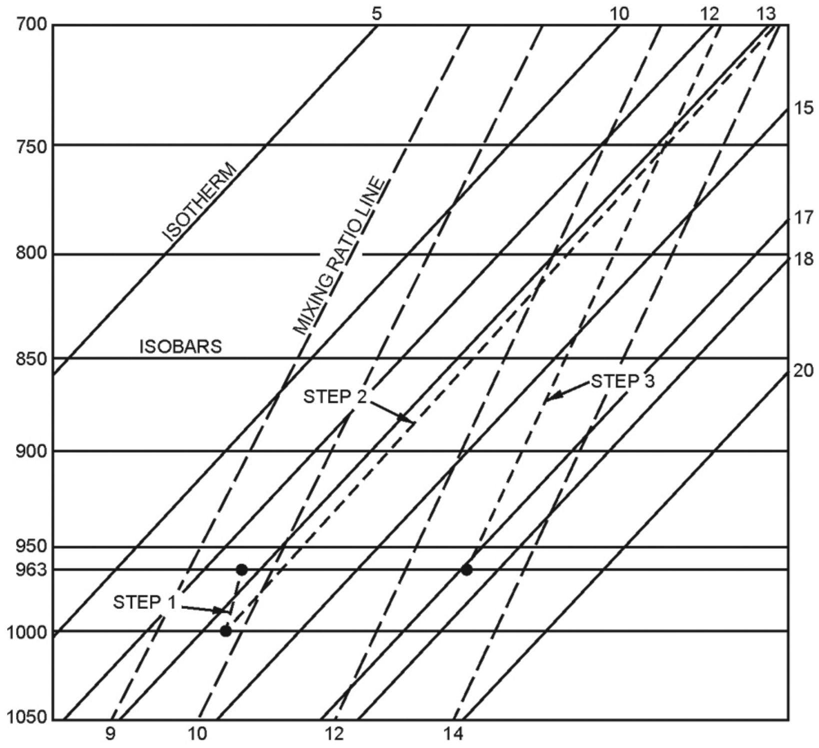

Figure 5–4. Relative humidity.

You can use two methods to compute RH. One method is strictly graphical, and the other is partially graphical and partially mathematical. The graphical method allows you to quickly calculate RH by entering a graph consisting of an X and Y-axis. Normally, one axis has temperature increments along it while the other axis has dewpoint temperature increments along it. The calculation is accomplished by entering the two respective values and following them both horizontally and vertically until they intersect. The intersection will be the RH based on that temperature and dew point. The mathematical method uses the information we previously discussed. The Skew-T acts as the graph, and the formula computes the values from the Skew-T into RH.

Because relative humidity shows how close the air is to saturation, it is dependent on both temperature and water vapor content. Cooling the air with a constant water-vapor content causes the RH to increase. Therefore, cold air has a higher RH than warm air with the same water vapor content.

261. Analyzing temperature parameters

Knowing what the respective lines and curves on a Skew-T represent is very important. Now that we know what these lines and curves mean, we can begin to analyze specific parameters to assist us in forecasting the weather. We begin by discussing the different temperature parameters. We'll discuss what these temperatures are, how to calculate them, and how to use them to assist in forecasting the weather.

Potential temperature

The potential temperature (θ) is the temperature an air sample would have if its pressure were increased to 1,000mb in a dry adiabatic process. By that, we mean the air is compressed to a pressure of 1,000mb with the heat of compression raising the temperature.

Procedure 1

To find the θ of a sample of air, you find the point on the Skew-T that represents its temperature and pressure. The value of the dry adiabat that passes through the point is the θ of the sample of air, labeled in °C.

Procedure 2

A second procedure involves locating the point depicting the temperature and pressure of a sample of air, following the dry adiabat through this point to the 1,000mb level, and reading the temperature there. This temperature is the θ, labeled in °C. To see this, let's compute the θ of a sample of air that has a temperature of 17.2°C and a pressure of 963. Looking at figure 5–5, you can see that we follow the dry adiabat (17.2°C and 963mb) down to the 1,000mb level and then follow the isotherm to the 1,050mb level. At the 1,050mb level, the temperature is 20.3°C, which is the θ of the sample of air. Looking back, you could have found the θ by estimating the value of the dry adiabat passing through the point 17.2°C, 963mb.

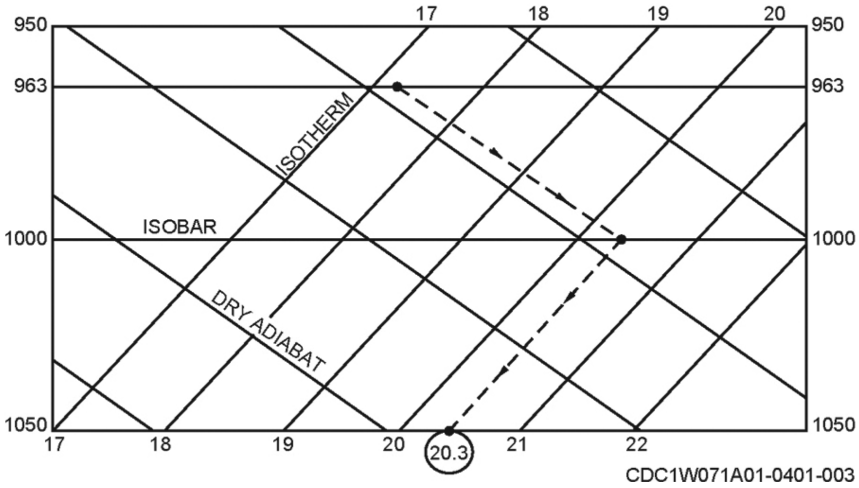

Figure 5–5. Potential temperature.

Wet-bulb temperature

Wet-bulb temperature (Tw) is the lowest temperature to which a parcel of air can be cooled by evaporating water into it at constant pressure. Of course, the heat required for evaporation is supplied by the cooling of the air.

To compute the Tw of a sample of air on the Skew-T, you must know three facts: (1) the temperature, (2) Td or mixing ratio, and (3) the pressure. You plot points at the temperature and Td on the isobar corresponding to the pressure of the air sample. You then draw two lines: one parallel to the mixing ratio lines through the Td and the second line parallel to the dry adiabats through the temperature. A moist adiabat passes through the point of intersection (lifting condensation level discussed later) of these two drawn lines. We follow the moist adiabat back to the original pressure level. The temperature shown by the intersection of the moist adiabat and the pressure is the T_w. The T_w *is always* a value between the T and the T_d.

In figure 5–6, the T_d is 12.8°C and the pressure is 963mb. Follow the 9.8 mixing ratio line until it crosses the 20.3°C dry adiabat line, which passes through the 17.6°C temperature point at 963mb. You see that the 17.9°C moist adiabat passes through this intersection. Follow the 17.9°C moist adiabat back to where the temperature is 14.4°C for the T_w of the sample of air.

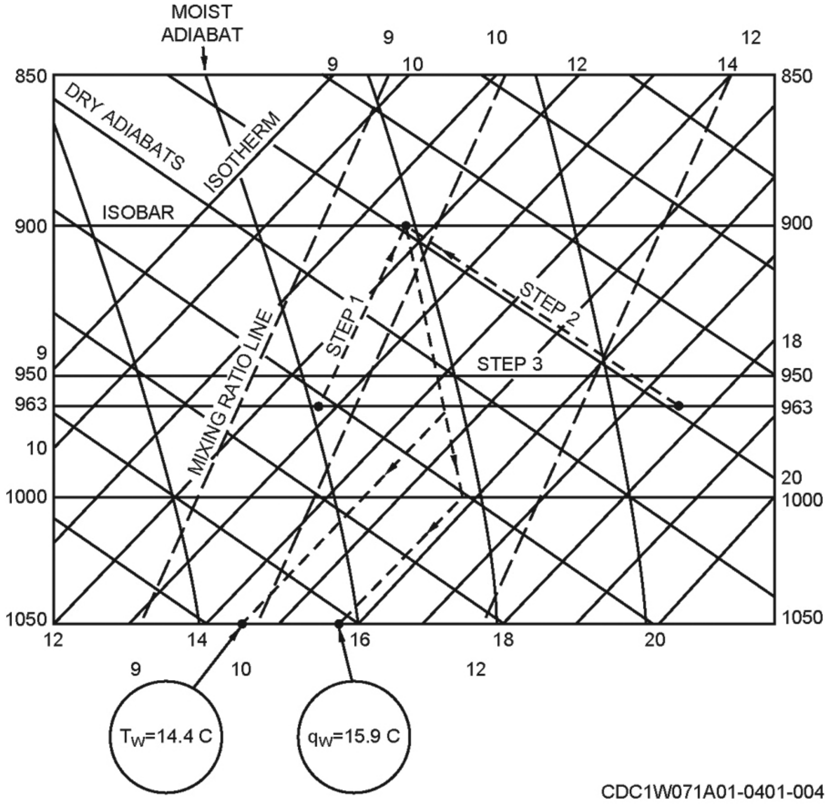

Figure 5–6. Wet-bulb and wet-bulb potential temperature.

The above procedure for calculating the T_w is actually a procedure for finding the pseudo-wet-bulb temperature. We use the two interchangeably because the difference between the two is less than the error involved in measuring the two.

T_w is used in making rain versus snow decisions. Air generally cools to T_w (or close to it) with the onset of precipitation. For example, suppose a situation arises where the temperature = 40°F and T_d = 23°F several hours before precipitation begins. Calculate the T_w and use it as an estimate of the amount of evaporative cooling that will occur when precipitation begins. This is a good indicator of whether the air will cool sufficiently to support snow. Also, the T_w is used in hail forecasting (covered in the severe weather forecasting section in course C).

Wet-bulb potential temperature

The wet-bulb potential temperature (θ_w) is the T_w a parcel of air would have if it were brought moist adiabatically to the 1,000mb level. Since the moist adiabats on the Skew-T are labeled in terms of θ_w, we simply locate the moist adiabat that passes through the T_w of a sample of air and read the value.

In figure 5–6, we have illustrated calculating the T_w of the sample of air that has a temperature of 17.2°C, and a dew point of 12.8°C, and a pressure of 963mb. Using these facts, we find the θ_w to be

15.9°C. Like the T_w, the θ_w that we calculate is the pseudo-wet-bulb potential temperature and can be substituted for the θ_w.

θ_w is used for finding downdraft temperatures of thunderstorms that are related to thunderstorm gust potential. Also, θ_w is used for severe weather indices.

Equivalent temperature

Equivalent temperature (T_e) is the temperature a parcel of air would have if it were cooled both dry and moist adiabatically until all of its moisture was condensed from it and then the sample was heated dry adiabatically to its original pressure. The latent heat of condensation is being used to heat the parcel. This shows the tremendous energy that the condensation of water vapor contributes to the atmosphere. This is assumed to occur by a lifting and sinking process. You can find the T_e of a sample of air by using the Skew-T.

To calculate T_e, follow the point that represents the temperature of the sample as it is lifted adiabatically until saturation is reached (lifted condensation level). Then go up the moist adiabat from the saturation level until the moist adiabats become parallel or nearly parallel to the dry adiabats. The moist adiabats become nearly parallel to the dry adiabats at approximately 200mb. From this level, extend a line down parallel to a dry adiabat to the original pressure original pressure level is reached. The T_e is then read from the value assigned to the isotherm going through the point.

To show this procedure for obtaining the T_e, follow the temperature changes of a sample that at 963mb has a temperature of 17.2°C and a T_d of 12.8°C (fig. 5–7). First from the temperature of 17.2°C, follow a dry adiabat until it crosses the mixing ratio line through the dew point. In our example, this is the 900mb level. From there, follow the moist adiabat up to the 200mb level. From the 200mb level, follow a dry adiabat down until you reach the 963mb level. The T_e is read from the temperature scale at that point and is 42.1°C. Again, this is the pseudo equivalent temperature, but it can be substituted for the T_e.

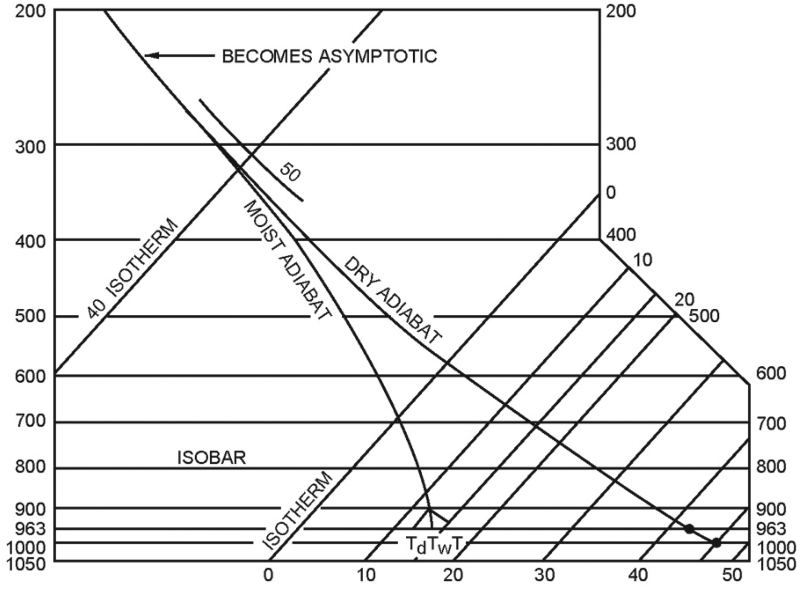

Figure 5–7. Equivalent and equivalent-potential temperature.

T_e is used to trace parcels to their origin in the study of thunderstorms and squall lines. It shows the magnitude of energy that water vapor contributes to the atmosphere.

Equivalent-potential temperature

The equivalent-potential temperature (θ_e) is the temperature a parcel of air would assume if, having reached its T_e, it was warmed by adiabatic compression to 1,000mb. This temperature can be obtained by reading the value of the dry adiabat passing through the T_e position on the Skew-T. Referring to figure 5–7, we have a T_e of 42.1°C; following the dry adiabat to 1,000mb gives us a θ_e reading of 45°C.

Virtual temperature

Virtual temperature (T_v) is the temperature at which a parcel of dry air would have the same density as a parcel of moist air at a given pressure. This is a calculated temperature and not a measured temperature. The T_v is also represented by the symbol T*. Both symbols are used in current publications; however, since T_v is used by most authors, we'll use it in this volume. The difference in the actual and (T_v)—a function of moisture content—can be computed and is approximately 1°C per 6 grams of water vapor. From this we arrive at the formula:

$$T_v = T + \frac{w}{6}$$

To illustrate, let us suppose that we have a sample of air with a T of –5°C and a w of 3 g/kg. The Tv of this parcel of air would be:

$$T_v = -5 + \frac{3}{6}$$

$= -5 + 0.50$

$= -4.5°C$

In most work done on the Skew-T, the observed temperature curve (the actual sounding) is used. In cases where air density is a factor, use the T_v curve. The need for a T_v curve arises in the determination of stability and thickness. This curve may be plotted by adding the required correction to the reported temperatures. The T_v curve is *always* located to the right of the actual sounding curve in the lower levels, and the two will nearly coincide in the upper levels. This shows T_v is always warmer than the environmental temperature, except at levels in the atmosphere where the RH is near zero. T_v is used in thickness computations because it considers the amount of moisture in the air.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

260. Basic components, terms, and definitions

1. Match eachmb level in column B with the associated height in meters in column A. Each item in column B may be used once, more than once, or not at all.

Column A	*Column B*
____ (1) 1500m.	a. 300mb.
____ (2) 3100m.	b. 600mb.
____ (3) 5500m.	c. 500mb.
____ (4) 9300m.	d. 850mb.
	e. 700mb.
	f. 1000mb.

2. How are CAA and WAA identified on the vertical wind profile of a Skew-T?

3. How are isotherms identified on a Skew-T?

4. What's the difference between dry and moist adiabats and how are they located on a Skew-T?

5. What is the thickness of the 850–500mb layer when the 850mb level is at 1510m and the 500mb level is at 5520m?

261. Analyzing temperature parameters

1. If we have a sample of air with a T of –12C and a w of 2g/kg, what would the (T_v) be?

2. What is potential temperature?

3. What is wet-bulb temperature?

4. What is wet-bulb potential temperature?

5. Define equivalent temperature.

6. What is equivalent potential temperature?

7. What is virtual temperature?

5–2. Determine Cloud Formation Parameters

Determining whether the atmosphere is stable or unstable is one of the first factors that you must determine by using an upper-air sounding. By knowing the atmosphere's stability over a region, the you can forecast a number of important features. Some of these features include whether clouds will form, the kinds of clouds that will form, and if precipitation is forecast, whether it will be stratiform or convective in nature. Unit three of this volume introduced how to determine the stability of the atmosphere. In this section we will discuss how to determine the stability of the atmosphere using an upper air sounding.

We also discuss the calculations we use to determine cloud formations due to the heating, lifting, and mixing of the atmosphere and at what respective temperature this occurs. We complete the section by discussing the different types of inversions and how they can be determined on a Skew-T diagram. Inversions are very important in that they can "cap" the atmosphere ultimately leading to explosive convective activity or they can keep the atmosphere in a relatively stable state.

262. Lifting/lifted condensation level

A parcel of air lifted will cool dry adiabatically until condensation occurs. The level at which this occurs is called the lifting/lifted condensation level (LCL). The LCL is the intersection obtained on the Skew-T by drawing the mixing ratio line (which passes through the plotted dew point) and the dry adiabat (which passes through the plotted temperature). In figure 5–8, the parcel, let's say, has a temperature of 17.2°C and a dewpoint temperature of 12.8°C at a pressure of 1,000mb. The intersection, formed by drawing the mixing ratio line passes through 12.8°C and the dry adiabat passes through the 17.2°C, shows the LCL to be at 947mb.

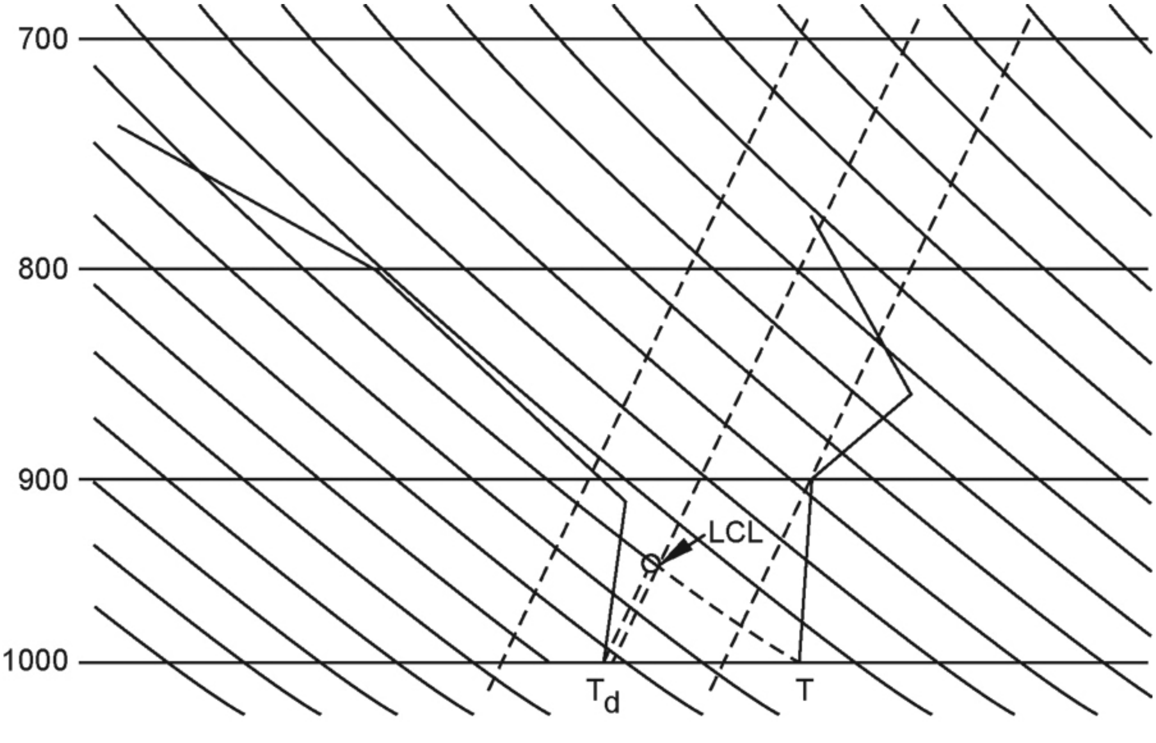

Figure 5–8. Skew-T example of a computed LCL.

263. Convective condensation level

The theoretical level where condensation in an ascending parcel is reached because of convection is called the convective condensation level (CCL). There are two methods of finding the CCL: (1) the parcel method (CCL_p) and (2) the moist layer method (CCL_{ml}). The method chosen depends on how you are going to use it.

Convective condensation level parcel

The first and easiest method is to project the mixing ratio line that passes through the surface dew point until it crosses the free-air temperature curve (fig. 5–9). The point where it intersects the temperature curve of the sounding is the CCL_p. This is the level where the bases of the cumulus clouds form if sufficient heating takes place from the surface level up to the level of the CCL. This is why this CCL is sometimes referred to as the heated method. In figure 5–9, we find the CCL_p to occur at approximately 908mb.

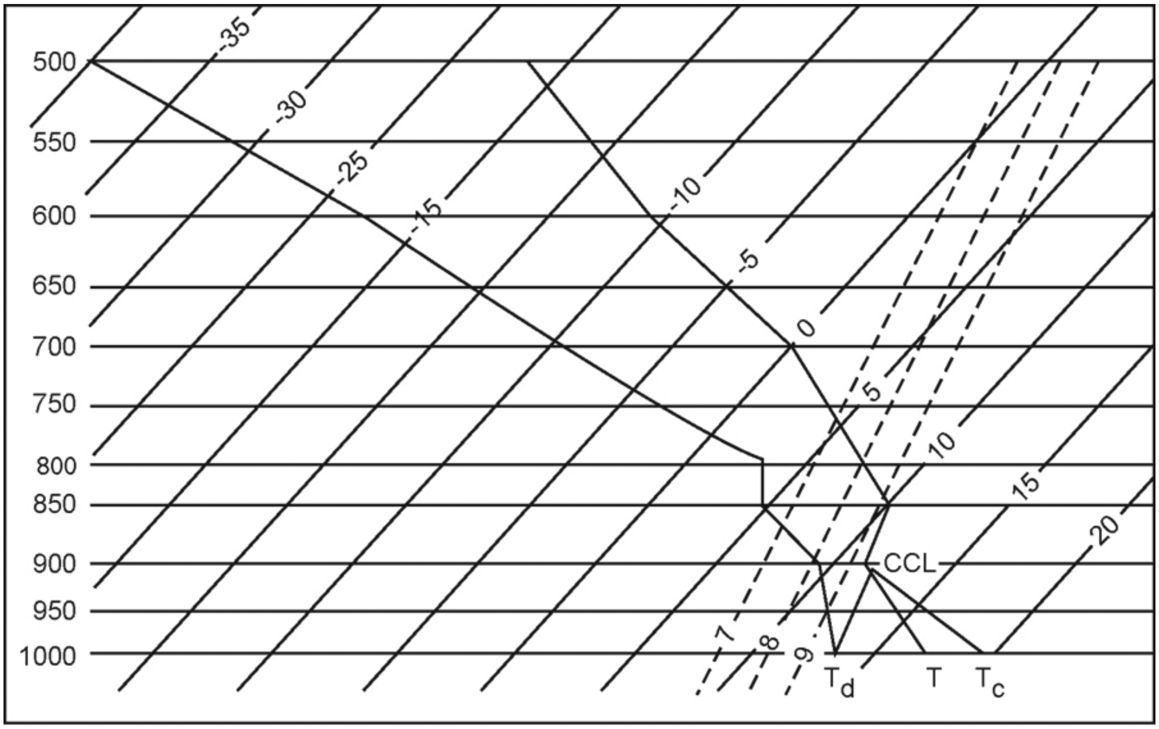

Figure 5–9. Skew-T example of CCL_p and T_c computations.

Convective condensation level moist layer

The second method of finding the CCL was developed to give increased accuracy in severe weather forecasting. Use this method when there is a highly variable moisture content in the lower layers of the troposphere (near the surface). Only use this method when the surface dewpoint depression is 6°C or less. The moist layer is required to have an RH of 65 percent or more at all levels, and in practice we do not use more than the lowest 150mb of the air mass sounding. An alternate method of finding the moist layer is to use a temperature/dewpoint spread of 6°C or less for the 65 percent RH factor.

After finding the depth of the moist layer (or the lower 150mb of the sounding), find the mean (average) saturation mixing ratio line of this layer. You can determine the mean saturation mixing ratio for the moist layer using the equal-area method. The equal-area method requires you to extend a line up parallel to a saturation mixing ratio line, dividing the dewpoint curve into two or more equal areas within the lowest 150mb. This is the mean saturation mixing ratio line. Follow the *mean mixing ratio line* of the moist layer to the point where it crosses the temperature curve of the sounding. The level of intersection is the CCL_{ml}.

264. Mixing condensation level

The mixing condensation level (MCL) is the lowest height in a layer, mixed by turbulence, at which saturation occurs after complete mixing of the layer. This is used for forecasting cloud bases in areas of strong mechanical mixing (i.e., terrain, orographics, etc.).

To determine the MCL, you must first determine the top of the mixing level (layer). There is no known objective way to determine this height. However, a subjective estimate based on local experience which considers such things as the expected lower-level wind speeds, the terrain roughness, and the original sounding through the lower layers, usually suffices.

Once you determine the mixing level, then you need to determine the mean mixing ratio for the mixed layer using the equal area method. Next, determine the mean potential temperature (θ) of the mixed layer by use of the equal area method. If the two lines intersect within the mixed layer, read the isobar. This is the MCL. If the two lines do not intersect, or intersect above the top of the mixed layer, then there is no MCL.

265. Convective temperatures

Convective temperature (T_c) is the temperature that air at the surface must reach by solar heating of the surface air layer. This solar heating initiates convective currents that extend high enough for the air to become saturated and produce convective clouds. We obtain the T_c by following a dry adiabat down from the CCL to the point where the dry adiabat intersects the isobar that represents the surface pressure. The temperature read at this intersection is the T_c. In our example (fig. 5–11), in which we use the surface dew point to find the CCL, we obtain a T_c of 19.7°C (approximately 78.0°F). You can see that a large variation can occur in convective temperatures.

We can also compute the convective temperature using the CCL_{ml} method. This is known as the convective moist layer temperature (T_{cml}) method. Compute the T_{cml} when the moisture content is highly variable in the lower levels near the surface (as we stated previously while discussing CCL_{ml}). The T_{cml} you obtain by following a dry adiabat down parallel from the CCL_{ml} to the point where the dry adiabat intersects the isobar representing the surface pressure. The temperature read at this intersection is the T_{cml}. This, again, is the temperature at which convection begins due to solar insolation.

266. Level of free convection

The level of free convection (LFC) lifted method is the height at which a parcel of air, which is lifted, first becomes warmer (less dense) than the environment. The parcel continues to rise freely until it becomes colder than the environment. The greater the temperature difference between the rising parcel and the surrounding air, the greater upward acceleration the parcel experiences.

Level of free convection lifted method

The LFC lifted method is located graphically on the Skew-T by calculating the mean wet-bulb temperature (T_w) of the moist layer by using the equal area method discussed previously. Then, from the point where it is graphically located, follow the moist adiabat through that point upward until it crosses the temperature curve of the sounding. This point of intersection is the lifted LFC.

For example, using the sounding in figure 5–10, let's compute the lifted LFC. The indicated moist layer extends above 6,000 feet, so we only use the first 150mb of the layer. After constructing the T_w curve for the moist layer (first 150mb, here), we can compute the mean T_w by the equal area method (the shaded area of fig. 5–10). The point at which you divide the T_w into approximate equal areas determines the mean T_w. We then extend a line up parallel to the *saturation adiabat* passing through the indicated mean T_w (shown by a dashed line in the figure) to where it crosses the temperature curve of the sounding. This intersection is the LFC. In figure 5–10, the LFC is shown at approximately the 710mb level of the sounding. However, some conditions occur where the mean T_W will not cross the temperature curve when extended upward. During such conditions, no LFC exists.

Level of free convection heated method

This type of LFC has the same definition as the lifted LFC, except surface heating provides the needed lift. If the parcel is heated, making it rise, then the wet bulb temperature corresponding to the convection temperature must be used.

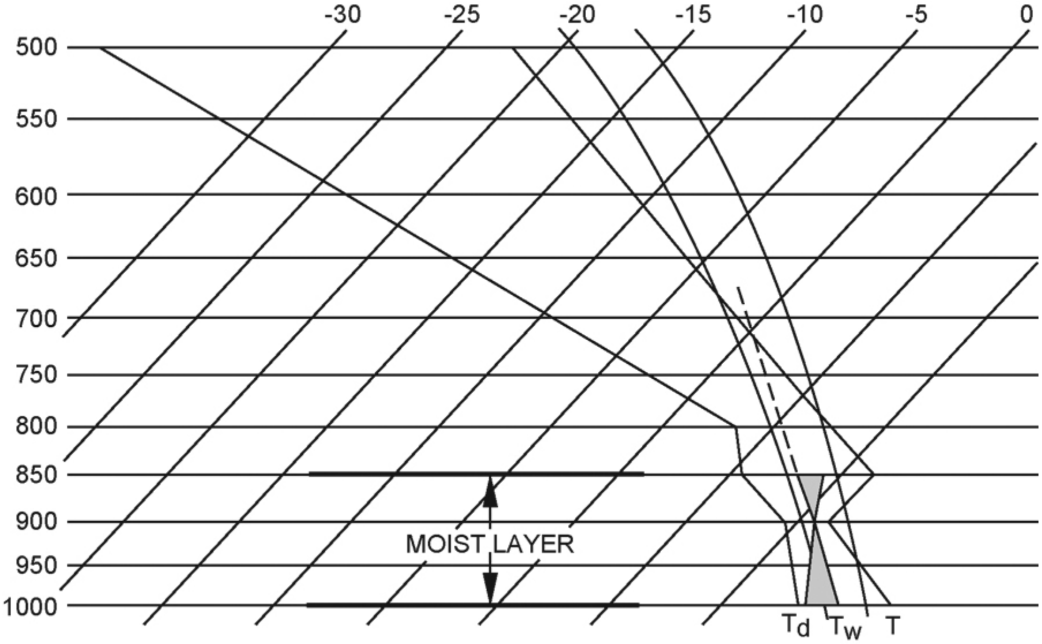

Figure 5–10. Skew-T examples of an equal area method and LFC computation.

267. Types of inversions

An inversion is a layer in the atmosphere where the temperature is either isothermal (remains the same) or increases with an increase in altitude. The lapse rate in the inversion is negative (warms with altitude) and represents a stable layer. The base of the inversion is where the temperature first becomes isothermal or increases. The top of the inversion is where the temperature begins to decrease again. We’ll discuss three types of inversions: (1) radiation inversion, (2) subsidence inversion, and (3) frontal inversion.

Radiation inversion

A radiation inversion is a surface-based inversion formed by rapid cooling of air in contact with the surface of the earth. It occurs in times of maximum radiational cooling, normally just before and after sunrise. Radiation inversions are frequently associated with fog. The mixing ratio is nearly constant through the inversion.

Subsidence inversion

A subsidence inversion is a mechanically produced inversion formed by adiabatic warming of sinking air. It forms in areas of high pressure. The dew point rapidly decreases at the base of the inversion as the sinking air warms and dries it out. This type of inversion is significant in that it suppresses convective activity.

Frontal inversion

The frontal inversion is a transition zone between a cold air mass and the warmer air mass above it. We use it to calculate frontal slope and intensity. The inversion is seen as a layer of weak warming and where the dew point increases through the inversion. The winds will back with altitude through the inversion if a cold front is depicted and the winds will veer with height through the inversion if a warm front is depicted.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

262. Lifting/lifted condensation level

1. What is the LCL?

2. The intersection of what two lines comprises the LCL?

263. Convective condensation level

1. What is the CCL?

2. What determines the method of finding the CCL?

264. Mixing condensation level

1. What is the MCL?

2. What is the MCL used for?

265. Convective temperatures

1. What is T_C?

2. What is T_{CML}?

266. Level of free convection

1. What is the LFC?

2. What is indicated when the T_W will not cross the temperature curve when extended upward?

267. Types of inversions

1. What kind of lapse rate is associated with an inversion?

2. What kind of inversion is surface based?

3. What is the difference between a subsidence inversion and a frontal inversion?

5–3. Severe Weather Indices

Stability indexes are a convenient set of numbers that allow you to rapidly evaluate the instability of an air mass. The higher the number, the higher is the instability of the air mass. You can also evaluate the changes in the atmosphere from one sounding to the next. For instance, a much lower index than 12 hours ago indicates the atmosphere is becoming more stable. All indexes use temperature to look at stability. Temperature is the most destabilizing parameter in the atmosphere.

Use the indices as guidelines when forecasting thunderstorms and severe weather. Often, indices contradict each other and can change rapidly in the course of a few hours. Soundings are most notably changed through thermal advection, moisture advection and evaporational cooling, especially low-level warm air advection, which is most conducive to instability. There are several ways to assess the stability of an air mass by performing calculations from a Skew-T diagram. The most useful indexes to access stability are CAPE, TT, SWEAT, K Index, and the SSI.

268. Convective available potential energy

Convective available potential energy (CAPE) is the buoyancy that a parcel of air has when we compare the temperature of the rising parcel and that of the environment. The CAPE index is an evaluation of the positive area between the parcel and its environment. The positive area on a sounding is proportional to the amount of CAPE. The unit of measurement for CAPE is Joules per kilogram (energy per unit mass). It is the most accurate index to forecast the occurrence and intensity of hail. High CAPE values mean storms build vertically very rapidly. The updraft speeds in these storms depend on the CAPE environment. As CAPE increases, especially above 2,500 J/kg, the hail potential increases dramatically. The following table shows the hail potential based on CAPE increases.

CAPE Hail	Potential
1–1,500	Low
1,500–2,500	ModerateTEST171
greater than 2,500	HighSTQ2

269. Total totals

Total totals (TT) is an index used to determine thunderstorm potential and storm strength. The TT is a combination of the vertical totals (VT) and the cross totals (CT). The VT is the temperature difference between 850 and 500mb. The CT is the 850mb dew point minus the 500mb temperature. TT is an average of the temperature and moisture profile and the potential to form thunderstorms. One pitfall to the total totals index is that it may not pick up a capping inversion that will prevent storms from forming. Also, the index will be too low if the available moisture is confined below the 850mb level.

The following table shows the total total values and the atmospheric potentials.

Total Total Value	Atmospheric Potential
<44	TSTMS Unlikely
44–50	TSTMS Likely
51–52	Isolated Severe
53–56	Scattered Severe
>56	Tornadic

270. Severe weather threat

Developed by the United States Air Force, the severe weather threat (SWEAT) index uses several variables to determine the probability of severe weather. It incorporates the temperature and dew points at both the 500 and 850mb levels, but also adds the wind speeds and directional shear at those levels as well. By doing this, it provides a better look at the potential of thunderstorms to produce tornado events. SWEAT values of about 300 or more indicate a greater potential for severe weather. Values over 400 indicate tornadic conditions.

SWEAT Value	Severe Weather Potential
150–300	TSTMS
300–400	Severe TSTMS
Greater than 400	Tornadic

271. K Index

The K Index (KI) is a combination of the vertical totals (VT) and lower level moisture characteristics. The VT is the temperature difference between the 850 and 500mb level while the moisture parameters used are the 850mb dew point and 700mb dewpoint depression. A high 850mb dew point and a low 700mb dewpoint depression indicates there is a deep layer of warm moist air in the lower to middle troposphere. This is very conducive to producing high levels of instability in the atmosphere. As with the TT index, the K Index may not detect a capping inversion that will prevent storms from forming. This index works fine for flat areas in low to moderate elevations, but does not work very well in mountains.The K index is best used for air-mass thunderstorms. We use this index to determine thunderstorm potential, not the severity of storms. When the VT is high, it causes the KI to be high, even when moisture is lacking. Index values vary with season and location. The table below shows the K-Index value and the convective potential.

K-Index Value	Convective Potential
15–25	Small Convective Potential
26–39	Moderate Convective Potential
40+	High Convective Potential

272. Showalter stability index

This index is simply the difference between the temperature of the air at the 500mb level and the temperature a sample of air at the 850mb level would have if it were lifted to the 500mb level. We use it to assess the instability of the 850mb parcel. The more negative the showalter stability index (SSI), the more unstable and buoyant the lower levels are. The SSI is great to use when there is a shallow layer of polar air in the planetary boundary layer (PBL). In those cases, lifting occurs above the polar air. The table below shows SSI values and their corresponding potentials.

SSI Value	Potential
–1 to –4	TSTMS
–5 to –7	Severe TSTMS
–8 or less	Tornadic

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

268. Convective available potential energy

1. What is CAPE?

2. What happens when the CAPE goes above 2,500 J/kg?

269. Total totals

1. What comprises the TT index?

2. What are the two main parameters we look at for the TT index?

270. Severe weather threat

1. What makes up the SWEAT index?

2. What does the SWEAT index evaluate?

271. K-Index

1. What moisture parameters are incorporated in the K Index?

2. Where would the K index be least effective: Patrick AFB, Scott AFB, Peterson AFB or Ft. Lewis?

272. Showalter stability index

1. What does a negative SSI indicate?

2. A Showalter's stability index of -9 indicates the potential for what type of weather?

5–4. Forecasting Hail and Convective Wind Gusts

Hail is regarded as one of the worst hazards of flying in thunderstorms. It usually occurs during the mature stage of cells having an updraft of more than average intensity and is found with the greatest frequency between 10,000 and 15,000 ft. As a rule, the higher the vertical extent of the storm's updraft, the more likely it is to have hail.

Wind gusts are also extremely hazardous to aircraft in flight. As we previously discussed, convective gusts in the form of outflow boundaries, downbursts, and microbursts are a common occurrence with thunderstorms. Aircraft are especially sensitive to gusty winds during take-off and landing. We'll now discuss how to forecast both hail and gusty winds associated with thunderstorms.

273. Hail occurrence

Hail, like the maximum wind gusts in thunderstorms, usually takes place in a narrow shaft which is seldom wider than a mile or two and usually less than a mile wide. Let's look at a few facts concerning the occurrence and frequency of hail in flight.

Since hail is normally associated with thunderstorms, the season of the maximum occurrence of hail coincides with the season of maximum occurrence of thunderstorms. When the storm is large and well developed, assume that it contains hail. Encounters of hail below 10,000 ft showed the hail distribution equally divided between the clear air alongside the thunderstorm, in the rain area below the storm, and within the thunderstorm itself. Ninety percent of the cases showed that the aircraft were within two miles of the storm, in it, or beneath it.

From 10,000 to 20,000 ft, the percentages ranged from 40 percent in the clear air alongside the storm to 60 percent in the storm, with 85 percent of the encounters outside the storm under the overhanging cloud. Above 20,000 ft, 80 percent of the hail was encountered in the storm with 20 percent in the clear air beneath the anvil or other clouds extending from the storm.

Climatology is important in predicting the hail occurrence, as well as its size. You can glean good estimates of the size of the hail from reports of the storm passage over nearby stations. Of course, you must consider modifying influences. It follows, then, that climatology of hail occurrences for your station and local area can give a good indication of what has taken place in the past and can serve as a guide for what is likely to occur with thunderstorm passage.

274. Predicting surface hail

Besides the fact that hail occurs with thunderstorms (both inside and outside the storm) and despite our knowledge of the relationship of hail to this type of storm, very little information is available as to forecasting the actual occurrence of hail. The following is an objective method of forecasting hail, using the parameters of the ratio of cloud depth below the freezing level and the height of the cloud top. The data used in this study were derived from 70 severe convective storms (34 hail producing and 36 non-hail producing storms) over the Midwest states.

Severe thunderstorms used in developing this technique were those thunderstorms causing measurable property damage due to strong winds, lightning, or heavy rain. Severe thunderstorms with accompanying hail were thunderstorms where the hail was listed as the prime cause of property damage although other phenomena may also have occurred. All tornadoes were excluded from consideration to avoid confusion. To determine whether hail will occur at the surface:

- Calculate the CCL.
- Calculate the EL.
- Determine the freezing level.

NOTE: All three heights are expressed in feet or millibars.

Determine the ratio of the cloud depth from the CCL to the freezing level in millibars and from the CCL to the clouds' estimated maximum vertical development in millibars. For example, if the CCL is at 760mb, the freezing level is at 620mb, and the EL is at 220mb, compute the cloud depth ratio as follows:

Cloud depth ratio = 140/540 = 0.26. Using the freezing level and the cloud depth ratio from above look at figure 5–11 to determine if hail should be forcast. (In the above example, hail would be forecast.)

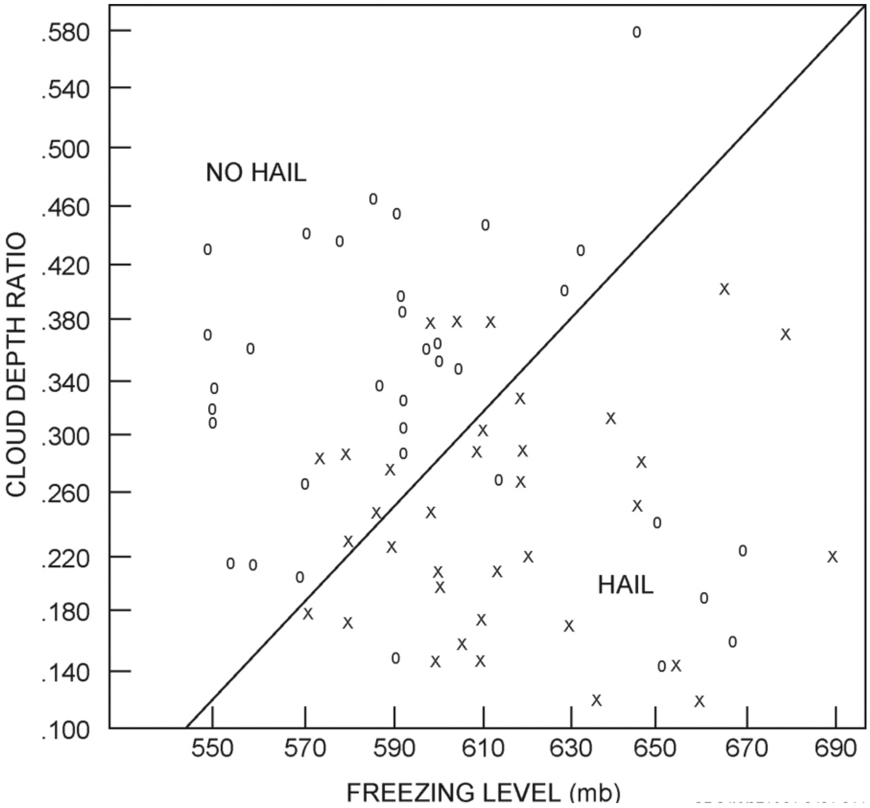

Figure 5–11. Hail scatter diagram.

275. Predicting hail size

As we discussed earlier, the incidence of large hail at the surface seems related to an optimum wet-bulb zero (WBZ) height around 8,000 ft above the terrain. Similarly, the incidence of large hail at the surface decreases sharply with WBZ heights above 10,500 ft. You can also use the height of the WBZ, then, as a preliminary indicator of possible hail occurrence at the surface. A point to remember is that an aircraft may encounter hail at *any* altitude in or near the thunderstorm.

The technique we use to forecast the size of hail involves a diagram based on wind tunnel tests and estimates of the updraft velocities in thunderstorms. Again, we derive the parameters necessary for entering the diagram from analysis of the Skew-T sounding. The first step is to determine the CCL. The parcel temperature at the CCL is projected upward moist adiabatically to the pressure level where the free-air temperature is –5°C. In the sounding shown in figure 5–12, the CCL is at point A. The moist adiabat from the CCL to the pressure level where the ambient temperature is –5°C is shown by the line AB’. The isobar from the point where the air temperature is –5°C to its intersection with the moist adiabat (AB’) is the line BB’. The triangle formed by connecting points A, B, and B’ represents a “positive” area indicative of the updraft velocity potential of the air column. The dry adiabat from the isobar BB’ through the triangle to the pressure level of the CCL is the line HH’.

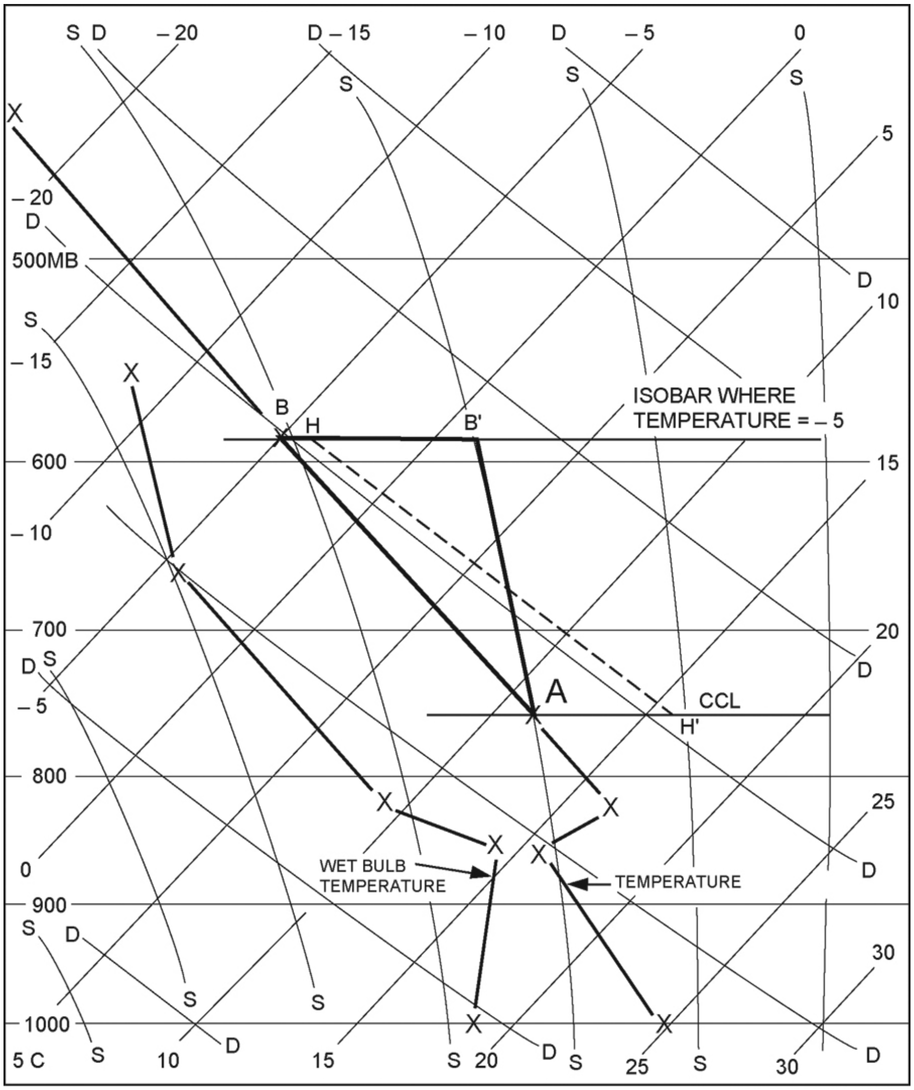

Figure 5–12. Example evaluation of sounding for hail.

The hail graph illustrated in figure 5–13 yields estimates of hail sizes based on the dimensions of the positive triangle measured in Celsius degrees. The horizontal coordinate in figure 5–13 is the length of the base of the positive triangle while the vertical coordinate is the length of the dry adiabat through the triangle. In figure 5–12, the base of the triangle, BB', is 6° (from +1° to –5°). The length of the dry adiabat through the triangle is 21° (from –5° to +16°). Looking at the hail graph with a horizontal coordinate of 6 and a vertical coordinate of 21 suggests that a forecast hail size of 1 inch would be most appropriate.

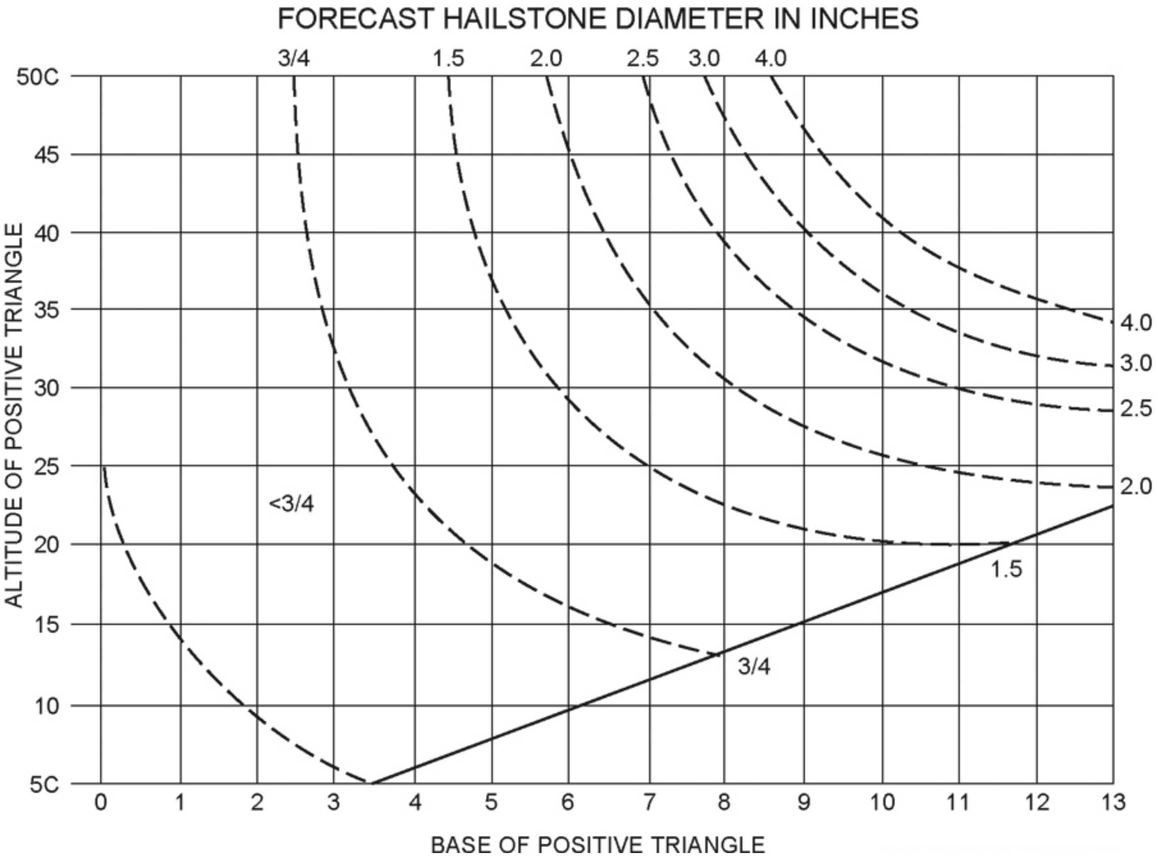

Figure 5–13. Forecasting hail size graph.

In the Gulf States, or in any air mass where the WBZ is above 10,500 ft, the hail size derived from the graph in figure 5–13 is too large. An additional graph was devised to determine hail size corrections as a function of WBZ heights above 10,500 ft. The correction graph is shown in figure 5–14. The horizontal coordinate of the graph represents the hail size derived from the graph in figure 5–13. Determine the corrected hail size by plotting the graph with the original hail size and the height of the WBZ (fig. 5–14). For example, an original hail size of 1 inch is reduced to ¼-inch hail forecast if the WBZ height is 11,800 ft above the ground.

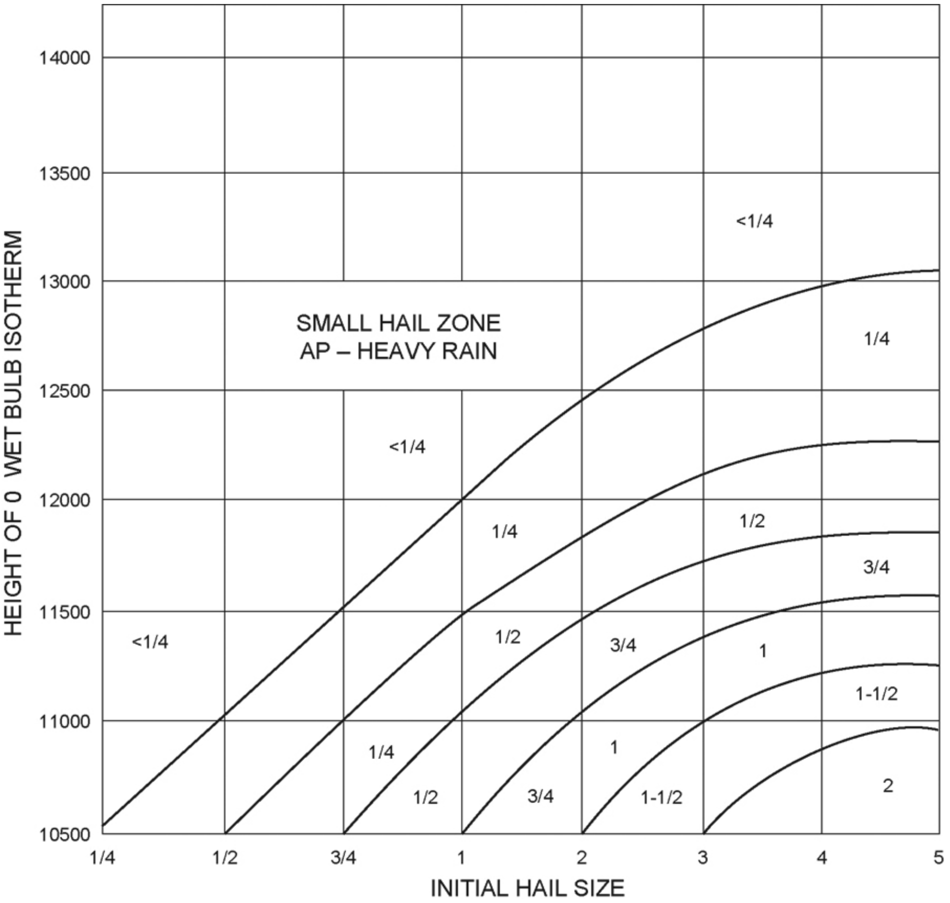

Figure 5–14. Hail size correction nomogram.

276. Forecasting maximum wind gusts

Remember that the 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. The probability of verifying such a forecast is comparatively low unless the thunderstorm is going to pass close to or directly over your station. Nevertheless, the possibility of damage to aircraft and installations is so great that you must use every available means to make the best and most accurate forecasts possible to forewarn the agencies concerned.

The most severe surface weather appears to occur within areas in which the height of the WBZ is less than 10,500 ft. An isopleth representing the 10,500 ft WBZ height effectively outlines this maximum threat area. WBZ heights between 7,000 and 9,000 ft above the terrain are most closely associated with the occurrences of destructive surface winds. It follows, then, that the height of the WBZ is a reliable preliminary indicator of the possibility of severe weather occurrence at the surface. The geographic distribution of WBZ heights identifying the areas of maximum threat is revealed through evaluation of the many atmospheric soundings available in the US. Those soundings on which the WBZ heights fall between 7,000 and 9,000 ft are subjected to additional empirical techniques devised at Air Force Weather Agency (AFWA).

There are two techniques used by the AFWA in forecasting maximum wind gusts associated with thunderstorms. One technique considers the difference between a parcel temperature raised moist adiabatically and the ambient temperature at 600mb. This temperature difference is designated at T_1. If the sounding under consideration has an inversion, the warmest point on the sounding is raised moist adiabatically to the 600-mb level. Figure 5–15 is an example of such a sounding. Note that the temperature of the raised parcel is 0.8°C and the ambient temperature at 600mb is –7.8° representing a T_1 of 8.6°C. An empirical formula was devised to express the relationship between this 600mb temperature difference and surface wind gusts.

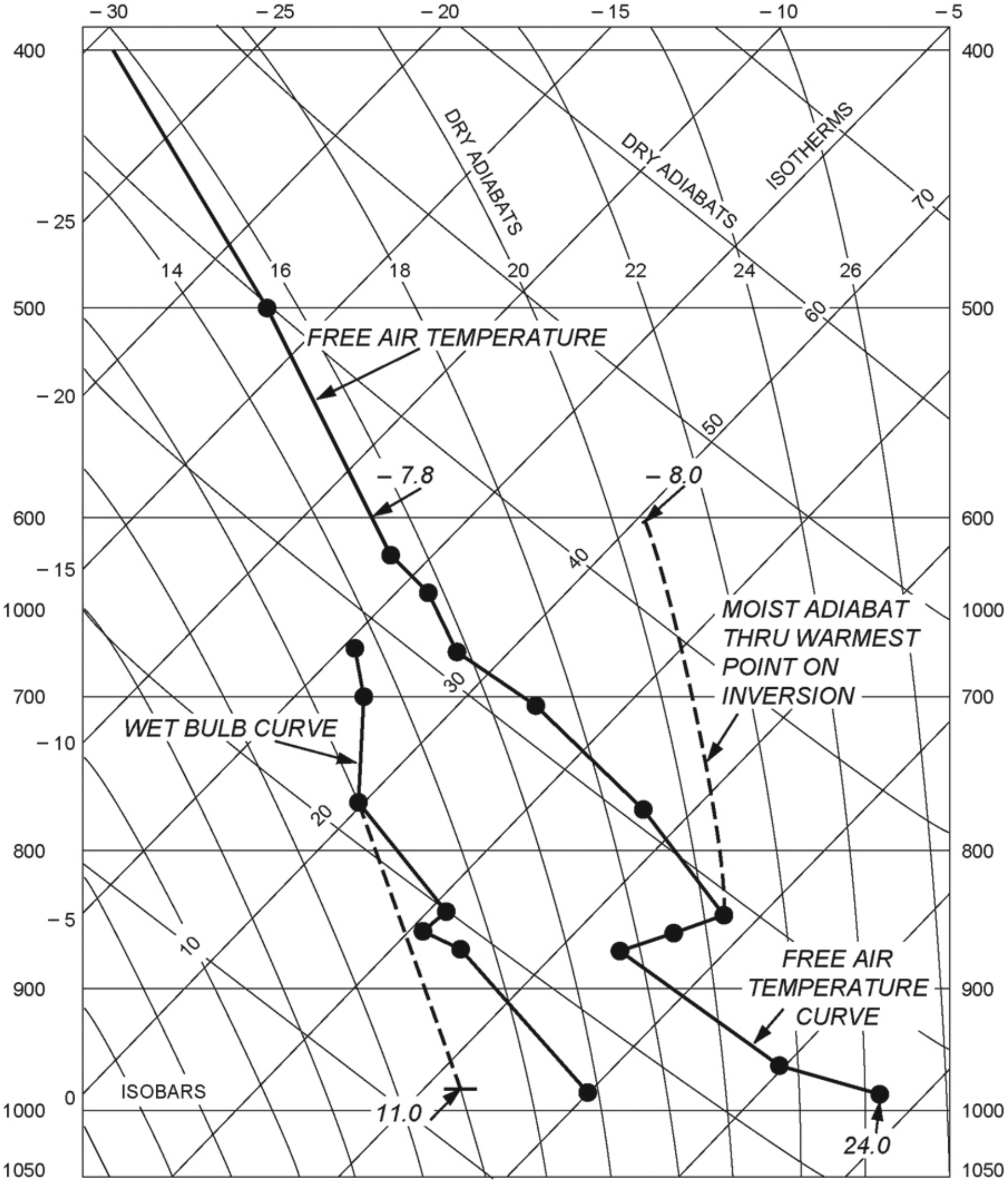

Figure 5–15. Example sounding evaluation for wind gust forecast.

The formula is $V = 13(\sqrt{\ T_1})$ where V represents the gust speed and T_1 the 600mb temperature difference. To further simplify gust computation, a table was prepared based on this empirical formula for a range of possible T_1 values. These T_1 values and their equivalent gust speeds (empirically modified) are given in the following table.

T_1 Values in °C	3	4	5	6	7	8	9	10	11	12	13	14
Max Gust Speed	17	20	23	26	29	32	35	37	39	41	45	47
T_1 Values in °C	15	16	17	18	19	20	21	22	23	24	25	
Max Gust Speed	49	51	53	55	57	58	60	61	63	64	65	

We use the warmest point on the inversion to determine the T_1 *if* that warmest point is within 150 to 200mb from the surface and surface convection does *not* dissipate the inversion. If the inversion is comparatively high (more than 200mb above the surface); or if surface convection dissipates the inversion or if no inversion is present; we use a different method to find T_1. In these latter cases, we use the maximum temperature expected at the surface as a basis for determining the T_1. The T_1 is then the difference between the free-air temperature at 600mb and that of a parcel projected moist adiabatically from the forecast maximum surface temperature. This yields a larger value for T_1, resulting in a higher gust forecast.

This method is quite reliable in showing maximum *average* wind gusts. To calculate the probable maximum peak gust, add one-third of the *mean* wind speed expected in the lowest 5,000 ft AGL to the average value obtained. The mean wind direction in the layer between 10,000 and 14,000 ft AGL has been found to closely approximate the direction of maximum gusts at the surface; use it in forecasting gust direction.

Back on figure 5–15, the T_1 was found to be 8.6°C. From the tabular values for T_1, 8.6° is rounded to 9° and indicates a gust speed of 35kt. Assuming a mean wind speed of 30kt in the lower 5,000 ft AGL, one-third of that wind, or 10kt, added to the maximum average gust results in a peak gust speed of 45kt. AFWA has found this method of gust forecasting to be quite representative of the gusts experienced in scattered thunderstorm situations in which the storms are near a station but do not necessarily pass directly overhead.

In the event the passage of an intense squall line or numerous thunderstorms is anticipated, greatly increasing the likelihood of a station being directly affected, AFGWC considers an alternate technique in wind gust forecasting. This alternate technique, T_2 is built around the graph in figure 5–16. The first step is to locate the 0°C isotherm on the wet-bulb curve of the sounding. In figure 5–15, the point of intersection of the wet-bulb curve and the 0°C isotherm projected downward moist adiabatically to the surface resulting in a “down rush” temperature of 11°C. The surface free air temperature is 24°C. The difference between this down rush temperature in centigrade degrees and the surface free-air temperature or forecast maximum temperature is termed T_2. Here, T_2 equals 13°C. Looking at the graph with a T_2 of 13 shows that the range of maximum gusts falls between 33 and 47kt, with a mean of approximately 40kt.

It is important to remember that this latter method indicates the expected maximum gusts. The thunderstorm must pass over the forecast point and moderate to heavy rain must occur, except in a type IV sounding for a forecaster to use the T_2 wind graph. Since these unique conditions must be met, the method may appear to over-forecast at times.

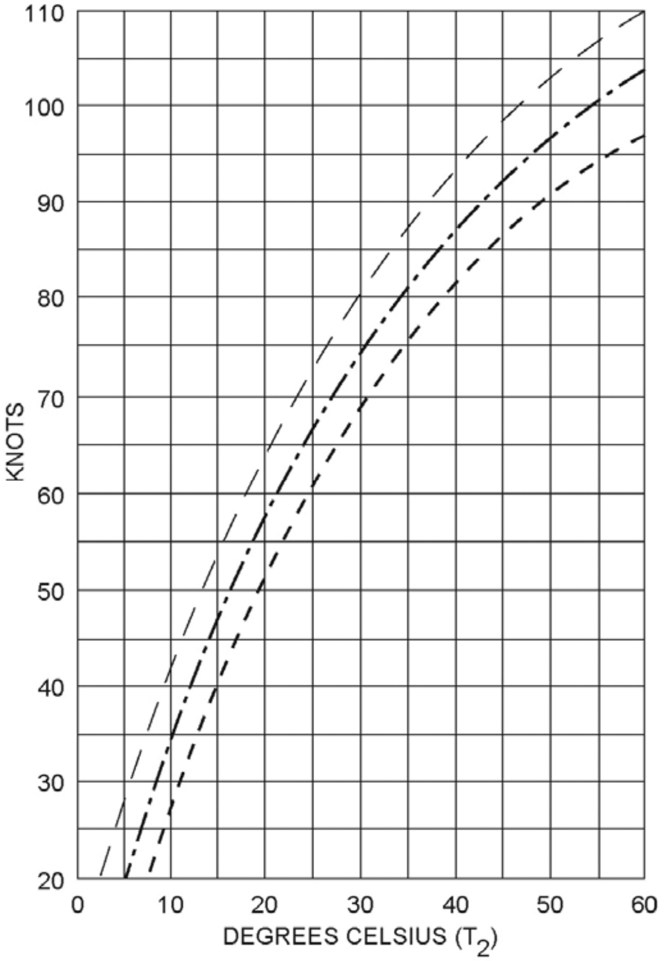

Figure 5–16. T_2 Wind gust graph.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

273. Hail occurrence

1. Below 10,000 ft, what is the hail distribution between areas in the storm and areas beneath the overhang of the cloud?

2. Between 10,000 and 20,000 ft, what is the hail distribution?

3. Above 20,000 ft, what is the hail distribution in the storm and in the clear air beneath the anvil or overhang of the cloud?

274. Predicting surface hail

1. Do you forecast hail if the CCL is at 830mb, the freezing level is at 700mb, and the EL is at 150mb?

2. Do you forecast hail if the CCL is at 795mb, the freezing level is at 590mb, and the EL is at 210mb?

275. Predicting hail size

1. What is an important point to remember about hail and aircraft?
2. What is the first step in forecasting hail?
3. Using figure 5–13, what size hail would you forecast with a base of the positive triangle as 7 and a length (or height) through the triangle of 35?

276. Forecasting maximum wind gusts

1. If a sounding on a Skew-T has an inversion, what point on the inversion do you use to calculate T_1?
2. What layer closely approximates the direction of maximum gusts at the surface?
3. What is important to remember in using the T_2 method?

5–5. Determine Turbulence Indicators from Skew-T Data

We cannot overemphasize the importance of turbulence forecasting to the flying customer. The impact of forecasting and classifying turbulence, however, is a challenge. The difficulty arises because factors creating turbulence in one instance may not cause turbulence in a similar situation. Complicating matters further is that while one aircraft may report “smooth sailing,” minutes later, another aircraft flying through the same airspace may report significant turbulence. Turbulence can rip an aircraft apart in flight, damage the airframe, and cause injury. Therefore, accurate turbulence forecasts are an important part of aviation weather briefings. If you as a forecaster understand the basics of atmospheric turbulence, you can better analyze and forecast this dangerous phenomenon.

277. Frontal zone characteristics

A front can’t exist unless two air masses of differing characteristics or properties are brought next to each other. To identify a front, you need to know the elements expected in the frontal zone. These characteristics tell you a lot about what weather to forecast for your station.

In this lesson, we’ll cover six front characteristics:

1. Types of fronts.
2. Thermal structure.
3. Humidity characteristics.

4. Vertical wind distribution.
5. Horizontal wind distribution.
6. Streamlines associated with frontal systems.

Types of fronts

A front or frontal zone marks the boundary between adjacent air masses. Whether you classify a front as warm or cold depends on the direction of movement of the air mass and the relative temperatures of the air masses involved. The boundary associated with a cold air mass displacing a warm air mass is a cold front. When a cold air mass retreats and is replaced by warmer air, the boundary between the air masses is a warm front. For example, a cold front precedes cP air moving out of Canada into the central US. When this cold air mass moves eastward, a warm front marks its trailing edge because warmer air is replacing the retreating colder air mass. Remember, because cold air is denser it can displace or move warm air out of the way. However, warm air can’t displace cold air; instead, it can only replace retreating cold air.

Thermal structure

A significant feature in frontal identification is the lapse rate through a front. The lapse rate in the frontal zone shows a strong stabilization or warming. Consequently, the frontal zone resists vertical exchanges of heat and moisture between the air masses on either side of the front. Figure 5–17 illustrates an idealized lapse rate through a frontal zone. An actual sounding might show a less definite frontal zone and many minor variations in the lapse rate. The sharpness and the extent of the frontal inversion show the height, strength, and degree of mixing between the air masses.

While frontal zones are depicted as blue or red lines on a weather product, there’s much more to understanding them. The frontal zone is an area of varying width that acts as a buffer or transition zone between the two types of air masses. Fronts are also zones of low pressure; as such, they’re associated with convergence and cyclonic circulation patterns at the surface. Frontal zones have certain other characteristics that aid in their identification and help to explain the type of weather that forms because of air mass movement.

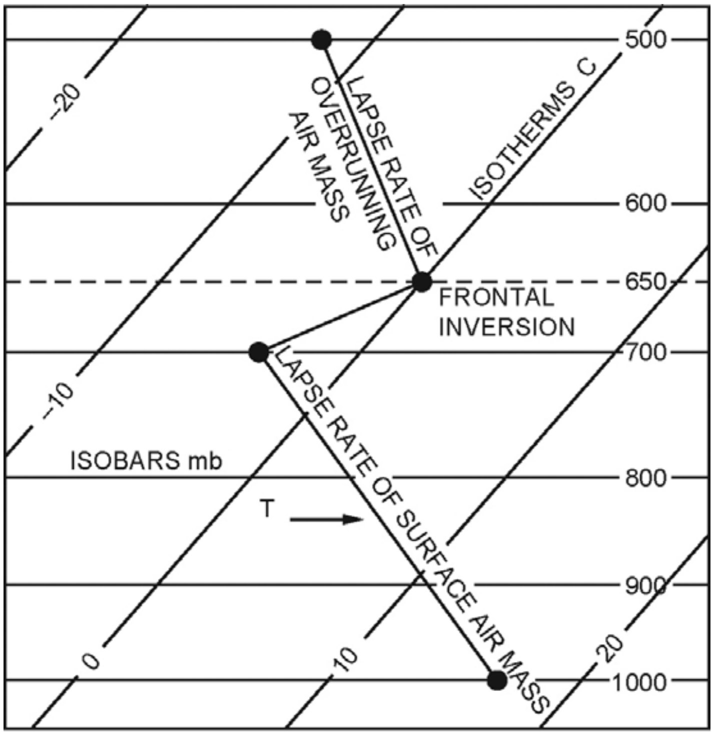

Figure 5–17. Temperature inversion through a front.

Humidity characteristics

Figure 5–18 illustrates the classical humidity pattern through a frontal zone. The dew point (Td) usually increases sharply with the inversion in the temperature curve indicating stabilization. Because of the many variations in the temperature curve, the dewpoint curve better indicates the frontal zone.

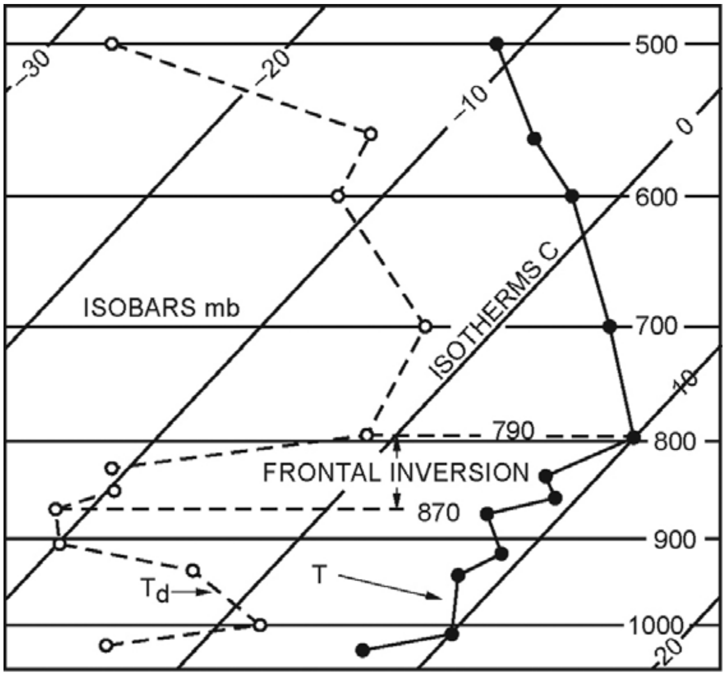

Figure 5–18. Dewpoint temperature curves through a frontal zone.

In association with a subsidence inversion, the dewpoint curve shows an entirely different trend. The dynamic warming of air as it subsides reduces the relative humidity. The dew point normally drops off to an undetectable value through a strong subsidence inversion. Figure 5–19 illustrates the characteristics of the dewpoint curve through a subsidence inversion and a frontal inversion as they might appear in association following a cold frontal passage. The usual subsidence within southward moving cold air often creates one or more inversions below the frontal inversion. These lower discontinuities can be mistaken for frontal zones, or may be so close to the frontal zone that only one deep inversion is evident. On the sounding shown in figure 5–19, the dewpoint curve identifies the inversion between 870 and 820mb as one caused by subsidence. If precipitation should fall from above, however, the raising of the dew point would leave little or no clear indication of the subsidence origin of the layer.

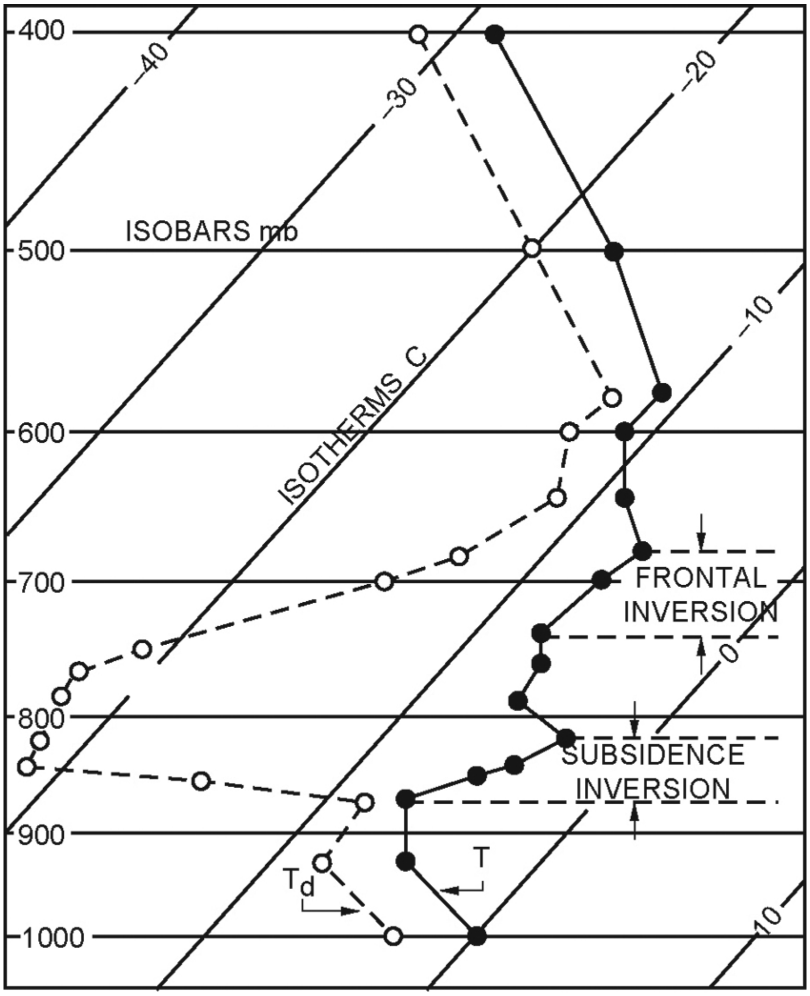

Figure 5–19. Frontal and subsidence inversion.

Vertical wind distribution

The best sources for identifying frontal inversions are a combination of temperature and dewpointcurves coupled with vertical wind distribution. Clues are provided by vertical wind direction changes as you view the Skew-T from the lower levels up through the upper levels. The wind direction backs (changes counterclockwise) with altitude through inversions associated with cold fronts and veers (changes clockwise) with altitude through inversions associated with warm fronts. In figure 5–20, look at the wind changes from the 800 to the 650mb levels. The winds veer (southwesterly changing to northwesterly) between the two levels; this identifies this as a warm frontal zone.

The boundary zone between two sharply differing air masses is usually accompanied by clear discontinuities in temperature, dew point, and vertical wind distribution. You must consider each of these parameters for the greatest degree of certainty in frontal zone identification. Any parameter by itself might prove inconclusive. For example, a backing wind in the vertical might suggest cold-air advection not associated with a cold front. Conversely, a veering wind in the vertical might suggest warm-air advection not associated with a warm front.

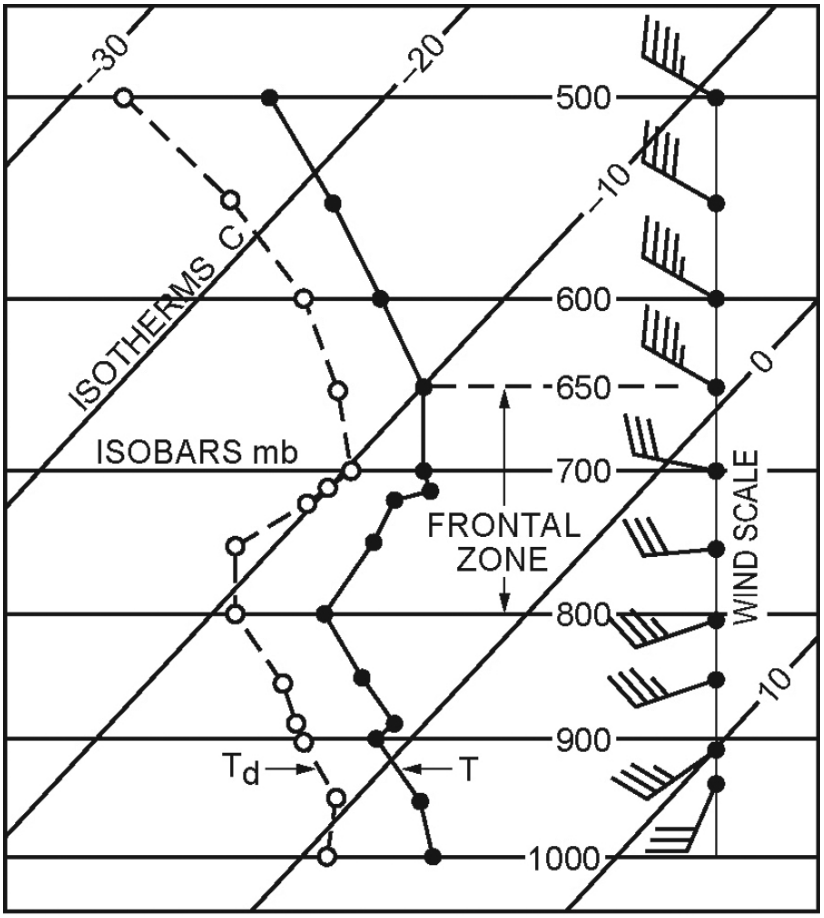

Figure 5–20. Vertical wind distribution through a frontal inversion.

Horizontal wind distribution

The vertical discontinuities of a frontal zone are also represented in the horizontal. Besides temperature and dew point discontinuities, an easily identifiable wind shift usually accompanies a frontal passage at the surface (particularly a cold frontal passage). The horizontal shift in wind direction is in a clockwise or veering sense (in the Northern Hemisphere) because fronts lie in a trough of low pressure. Usually, wind speed increases when a wind shift occurs. Greater horizontal discontinuities accompany cold frontal passages. Direction and speed discontinuities associated with a warm frontal passage are usually not as sharp, but are normally identifiable.

Streamlines associated with frontal systems

A streamline is a line showing the direction all air parcels are moving at any instant in time. A baroclinic low has three such streamlines we call conveyor belts. Figure 5–21 illustrates the three different airflow patterns that flow around and through a developing baroclinic low. These different patterns are composed of three conveyor belts: 1) warm, 2) cold, and 3) dry-air.

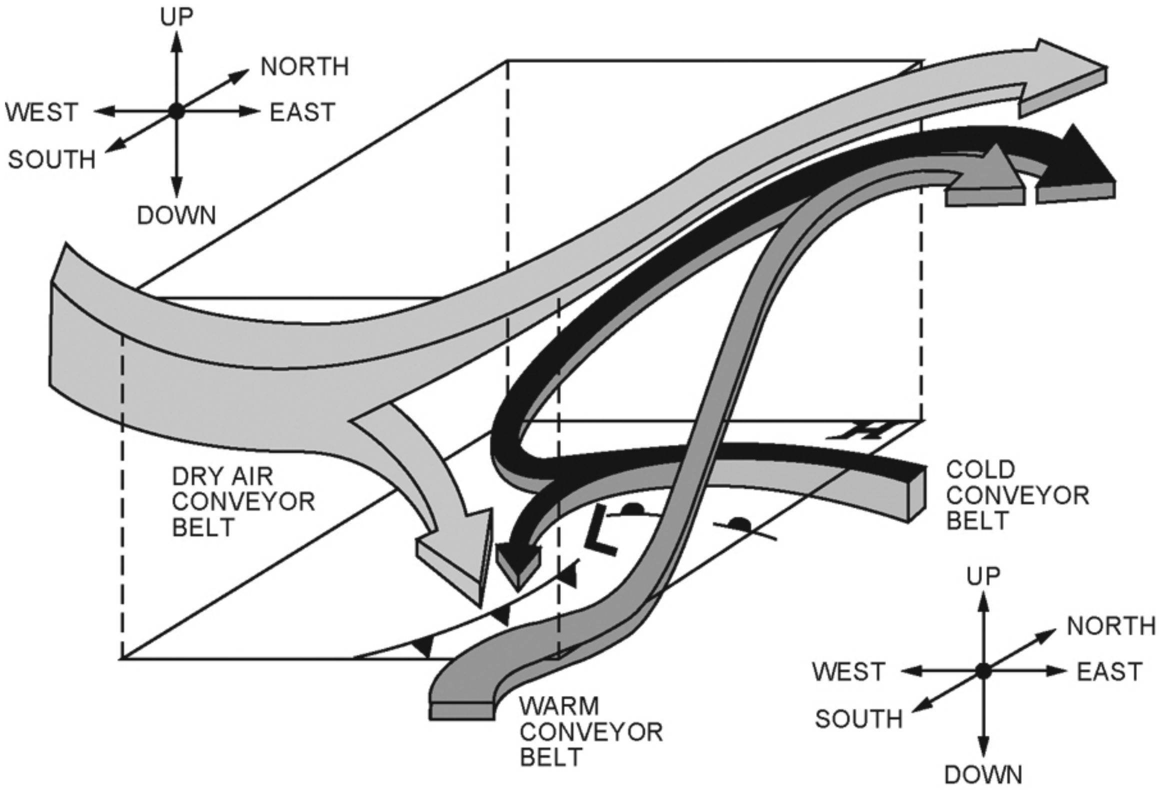

Figure 5–21. Conveyor belts.

Warm conveyor belt

The warm conveyor belt is a set of streamlines that originates at low levels in the moist tropical air mass equatorward of the surface low center. The warm conveyor belt flows northward, ascends, and turns anticyclonically at the jet-stream level. The strongest ascent occurs near the surface low. Baroclinic zone cirrus forms within the northern portion of the conveyor belt, where the warm moist air is being lifted.

The warm conveyor belt flows northward ahead of the cold front in one of two configurations. If the front is an active cold front, the warm conveyor belt ascends the frontal surface. This is called *rearward sloping ascent*. If the front is an inactive cold front, the warm conveyor belt won't rise over the frontal surface. The strongest lift is over the warm frontal surface. This is called *forward sloping ascent*.

Cold conveyor belt

The cold conveyor belt originates in the low levels in the cold air east of a low center. The cold conveyor belt flows westward beneath the warm conveyor belt and is associated with subsidence well ahead of the low center. As it nears the low center, it ascends rapidly and emerges from beneath the warm conveyor belt west of the low center. This strong ascent is associated with widespread cloudiness and precipitation. These clouds compose the "head" of the comma-shaped cloud system we see on satellite imagery.

Once west of the low center, the cold conveyor belt may take one of two paths. It may continue to rise, turn sharply anticyclonically, and join the warm conveyor belt in the upper levels, or it may turn cyclonically and descend well west of the low center.

Dry-air conveyor belt

The dry-air conveyor belt originates at upper levels, upstream from the major short-wave trough supporting the low. As the dry-air conveyor belt approaches the short-wave trough, it undergoes strong subsidence and drying. As it nears the low, it splits into two branches. One of these branches turns cyclonically and flows north and west of the warm conveyor belt. The boundary between the two conveyor belts is visible as the smooth high cloud border north of the low. The other branch turns anticyclonically and descends to low levels well behind the low. This branch of the dry-air conveyor belt helps to bring cold air from the upper levels, down to the surface behind the cold front.

278. Forecasting turbulence in convective clouds

This lesson describes a method for forecasting turbulence in convective clouds using a Skew-T. This method considers two layers of the atmosphere: surface to 9,000 feet MSL and above 9,000 feet MSL (fig. 5–22). The forecast is designed for Category II aircraft and must be modified for other types of aircraft.

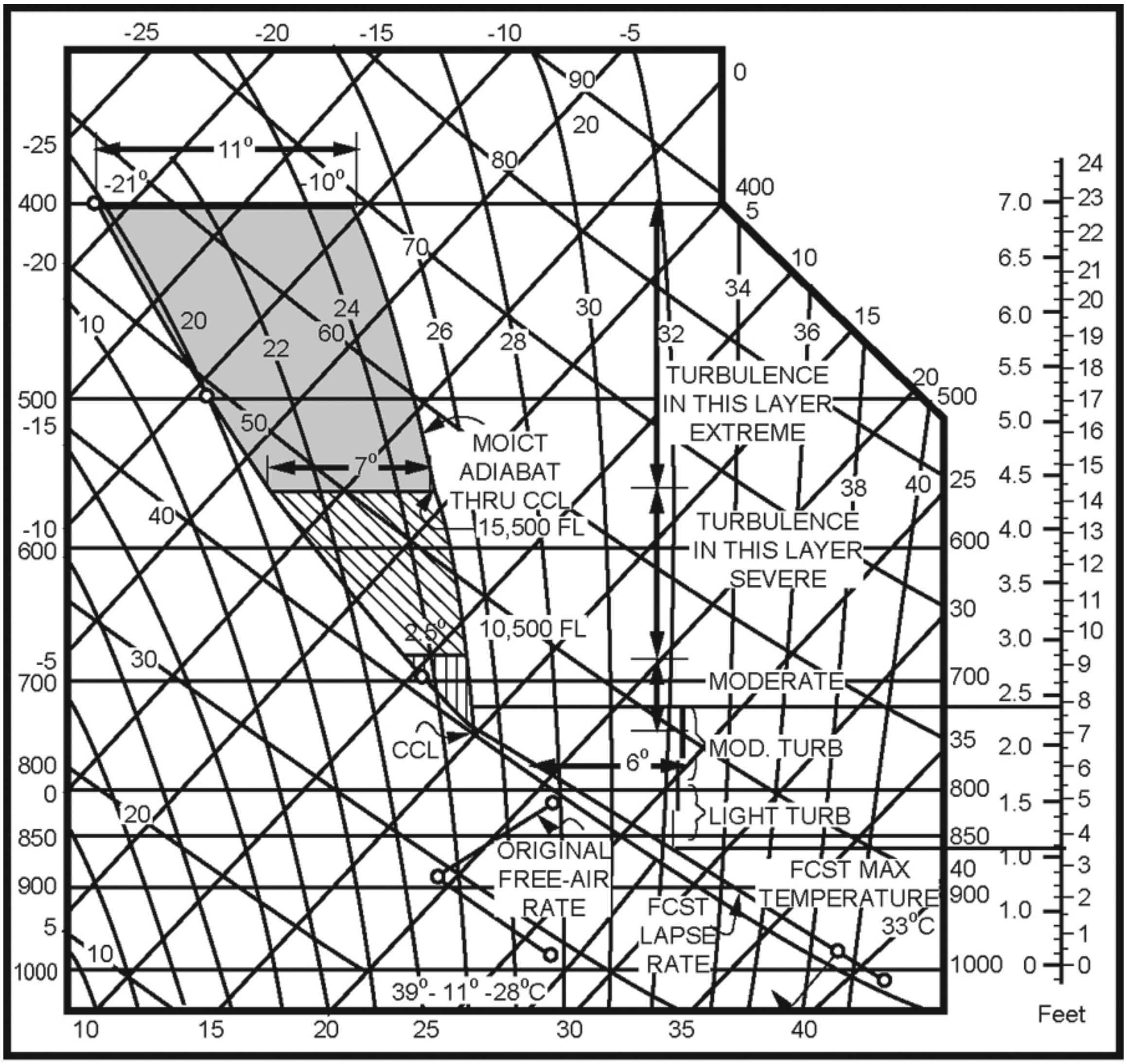

Figure 5–22. Turbulence Skew-T example.

Layers from surface to 9,000 feet

Use the steps below to estimate the buoyant potential in the lower atmosphere. Use the results obtained by this method to estimate turbulence in thunderstorms. Use the convective temperature to forecast the maximum surface temperature.Project a dry adiabat from the CCL to the surface. This gives the convective temperature. Adjust this temperature using temperature curves for local effects. Subtract 11°C from the final forecast maximum temperature. Follow this isotherm to its intersection with the dry adiabat projected upward from the forecast maximum temperature. If the intersection is above 9,000 feet MSL, forecast no turbulence below 9,000 feet MSL. If the intersection is below 9,000 feet, draw a moist adiabat from the intersection of the isotherm and the dry adiabat upward to the 9,000 feet level. The temperature difference between this moist adiabat and the free-air temperature curve determines the severity of the turbulence as well as the limits of the layers of each degree of turbulence. Apply the temperature differences to the layers “below 9,000 feet table” below.

Layer temperature difference	Turbulence forecast
0 to 6 degrees C	Light
6 to 11 degrees C	Moderate
11 degrees C or more	Severe

Layers above 9,000 feet

Follow the moist adiabat that passes through the CCL upward to the 400mb level. The maximum temperature difference between this moist adiabat and the forecast free-air temperature curve is the central portion of the most turbulent area. The intensity of the turbulence is found in the “above 9,000 feet table” below.

Layers where temperature difference is	Forcast turbulence as
0 to 2.5 degrees C	Moderate
2.5 to 7 degrees C	Severe
7 degrees C or more	Extreme

279. Forecast sounding

The development of a forecast sounding is simply a matter of forecasting local temperature, dew point, and winds for specific mandatory pressure levels at a predetermined time. We can use this technique anywhere in the world; it is especially useful where radiosonde observation (RAOB) file times do not match up with local heating or changes in air mass.

Although this method is explained as though you were constructing a manually plotted Skew-T, you will more than likely use the programs from the OWS or a Skew-T computer program to construct a forecast sounding. For example, you can quickly modify the Skew-T and create a forecast sounding after you edit the raw data and generate new parameters from the data. The stability indices are automatically and almost instantaneously calculated. However, regardless of how the data is processed, the final result is still the same—a sounding that you can use to determine severe weather potential.

In severe weather forecasting, forecasting the sounding for maximum heating is the best time period to apply this technique. You can extract the data, for the most part, from the standard 1200Z surface and upper-air products. Following are the levels you use and the elements of each level that you need to forecast.

Surface

Warm, moist air is favorable for air mass destabilization. Most units routinely forecast and post a daily maximum temperature. Plot this forecast maximum temperature on your forecast sounding. To forecast a surface dewpoint temperature, next analyze a regionalized surface chart with isodrosotherms (lines of equal dewpoint temperatures). Look for the moisture patterns upstream by noting the low-level winds and advect accordingly. Forecast the local dewpoint temperature for the maximum heating time on your sounding.

850 millibar temperature

Warm, moist air at this level is also favorable for destabilization. It’s important that you understand that there will be times when the moist air is below the 850-mb level. Using the 12Z 850mb product, analyze for temperature and dew points (not depressions) upstream. Forecast the temperature based on the advection flow.

Estimating the 850 millibar dewpoint temperature

As we already discussed, a key input to making a forecast sounding is low-level moisture. Low-level moisture information is limited to the 850mb and surface dewpoint temperatures. We can measure the available moisture at either level by using a Skew-T. A dewpoint temperature and corresponding saturation mixing ratio value provides us with this information. Which level, 850mb or surface, do we use to forecast the 850mb dewpoint temperature for the forecast sounding?

During the course of the day, moisture is advected horizontally from upstream areas. We can easily horizontally advect a dew point at 850mb and the surface. On the other hand, the high moisture content trapped near the surface is gradually lifted vertically due to turbulent mixing (convection, winds, and terrain). Somehow, we must consider both the horizontal and vertical moisture contributions when we forecast the 850mb forecast sounding dewpoint temperature.

Calculating 850mb horizontal moisture contribution

Follow these steps to forecast the 850mb horizontal moisture content:

- Forecast an 850mb dew point based on the advection of upstream air.
- Using a Skew-T Log P diagram, find and record the saturation mixing ratio that is equivalent to the dewpoint temperature found in the previous step.

Calculating 850mb vertical moisture contribution

Using a Skew-T Log P diagram, find and record the saturation mixing ratio that is equivalent to the surface dewpoint temperature that you previously calculated.

The forecast 850mb dew point

Average the horizontal and vertical values obtained above. Using the forecast sounding Skew-T, plot this average saturation mixing ratio on the 850mb level. The temperature at this point is the 850mb dewpoint temperature for the forecast sounding.

However, when the 850mb level is extremely dry (greater than or equal to 20°C dewpoint depressions), use *only* the vertical moisture contribution. This accounts for a shallow moisture layer that does not reach the 850mb level. Therefore, in this situation, use only the saturated mixing ratio value of the surface dewpoint temperature that intersects the 850mb (read in °C). This will be the forecasted dewpoint temperature for the 850mb level.

700 millibar

Dry-air advection is favorable for air mass destabilization. Forecast a temperature and a dew point for the maximum heating timeframe based on advection rate indicated by your analysis of data upstream (200–250 nm).

500 millibar

Cold-air advection is most favorable for air mass destabilization (however, weak warming when the low levels have undergone significant warming may still result in explosive convection or destabilization). Look upstream (250–400 NM) and allow for deepening of troughs (cold-air advection). Forecast a temperature for the maximum heating time frame. Forecast a dewpoint temperature only when you need to construct a wet bulb curve for computing the WBZ height.

Forecast sounding winds

A 40 to 90 degree veering wind shear between the 850mb (or low-level jet) and 500mb favors destabilization.

850 millibar winds

Plot the direction and speed of the low-level maximum wind upstream.

500 millibar winds

The direction of the forecast 500mb wind should reflect any changes in contour amplitude, which can result due to significant deepening of a trough and/or building of a ridge. Changes to the upstream amplitude of the contours result in the same changes to the wind direction. Use an initialized numerical package as a "first guess" wind direction for the forecast sounding. Find the speed of the 500mb wind by using the maximum wind speed within 250 NM upstream based on the 12Z analysis data.

Use of the forecast sounding

Now that the forecast sounding construction is complete, you can calculate severe weather indices to determine the severe weather potential for your area. As usual, the forecast sounding is a tool you do *not* use as a stand-alone product. Use it to help you in determining whether you need to issue an advisory or warning once thunderstorms develop or are forecasted to develop.

During forecast discussions, the forecast sounding provides a sound meteorological platform for severe weather dialogue. The forecast sounding technique works if you use it properly.

Self-Test Questions

After you complete these questions, you may check your answers at the end of the unit.

277. Frontal zone characteristics

1. What is the better indicator of a frontal zone, the temperature or dewpoint curve?

2. What are the best sources for identifying frontal inversions?

3. Do wind speeds usually increase or decrease with a wind shift?

278. Forecasting turbulence in convective clouds

1. What determines the severity when forecasting turbulence in convective clouds?

2. From the surface to 9,000 feet, what do we use to forecast the maximum surface temperature?

279. Forecast sounding

1. When is the best time to apply forecast sounding?

2. In the low levels what parameter is most favorable for destabilization?

3. What parameter is most conducive to destabilization at 700mb?

4. Is cold air advection or weak warm air advection favorable for destabilization at 500mb?

Answers to Self-Test Questions

260

1. (1) d.
   (2) e.
   (3) c.
   (4) a.
2. Backing or veering of the winds with height.
3. Straight equally spaced lines that slope from the lower left to the upper right of the diagram.
4. Dry adiabits are slightly curved lines that slope from the lower right to the left of the chart. They indicate the rate of temperature change in parcel of dry air rising or descending adiabatically. Moist adiabits are slightly curved lines that slope from the lower right to upper left of the chart, but only extend to the 200mb level. Moist adiabats represent the temperature change undergone by a saturated parcel of air as it rises or descends through the atmosphere.
5. 4010m.

261

1. –11.67C.
2. The temperature an air sample would have if its pressure were increased to 1,000mb in a dry adiabatic process.
3. It is the lowest temperature to which a parcel of air can be cooled by evaporating water into it at a constant pressure.
4. It is the wet bulb temperature a parcel of air would have if it were brought moist adiabatically to the 1,000mb level.
5. Equivalent temperature is the temperature a parcel of air would have if it were cooled both dry and moist adiabatically until all of its moisture was condensed from it.
6. It is the temperature a parcel of air would assume if, having reached its equivalent temperature, it was warmed by adiabatic compression to 1,000mb.
7. The temperature at which a parcel of dry air would have the same density as a parcel of moist air at a given pressure.

262

1. The level at which a parcel of air, when lifted, will cool dry adiabatically until condensation occurs.
2. The mixing ratio line and the dry adiabat.

263

1. The theoretical level where condensation in an ascending parcel is reached because of convection.
2. How you are going to use it.

264

1. The MCL is the lowest height in a layer, mixed by turbulence, at which saturation occurs after complete mixing of the atmosphere.
2. Forecasting cloud bases in areas of strong mechanical mixing.

265

1. The temperature that the surface air must reach to initiate convective currents that will extend high enough for the air to become saturated.
2. A method to compute the T_C when the moisture content is highly variable in the lower levels near the surface.

266

1 The height at which a parcel of air , which is lifted, first becomes warmer (less dense) than the environment.

2 No LFC exists.

267

1. Negative.
2. Radiational.
3. The dew point associated with a subsidence inversion rapidly decreases at the base of the inversion. The dew point associated with a frontal inversion increases.

268

1. The buoyancy that a parcel of air has as compared to its environment.
2. Hail potential increases dramatically.

269

1. Vertical totals and cross totals.
2. Temperature and moisture.

270

1. Temp and dew point at both 500 and 850mbs plus wind speeds and directional shear.
2. The potential of thunderstorms to produce tornadoes.

271

1. 850mb dew point and 700mb dewpoint depression.
2. Peterson AFB.

272

1. Unstable and buoyant lower levels of the atmosphere.
2. (1) d.
   (2) b.
   (3) c.
   (4) a.

273

1. Equally divided.
2. 40 percent in the clear air, 60 percent in the storm, and 85 percent under the overhang.
3. 80 percent in storm with 20 percent in the clear air or overhang.

274

1. 130/680 = .19 Yes.
2. 205/585 = .35 No.

275

1. Aircraft may encounter hail at any altitude in or near thunderstorms.
2. Determine the CCL.
3. 2 inches.

276

1. The warmest part of the inversion.
2. Between 10,000 and 14,000 feet.

3. Indicates the maximum gusts expected and that the thunderstorm must pass over the forecast point and moderate to heavy rain must occur, except in a Type IV sounding.

277

1. Dewpoint curve.
2. A combination of temperature and dewpoint curves coupled with the vertical wind distribution.
3. Increase.

278

1. The temperature difference between the moist adiabat and the free-air temperature curve.
2. The convective temperature.

279

1. Max heating.
2. Moisture.
3. Dry-air advection.
4. Cold-air advection, occasionally weak warm air advection.

Do the unit review exercises before going to the next unit.

Unit Review Exercises

Note to Student: Consider all choices carefully, select the *best* answer to each question, and *circle* the corresponding letter.

76. (260) Why does the isobar spacing on a Skew-T chart increase from the bottom to the top of the chart?
a. Stronger upper level winds.
b. Warmer air in the lower levels.
c. Accounts for the increase in density of the atmosphere with height.
d. Accounts for the decrease in density of the atmosphere with height.

77. (261) Potential temperature is the temperature an air sample would have if its pressure were increased to
a. 1,000 millibars (mb) in a dry adiabatic process.
b. 1,050mb in a dry adiabatic process.
c. 1,000mb in a moist adiabatic process.
d. 1,050mb in a moist adiabatic process.

78. (262) The lifted condensation level is the intersection obtained on the Skew-T chart between the
a. saturation adiabat and dry adiabat.
b. mixing ratio line and dry adiabat.
c. saturation adiabat and dew point.
d. mixing ratio line and dew point.

79. (263) A second method for finding the convective condensation level was developed to give increased accuracy in
a. air-mass analysis.
b. pressure system analysis.
c. severe weather forecasting.
d. frontal weather forecasting.

80. (264) To determine the mixing condensation level, you must *first* determine the
a. top of the mixing level.
b. bottom of the mixing level.
c. mean mixing ratio for the mixed layer by use of the equal area method.
d. mean potential temperature of the mixed layer by use of the equal area method.

81. (265) Convective temperature is the temperature that surface air must reach to initiate convective currents that will
a. increase atmospheric pressure.
b. extend wide enough for the air to become dry.
c. extend high enough for the air to become dry.
d. extend high enough for the air to become saturated.

82. (265) What temperature can be obtained by following a dry adiabat air parcel down from the convective condensation level to the point where the dry adiabat air parcel intersects the isobar representing the surface pressure?
a. Convective.
b. Condensation.
c. Free convective.
d. Free condensation.

83. (266) The height at which a parcel of air that is lifted, first becomes warmer (less dense) than the environment is the
a. level of free convection.
b. lifted condensation level.
c. adiabatic condensation level.
d. convective condensation level.

84. (267) What indicates a temperature increase with an increase in height indicating a negative lapse rate in a layer of the atmosphere?
a. Condensation.
b. Convection.
c. Inversion.
d. Mixing.

85. (268) What is the potential for thunderstorms with a convective available potential energy (CAPE) of 2250 joule per kilogram (J/kg)?
a. Low.
b. Moderate.
c. High.
d. Extreme.

86. (269) What would be the likely coverage and intensity of thunderstorms with a cross total of 30 and a vertical total of 23?
a. Isolated moderate and few severe.
b. Scattered severe.
c. Scattered moderate and few severe.
d. Tornadic.

87. (270) Tornado occurrences are grouped in severe weather threat index values of
a. 150 to 300.
b. 300 to 400.
c. greater than 400.
d. greater than 600.

88. (271) The K-Index is a measure of a thunderstorm’s
a. intensity.
b. rain fall.
c. potential.
d. hail production.

89. (272) What index is used to assess the instability of the 850 millibar (mb) parcel?
a. K.
b. Convection.
c. Showalter stability.
d. Free-lift convection.

90. (273) The season of *maximum* hail occurrences coincides with
a. summer.
b. spring.
c. fall.
d. the season of maximum thunderstorms.

91. (273) What percentage of hail occurrences in flight is attributed to have happened under overhanging clouds between 10,000 and 20,000 feet?
a. 50.
b. 75.
c. 85.
d. 90.

92. (274) What is the cloud depth ratio if the convective condensation level is at 770 millibars (mb), the freezing level at 550mb, and the equilibrium level at 270mb?
a. 0.24.
b. 0.34.
c. 0.37.
d. 0.44.

93. (275) What is considered the *optimum* wet bulb zero (WBZ) height, in feet, above the terrain for hail?
a. 12,000.
b. 10,500.
c. 8,000.
d. 5,000.

94. (275) What *must* be determined as the *first* step in forecasting the size of hail?
a. Wet-bulb zero.
b. Level of free convection.
c. Convective condensation level.
d. Ratio of cloud depth below the freezing level.

95. (276) In using the T_1 method of forecasting maximum wind gusts, the warmest point on the inversion is used to determine the T_1 if surface convection *does not* dissipate the inversion and if that warmest point is within how many millibars (mb) of the surface?
a. 100.
b. 150 to 200.
c. 200 to 250.
d. 250.

96. (276) What preconditions *must* exist to make the T_2 method over forecast maximum gusts?
a. The difference between the wet bulb zero (WBZ) and the 600 millibar (mb) free-air temperature must exceed 10 degrees (°) Celsius (C).
b. The difference between the WBZ and the 600mb free-air temperature must be less than 10°C.
c. The thunderstorm must pass over the forecast point and moderate to heavy rain must occur.
d. The mean wind direction at the 700mb level must be from the south to southwest direction.

97. (277) A veering wind in the vertical suggests a
a. mesocyclone.
b. thunderstorm formation.
c. cold air advection not associated with a warm front.
d. warm air advection not associated with a warm front.

98. (278) When forecasting turbulence in convective clouds below 9,000 feet what determines the severity of the turbulence?
a. Temperature difference between the moist adiabat and the free-air temperature curve.
b. Updrafts and downdrafts.
c. Dew point.
d. Anvil.

99. (278) When forecasting turbulence in convective clouds above 9,000 feet, where is the central portion of the most turbulent area?
a. Where the maximum temperature difference between the moist adiabat and the free-air temperature is.
b. Where the minimum temperature difference between the moist adiabat and the free-air temperature is.
c. Where the temperature line crosses the mixing ratio line.
d. Where the dew point line crosses the mixing ratio line.

100. (279) In severe weather forecasting, the *best* time period to forecast a Skew-T sounding is during the
a. morning.
b. evening.
c. afternoon.
d. maximum heating.

Student Notes

Glossary of Abbreviations, Acronyms, and Terms

θ	potential temperature
θ_e	equivalent-potential temperature
θ_w	wet-bulb potential temperature
AFMAN	Air Force Manual
AFOS	automation of field operations and services
AFWA	Air Force Weather Agency
AGL	above ground level
AIRMET	Airmen's Meteorological Information
APP	Code used in airways on three and six-hourly observations to include A, which indicates a characteristic of barometer tendency, and PP, which indicates the amount of barometric change.
AWC	Aviation Weather Center
AWSTR	air weather service technical reference
BLC	boundary layer convergence
BRN	bulk Richardson number
BWER	bounded weak echo region
C	celsius
CAA	cold-air advection
CAPE	convective available potential energy
CAT	clear air turbulence
CCL	convective condensation level
$\mathbf{CCL_{ml}}$	convective condensation level moist layer
$\mathbf{CCL_p}$	convective condensation level parcel
CCN	cloud condensation nuclei
CDC	career development course
CI	cirrus
CONUS	Continental United States
cP	continental polar air mass
cPk	continental polar, cold air mass
CS	cirrostratus
cT	continental tropical
CT	cross totals
dBZ	decibels
e	vapor pressure
EL	equilibrium level
ERL	early run forecast model
$\mathbf{e_s}$	saturation vapor pressure
ET	echo tops
FFD	forward-flank downdraft
FNL	final run forecast model
FOD	foreign object damage

FOUS	forecast output United States
GADB	global applications data base
fps	feet per second
g/kg	grams per kilogram
GFS	Global Forecast System
GOES	geostationary operational environmental satellite
GSM	global spectral model
GTP	Graphical Turbulence Product
GTWAPS	Global Theater Weather Analysis and Prediction System
Hg	mercury
ICAO	International Civil Aeronautical Orginization
IN	ice nuclei
IP	ice pellets
IR	infrared
ISMCS	International Station Meteorological Climate Study
JAAWIN	Joint Air Force and Army Weather Information Network
JTWC	Joint Typhoon Warning Center
KI	K index
km	kilometer
Kts	knots
LAFP	local Analysis and Forecast Program
LC	low cloud
LCL	lifted condensation level
LEWP	line echo wave pattern
LFC	level of free convection
LFM	limited fine mesh
LI	lifted index
LLJ	low-level jet
LLWS	low-level wind shear
mb	millibar
MCC	mesoscale convective complex
MCL	mixing condensation level
MESO	mesocyclone
METSAT	meteorological satellite
METWATCH	meteorological watch
MM5	mesoscale model 5th version
MODCURVES	modeled curves
MOS	modeled output statistics
mP	maritime polar air mass
MRF	medium range run forecast model
MSL	mean sea level
mT	maritime tropical air mass
mTw	maritime, tropical, warm air mass

MWA	military weather advisories
NAM	North American Model
NASA	National Aeronautics and Space Administration
NCEP	National Center for Environmental Prediction
NEXRAD	next generation weather radar
NGM	nested grid model
NHRL	National Hurricane Research Laboratory
nm	nautical miles
NMC	National Meteorological Center
NOAA	National Oceanic and Atmospheric Administration
NOGAPS	Navy Operational Global Atmospheric Prediction System
NSSFC	National Severe Storms Forecast Center
N-TFS	New Tactical Forecast System
NTSB	National Traffic Saftey Board
NVA	negative vorticity advection
NWA	National Weather Agency
NWF	northwest flow
NWP	numerical weather prediction
NWS	National Weather Service
OCDS	Operational Climatic Data Summary
OI	optimum interpolation
OJT	on the job training
OWS	operational weather squadron
PA	pressure altitude
PBL	planetary boundry layer
PEA	positive energy area
PFJ	polar-front jet
PGF	pressure gradient force
PIBAL	pilot balloon
PIREP	pilot report
PST	polar stereographic grid
PUP	principal user processor
PVA	positive vorticity advection
R	reflectivity
RAFP	regional area forecast program
RAOB	radiosonde observation
RCS	reflectivity cross section
RDA	radar data acquisition
RFD	rear-flank downdraft
RGL	regional run forecast model
RH	relative humidity
ROT	rules of thumb
RPG	radar product generator

RWM	relocatable window model
SAFWIN	Secure Air Force Weather Information Network
SBLI	surface based lifted indexes
SHARP	Skew-T, Hodograph Analysis and Research Program
SIGMET	significant meteorological information
SLP	sea-level pressure
SLW	supercooled liquid water
SOCS	surface observation climatic summaries
SPC	Storm Prediction Center
SSI	showalter stability index
SST	Sea surface temperature
SW	spectrum width
SWEAT	severe weather threat index
SWP	severe weather probability
T	temperature
TAF	terminal aerodrome forecast
TBM	theater battlefield management
T_c	convective temperature
T_{cml}	convective moist layer temperature
T_d	dew-point temperature
Td	dew point
T_e	equivalent temperature
TPC	Tropical Prediction Center
TT	total totals
T_v	virtual temperature
T_w	wet-bulb temperature
UTC	coordinated universal time
VAD	vertical azimuth display
VIL	vertically integrated liquid
VIV	verification/initialization/verification
VT	vertical totals
VWP	velocity azimuth display wind profile
VWP	Velocity Azimuth Display Wind Profile
w	mixing ratio
WAA	warm-air advection
WBZ	wet-bulb zero
WER	weak echo region
WMOTWC	World Meteorological Organization Typhoon Warning Central
WRF	weather research and forecast
w_s	saturation mixing ratio
WSCC	wind stratified conditional climatology
ZR	freezing rain
WSR–88D	weather surveillance radar–1988 Doppler

Terms

adiabatic process—A thermodynamic change of state of a system in which there is no transfer of heat or mass across the boundaries of the system.

ambient—Surrounding, encircling.

black stratus—The net effect of a warm earth in areas covered by air with high moisture content appears darker on IR imagery. These boundaries can be monitored and used as an estimate of the extent of nocturnal fog and stratus development and advection.

anticyclogenesis—The intensification or development of anticyclonic circulation.

baroclinic—When contours and isotherms are out-of-phase, advection is occurring and the atmosphere is described as baroclinic. In a baroclinic system, the axis tilts with height.

baroclinic instability—A process by which short waves amplify (increase amplitude with time) by extracting energy from the north/south temperature gradient.

barotropic—When contours are in-phase, the atmosphere is said to be equivalent barotropic - or simply barotropic. There is no temperature advection.

convergence—A measure of the rate of the net addition of mass into a volume of air above a given point (usually the LND). Promotes surface pressure and low-level height rises.

cyclogenesis—The intensification or development of cyclonic circulation.

deepening—A decrease in the central pressure/height of a low or trough.

dewpoint depression—The difference in degrees between the air temperature and the dew point.

diabatic process—Thermodynamic change of state of a system in which there is transfer of heat across the boundaries of the system.

diametrically—Of or along a diameter.

decibels—A measure of the relative power densities.

difluence/diffluent—Flowing apart resulting in mass being removed from an area.

divergence—A measure of the rate of net removal of mass out of a volume of air above a given point (usually the LND). Promotes surface pressure and low-level height falls.

dryline—A severe weather producing system that is located over western TX, OK, and southern KS.

derecho—A rapidly moving extratropical convective system known to produce widespread significant straight-line wind damage.

dynamics—Atmospheric motions.

echo—In radar, a general term for the appearance, on a radar indicator, of the radio energy returned from a target.

eccentricity formula—Formula used to forecast the movement of closed lows at the 500-mb level.

extrapolation—To estimate something unknown based on known facts.

funnel cloud—A tornado that does not touch the ground.

filling—An increase in the central pressure/height of a low or trough.

geostrophic wind—The wind that would result if there were a balance between the coriolis force and the pressure gradient force over a region.

hodograph—Displays vertical wind profiles as vectors.

helicity—A measure of a component's vertical shear in a storm's relative inflow that is parallel to the mean wind. A term for horizontal vorticity.

katallobaric—Of, or pertaining to, a decrease in atmospheric pressure.

kinematic—Description of the motion of bodies or fluids without reference to the forces producing the motion. In meteorology, the analysis of the motion of isobars, contours, and fronts when treated as geometric features of the pressure or height fields.

kinetic energy—The energy which a body possesses as a consequence of its motion.

leeside—The side apart or away from the wind.

lapse rate—The decrease of an atmospheric variable with height, the variable being temperature, unless otherwise specified.

low-level wind shear—The low-level local variation in the wind vector or any of its components in a given direction. Usually occurs within 2,000 feet above ground level (AGL).

line echo wave—A line of radar echoes subjected to acceleration along one portion of the line.

mesocyclone—A cyclone or low pressure area smaller than the synoptic scale but larger than the microscale.

mercator grid—A map grid that shows the earth's surface as a rectangle. The meridians are parallel straight lines spaced at equal intervals and the parallels of latitude are depicted as parallel straight lines intersecting the meridians at right angles but spaced further apart as their distance from the equator increases.

occluded—A composite of two fronts, formed as a cold front overtakes a warm front.

orographic lifting—Lifting due to mountainous or "hilly" terrain.

polar stereographic grid—Fixed spacing and no variable grid spacing and are centered on the hemisphere (north or south) to include the entire hemisphere.

potential energy—The energy which a body possesses as a consequence of its position in the field of gravity.

pseudo front—A small-scale front, formed in association with organized severe convective activity, between a mass of rain-cooled air from the thunderstorm clouds and the warm surrounding air.

progging—A common term used interchangeably with the word forecasting.

radiation inversion—A surface-based inversion formed by rapid cooling of air in contact with the surface of the earth. Occurs in times of maximum radiational cooling, normally just before and after sunrise. Frequently associated with fog.

resultant—The direction of a parcel's movement found by the vector sum of all the forces acting on the parcel.

retrogression—A pattern that moves backward and only occurs with the long-wave pattern.

shear zone—A zone across which there is an abrupt change in the horizontal wind component parallel to the zone.

subgradient winds—A wind which is adjusting to a rapid increase in contour (or pressure) gradient force downstream.

subsidence inversion—A mechanically produced inversion formed by adiabatic warming of sinking air. Forms in areas of high pressure. The dew point will rapidly decrease at the base of the inversion as the sinking air warms and dries out. These types of inversions suppress convective activity.

supergradient winds—A wind which is adjusting to a rapid decrease in contour (or pressure) gradient force downstream.

speed shear—The variation of wind speed of a vector field along a given direction in space.

synoptic scale—The scale of the migratory high- and low-pressure systems of the lower troposphere, with wave lengths of 1,000 to 2,500km.

temperature advection—The process of transporting temperatures solely by the mass motion of the atmosphere.

terminal velocity—The particular falling speed, for any given object moving through a fluid medium of specified physical properties, at which the drag forces and the buoyant forces exerted by the fluid on the object just equal the gravitational force acting on the object.

thickness—A type of synoptic product showing the thickness of certain physically defined layers in the atmosphere (i.e., 1000 to 500-mb thickness product).

unstable wave—A baroclinic low in which the amplitude of the wave on a frontal system increases (deepens) with time.

vector—Any quantity, such as force, velocity, or acceleration, which has both magnitude and direction at each point in space, as opposed to scalar which has magnitude only.

viscous fluid—A fluid whose molecular viscosity is sufficiently large to make the viscous forces a significant part of the total force field in the fluid.

vorticity—A measure of local rotation or “spin” in a fluid flow.

water spout—A tornado over a body of water.

weakening—A decrease in the central pressure/height of a high or ridge.

Student Notes

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