CDC 1W051A

Weather Journeyman

Volume 1. General Meteorology and Surface Weather Observations

Extension Course Program (A4L)

Air University

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CONGRATULATIONS on your completion of the formal, 3-level, initial skills course training to become a weather journeyman. Your journey toward becoming a skilled 5-level journeyman is just beginning. If you are in the 1W0X1 career field it's likely you are currently in an operational weather squadron (OWS). Assigned to an OWS, you have probably already begun the first phase of your upgrade training, which is called on-the-job training (OJT). Air Force Weather has developed a top-notch field-training program utilizing this course, qualification training packages (QTP), and OJT. You started the first phase of your training when you were assigned to work with a trainer. Your trainer will work with you to refine the skills you need to perform your job. Your trainer will also provide training with QTPs, which are designed to assist in the learning of core tasks associated with the career field. The second phase of the training consists of self-study to gain career knowledge. This self-study is provided by specially designed career development courses (CDC). Using these carefully planned and prepared texts, you will acquire the necessary career knowledge with some assistance from you trainer. If you are in the 1W0X2 career field this is the first phase of your 5 level upgrade training. Upon completion of the 1W0X1 CDC's you will be enrolled in the 1W0X2 CDC's.

CDC 1W051A offers you additional theory and information necessary to compliment the OJT process leading to eventual upgrade to AFSC 1W051A. It enlarges and expands the training you received in the 1W031 resident course. The 1W051 CDC is broken down into two courses. The first course contains three volumes: volume 1, *General Meteorology and Surface Weather Observations*; volume 2, *Weather Radar*, and volume 3, *Basic Metwatch and Space Environment Concepts*.

Volume 1 has four units. The first unit deals with the structure and properties of the atmosphere. You will read material on atmospheric composition and circulation, jet streams, and air masses. The second unit covers synoptic weather systems. In it, you will learn about low and high pressure systems, cyclones and anticyclones, and fronts. The third unit begins information on surface weather observations. Specifically, you will learn about the 27 different states of the sky and how the sky condition is encoded in an observation. The fourth unit covers other weather observation elements.

A glossary of terms, abbreviation, and acronyms is included for your use.

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This volume is valued at 24 hours and 8 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.

Page

Unit 1. Understanding the Atmosphere

OUR ATMOSPHERE has an enormous impact on our lives and, together with the sun, sustains life on our planet. The origin of the word *atmosphere* is Greek, from *atmos* meaning vapor and *sphaira* meaning sphere. It’s literally a sphere of vapor, or a gaseous envelope, surrounding a sphere or globe. The atmosphere of the earth is an envelope surrounding the earth. Space exploration has shown that other planets have atmospheres but none have an atmosphere like our earth.

To understand atmospheric dynamics and weather you have to understand the atmosphere and its structure and properties. Concerning understanding, you might well remember these words of advice from an ancient philosopher: “Wisdom is the principal thing; therefore get wisdom, and with all your getting of wisdom—get understanding.” This implies that our understanding of a subject depends on our knowledge of the subject at our command. This unit aims toward increasing your understanding of the earth’s atmosphere.

In our first section, we cover the structure and composition of the atmosphere. In our second section, we discuss the general and tertiary circulations we commonly call *wind*. We also look at the forces that cause them. In our third section, we discuss the structure and properties of jet streams. Finally, we will discuss air masses and understanding the process of their modification.

1–1. Atmospheric Structure and Composition

From the gases that make up the air we breathe to why the earth is heated unevenly, you need to know basic facts about the atmosphere before you can understand why the skies are sunny or cloudy or why it’s raining or the wind is blowing. In this section, we’ll cover these three topic areas:

- The divisions of the atmosphere.
- The different characteristics of atmospheric gases.
- The differential heating of the atmosphere.

001. Divisions of the atmosphere

We classify our atmosphere into seven layers and zones. In ascending order, the names of these layers are as follows:

1. Troposphere.
2. Tropopause.
3. Stratosphere.
4. Stratopause.
5. Mesosphere.
6. Mesopause.
7. Thermosphere.

Figure 1–1 illustrates the relative thickness, height, and temperature of these layers.

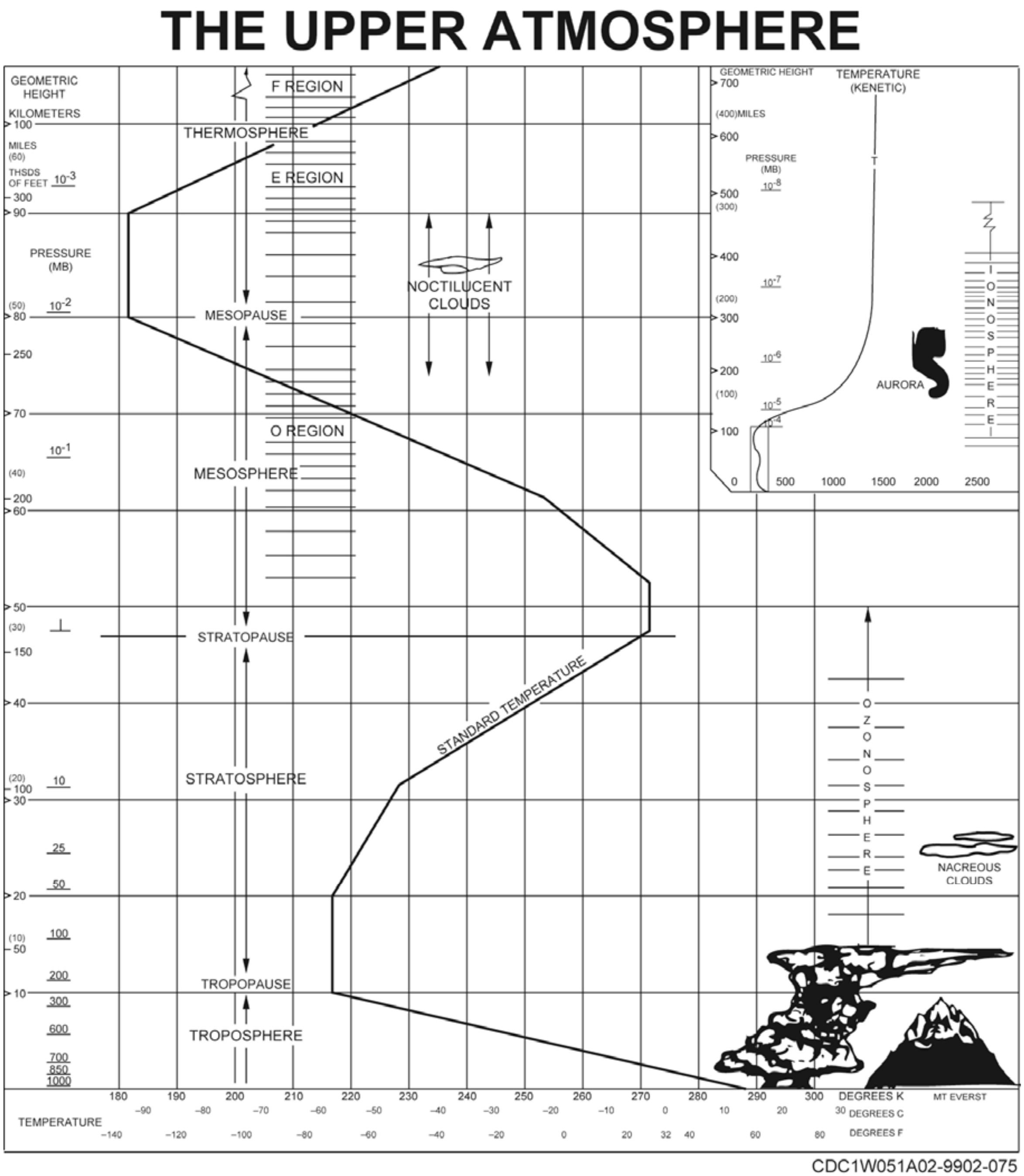

Figure 1–1. Vertical structure of the earth's atmosphere.

Troposphere

The troposphere, whose thickness varies with location and season, has an average height of 38,000 feet (7 miles or 11 km). Within the troposphere, the temperature decreases with height to a range of about –55˚ Centigrade (C) to –60˚C. We usually assume a standard temperature decrease in the troposphere of about 6.5˚C per 1,000 meters of altitude. Because the troposphere contains the most significant vertical wind motions (unstable) and water vapor and experiences all the weather affecting the earth’s surface, it captures most of the attention of forecasters.

Tropopause

The tropopause is a thin zone of transition between the troposphere and the stratosphere. Obviously, the tropopause height varies directly with the troposphere thickness and depends largely on geographical location, season, and other factors. For example, over the poles the average tropopause height is only 5 miles, but over the equator it’s 10 miles. Below the tropopause, the temperature normally decreases as height increases. At the tropopause the temperature remains constant with height (isothermal). The tropopause extends only a few thousand feet but acts as the upper limit for the occurrence of weather.

Stratosphere

The next layer is the stratosphere which extends from the tropopause to a height of about 154,000 feet (31 miles or 48 km). Temperature within the stratosphere remains isothermal up to 100,000 feet; there it begins to warm with altitude. Temperature reaches a maximum of 7°C or 45° Fahrenheit (F) at the top of the layer.

The stratosphere contains most of the ozone of our atmosphere; it’s concentrated between 65,000 and 100,000 feet. The *nacreous* cloud (mother-of-pearl) also occurs in the stratosphere. Scientists believe there may be an association between the ozone concentration and the nacreous cloud. Because no weather occurs in the stratosphere, flying conditions are excellent.

Stratopause

At the height where the stratospheric temperatures stop increasing, there’s a thin zone called the stratopause. This zone represents a transition between the increasing temperatures of the stratosphere and the decreasing temperatures of the mesosphere.

Mesosphere

The mesosphere extends to a height of about 262,000 feet (50 miles). In contrast to the warm upper stratosphere, temperature decreases with altitude in this layer to a minimum near –100°C (–148°F). This rapid temperature decrease causes large convective (vertical) currents. As a result, we consider the mesosphere unstable.

A cloud that occurs in the mesosphere is called the *noctilucent* cloud (meaning “luminous night clouds”) because it appears to glow at night. It’s believed that this cloud is composed of meteoric or cosmic and volcanic dust.

Also found within the mesosphere is a concentration of electrons called the “D” layer, which occurs at or about 230,000 feet. This layer reflects radio waves, thus making long distance communications possible.

Mesopause

The last transition zone is the mesopause. This zone marks the end of the temperature decrease within the mesosphere and occurs at or about 262,000 feet. Above this zone extends the atmosphere’s final layer, the thermosphere, which lacks a distinct upper boundary.

Thermosphere

Within the thermosphere, the temperature remains isothermal to 300,000 feet and then begins to increase with altitude. This layer contains noctilucent clouds and two layers that reflect radio waves.

These layers, like the "D" layer, are concentrations of electrons. They're called the "E" and "F" layers.

002. Characteristics of atmospheric gases

An average sample of pure dry air taken from the atmosphere contains (by volume) 78 percent nitrogen, 21 percent oxygen, almost 1 percent argon, and 0.03 percent carbon dioxide. Two other important gases might also be present in variable amounts: ozone (trace amounts) and water vapor (1.2 percent).

Nitrogen

This is a colorless, tasteless, odorless, gaseous element. It enters the atmosphere from volcanoes and from the decay of organic matter. Plant life removes nitrogen from the atmosphere.

Oxygen

Oxygen is also colorless, tasteless, and odorless. It's a prerequisite of most forms of animal and plant life. At altitudes above about 13 miles (20 km), the radiation from the sun breaks the oxygen down into ozone.

Argon

Just like nitrogen and oxygen, argon is colorless and odorless. It's considered an inactive gas.

Variable gases

Carbon dioxide, ozone, and water vapor are the more variable gases that significantly affect the weather in the atmosphere.

Carbon dioxide

Carbon dioxide results from the decay of vegetation, combustion, volcanic action, and similar processes. Usually, its maximum concentration occurs around cities and industrial areas. However, the actual amount of carbon dioxide varies with the seasons. There's more combustion and, therefore, more carbon dioxide produced during winter heating.

Ozone

Ozone reaches a maximum in the stratosphere. It's important because of its ability to "absorb ultraviolet radiation." Without this protection, the sun would burn everything on earth.

Water vapor

The third of the variable gases, water vapor, is the *most* important in determining weather. We couldn't have the type of weather we're familiar with without water vapor. Water vapor in a parcel of air makes the air lighter than a parcel of dry air. This affects the stability of the atmosphere. The warmer the air, the more water vapor it can hold. The maximum amount of water vapor the air can hold is 4 percent by volume.

003. Differential heating of the atmosphere

Earlier, you learned that water vapor is the *most* important gas in weather formation. Weather also depends on another important variable—heat. In fact, how the earth and its atmosphere are heated is the catalyst for weather formation.

In general, the earth and its atmosphere maintain a constant average temperature. This is due to a balance between the amount of radiated energy received from the sun and the amount of radiated energy the earth loses to outer space. This balance of radiation is maintained by a complex system of reradiation of thermal energy between the earth and water vapor in the atmosphere.

While it's true that over long periods the earth's temperature is somewhat constant, temperature variations occur because of differential heating of the earth and its atmosphere. On a global scale, differential heating results from latitudinal variation in the amount of heating of the earth and its atmosphere. On a smaller scale, other causes of differential heating exist, such as land versus water, hills versus valleys, and so forth. Differential heating is the true cause of the weather you observe. In this lesson, we cover two major subject areas. They are as follows:

1. Factors that influence earth's heating and atmosphere.
2. Barriers to short-wave radiation.

Factors that influence earth's heating and atmosphere

The sun is the source of heat for the earth. To understand how differential heating occurs, you need to be familiar with factors that influence how the sun's energy heats the earth and its atmosphere. These factors include the following:

- Radiation processes.
- Absorption.
- Angle of incidence of solar radiation.

Radiation processes

Radiation from the sun is responsible for heating the earth's surface. Yet, direct solar radiation accounts for only a small part of heat the atmosphere absorbs. Most radiative heating of the atmosphere is the result of reradiated energy from the earth's surface. Absorption of radiated energy depends partly on the wavelength of radiated energy. The wavelength of radiated energy depends on the temperature of the emitting body—the hotter the body, the shorter the wavelength of emitted energy.

Solar radiation has the full spectrum of electromagnetic energy, including gamma, ultraviolet, visible, infrared, and microwave radiation. Still, as the hottest body in the solar system, the bulk of the sun's energy is radiated in short visible wavelengths (less than 3 microns). As a significantly cooler body, the earth's radiation is limited to longer infrared wavelengths between 3 and 80 microns.

We can divide the spectrum into two categories—short-wave solar radiation (less than 3 microns) and long-wave terrestrial radiation (3 to 80 microns). Figure 1–2 shows the electromagnetic spectrum of radiated energy.

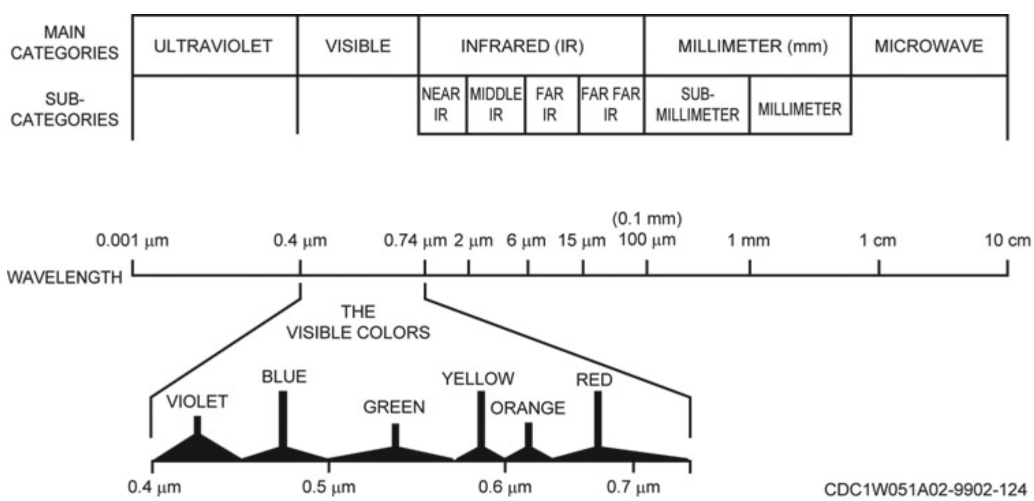

Figure 1–2. Electromagnetic spectrum.

Absorption

Heating of the earth and its atmosphere depends on the ability of the individual masses involved to absorb and emit the two types of radiation. Because the atmosphere is largely transparent to incoming short-wave radiation, a large quantity reaches the earth's surface where it becomes absorbed and heats the earth's surface. The earth then emits long-wave radiation that's readily absorbed by atmospheric gases, such as carbon dioxide and water vapor.

The only source of short-wave radiation is the sun. All other masses emit only long-wave radiation. Therefore, any radiation (long-wave or short-wave) absorbed by water vapor and other atmospheric components becomes reradiated as long-wave radiation.

Angle of incidence of solar radiation

Because of the curvature of the earth's surface, solar radiation strikes at different angles. These are called *angles of incidence*. An angle of incidence is measured from a line perpendicular to the earth's surface. When solar radiation strikes perpendicular to the earth's surface, the angle of incidence is 0°. If it strikes at angles other than perpendicular, the angle of incidence increases as seen in figure 1–3.

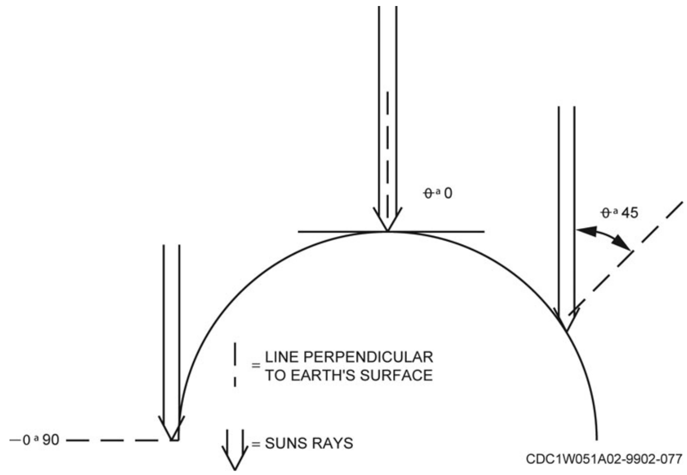

Figure 1–3. Different angles of incidence.

The smaller the angle of incidence, the more the earth's surface is heated. The concentration of solar energy is greatest at lower angles of incidence. Notice in figure 1–4 that when the angle of incidence increases, radiation spreads over a larger area. Because the same amount of energy affects a larger area, the earth's surface is heated less. This effect becomes magnified over the earth's curved surface.

There are two separate motions of the earth that affect angle of incidence:

1. Rotation.
2. Revolution.

Earth's rotation

The earth spins about its axis once each 24 hours. Its rotation is responsible for daily heating and cooling as the angle at which solar rays strike the earth's surface varies from sunrise to sunset. The angle of incidence is greatest at sunrise and sunset—it is least near noon.

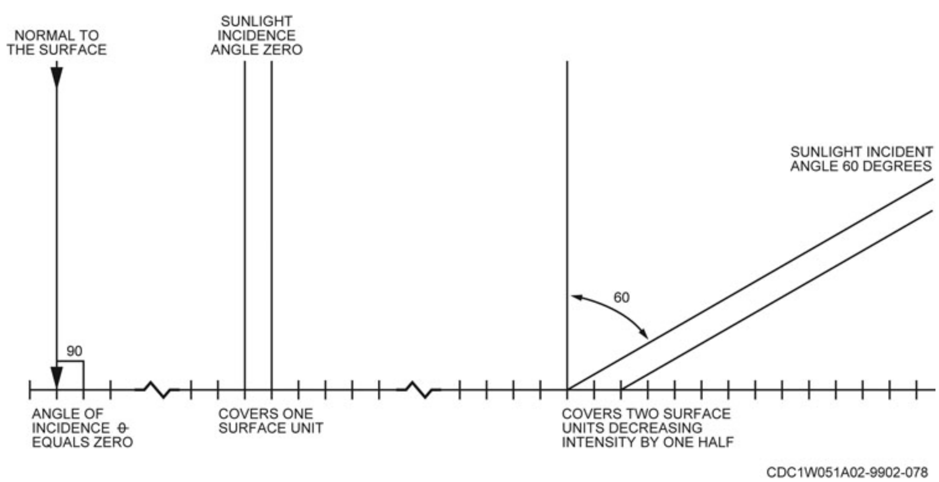

Figure 1–4. Angle of incidence comparisons.

Earth's revolution

Besides experiencing daily variations in heating, the earth experiences variations from season to season. These variations are due to earth's annual revolution around the sun and the tilt of earth's axis. Earth's axis is tilted 23$^{1}/_{2}$° from perpendicular to its orbit (fig. 1–5). The tilt is called the angle of inclination. Because of the inclination, as the earth orbits the sun, the most direct radiation strikes the earth at different latitudes throughout the year. The latitude varies between 23 $^{1}/_{2}$°N (tropic of Cancer) and 23$^{1}/_{2}$°S (tropic of Capricorn).

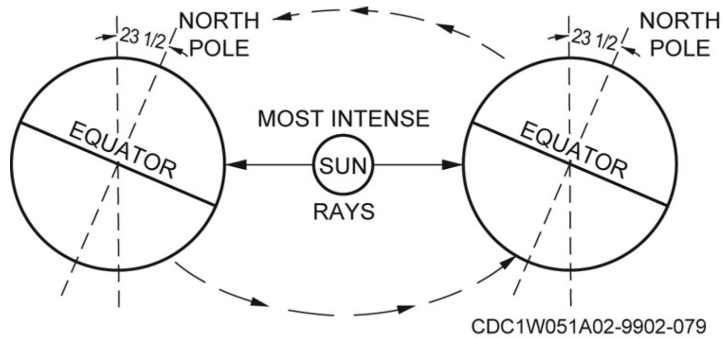

Figure 1–5. Effects of earth's revolution around the sun.

The seasonal variations in latitude of the sun's most direct radiation are marked by these four distinct events:

Seasonal Event	Description
Summer solstice	Occurs on June 21. This is when direct solar radiation is over 23.5 °N.
Winter solstice	Occurs on December 22. This is when direct solar radiation is over 23.5°S.
Autumnal equinox	Occurs when direct solar radiation is over the equator or at 0°. The autumnal equinox occurs on September 23. This is when the direct solar radiation is shifting from the northern latitudes to the southern latitudes.
Vernal equinox	Occurs when direct solar radiation is over the equator or at 0°. The vernal equinox occurs on March 21. This is when the direct solar radiation is shifting from the southern latitudes to the northern latitudes.

NOTE: The dates stated for these events are approximate. They vary from year to year because the calendar year is 365 days long. The solar year is 365¼ days long.

Figure 1–6 shows the earth's position relative to the sun for each of these dates.

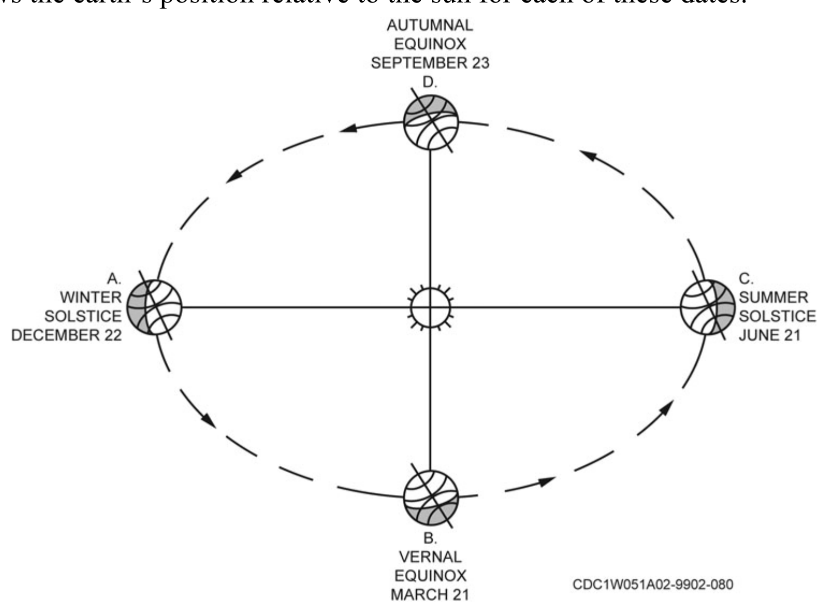

Figure 1–6. Seasonal variations of the earth.

Barriers to short-wave radiation

As we stated earlier, the atmosphere is somewhat transparent to incoming solar radiation; however, because of barriers, not all the incoming radiation passes unimpeded through the atmosphere. Only about 53 percent of incoming solar radiation reaches the earth's surface. In this lesson segment, we'll look at how the atmosphere absorbs and scatters some radiation. You'll also see that some surfaces absorb radiated energy better than others. This ability to absorb energy is measured by a surface's reflectivity or albedo.

Atmospheric absorption

Certain gases absorb some wavelengths of incoming solar radiation very well. Most incoming radiation absorbed by the atmosphere is in ultraviolet wavelengths.

The two gases most responsible for absorption of incoming solar radiation are oxygen and ozone. They are so efficient almost no ultraviolet radiation reaches earth. Atmospheric absorption accounts for a 20 percent reduction of solar radiation reaching earth's surface.

Atmospheric scattering

Scattering is the deflection of incoming solar radiation in all directions. Scattering of radiation depends to a large extent on particle size. For example, atmospheric particles smaller than the wavelength of visible light (0.5 microns) scatter sunlight. The particles scatter at a rate inversely proportional to the fourth power of the light's wavelength. Therefore, shorter wavelengths are much more susceptible to scattering than longer wavelengths.

Atmospheric scattering of solar radiation is apparent in the brightness of the daytime sky. The shorter visible wavelengths of blue and violet light are readily scattered by the particles of the atmosphere. This is why on a clear day the sky is blue. When larger particles are present, such as fog and cloud droplets, all visible wavelengths are scattered equally. This gives the sky the appearance of being white. Consider if the earth lacked an atmosphere, the sky would be black because there would be no scattering of incoming solar radiation. The atmosphere scatters about 5 percent of the incoming short-wave radiation.

Reflectivity (albedo)

Another important variable controlling the amount of solar radiation absorbed by the earth is the reflectivity of the earth's varied surfaces. Albedo, or the reflectivity of a surface, is the ratio of reflected radiation to average total incoming radiation. Albedo varies greatly from surface to surface and can be as high as 90 percent for fresh snow and as low as 15 percent for bare ground. For example, if you're walking barefoot on a sunny day, you're aware of the difference in a bare patch of ground compared with a cooler, grassy surface. The albedo of the bare ground is 7 to 20 percent while the grass surface's albedo is 14 to 37 percent. This shows the importance of higher albedo values yielding cooler natural surfaces.

Albedo also explains why the earth's surface is cooler on an overcast day. Clouds are also reflecting surfaces with an average albedo between 50 and 80 percent, depending on cloud type and thickness. Clouds in the atmosphere account for another 22 percent of solar radiation that doesn't reach earth's surface. The earth's surface reflects another 3 percent of incoming solar radiation.

Self-Test Questions

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

001. Divisions of the atmosphere

1. Match the layer or zone in column B to its appropriate characteristic(s) in column A. Items in column B may be used more than once.

Column A	*Column B*
____ (1) A thin layer of transition between stratosphere and mesosphere.	a. Troposphere.
____ (2) Excellent flying conditions exist because of a lack of weather in this layer.	b. Tropopause.
____ (3) All our weather occurs in this layer/zone.	c. Stratosphere.
____ (4) This layer has an average lapse rate of 6.5°C per 1,000 meters.	d. Stratopause.
____ (5) Contains the “E” and “F” layers.	e. Mesosphere.
____ (6) Noctilucent clouds occur in this layer/zone.	f. Mesopause.
____ (7) Contains most of the ozone of the atmosphere.	g. Thermosphere.
____ (8) Layer of the earth’s atmosphere extending from 262,000 feet to infinity.	
____ (9) The nacreous, or mother-of-pearl cloud, occurs in this layer.	
____ (10) Average height of this layer, over the equator, is 10 miles.	
____ (11) Contains the “D” layer.	
____ (12) An unstable layer of the earth’s atmosphere.	
____ (13) A thin layer of transition between the troposphere and stratosphere.	
____ (14) A thin layer of transition between the mesosphere and thermosphere.	
____ (15) Over the poles, average height of this layer is 5 miles.	
____ (16) Contains layers of electrons that reflect radio waves.	

002. Characteristics of atmospheric gases

1. Match the list of gases in column B with their characteristics in column A. Items in column B may be used more than once.

Column A	*Column B*
____ (1) Occupies 21 percent by volume of the atmosphere.	a. Nitrogen.
____ (2) Occupies 1.2 percent by volume of the atmosphere.	b. Oxygen.
____ (3) Occupies almost 1 percent by volume of the atmosphere.	c. Argon.
____ (4) Occupies 78 percent by volume of the atmosphere.	d. Carbon dioxide.
____ (5) Occupies 0.03 percent by volume of the atmosphere.	e. Ozone.
____ (6) The most important gas to meteorology.	f. Water vapor.
____ (7) The amount of this gas varies with the seasons.	
____ (8) The colder the air, the less of this gas there is in the air.	
____ (9) Maximum quantities are found in the stratosphere.	
____ (10) The more of this gas the air holds, the lighter it will be.	
____ (11) This gas is at a maximum around cities and industrial regions.	
____ (12) Four percent by volume is the maximum amount of this gas the air can hold.	
____ (13) This absorbs ultraviolet radiation.	
____ (14) All weather clouds and precipitation are produced by this gas.	
____ (15) This layer prevents the sun from burning up the earth.	

003. Differential heating of the atmosphere

1. What is the catalyst for the formation of weather?

2. As the angle of incidence increases, what happens to the concentration of solar radiation?

3. Name the two motions of the earth that have an effect on weather.

4. The angle of inclination combined with the revolution of the earth around the sun causes what events?

5. When an equinox occurs, where is an equal amount of sunshine received?

6. What two atmospheric gases are responsible for absorbing incoming solar radiation?

7. Which wavelengths are most susceptible to atmospheric scattering?

8. Approximately how much of the incoming solar radiation is absorbed by the atmosphere?

9. Approximately how much solar radiation is reflected by the earth’s surfaces?

1–2. Atmospheric Circulation

Earlier in this unit, you saw how various factors influence heating of the earth and its atmosphere. For example, angle of incidence is responsible for the wide variation in temperature over different latitudes (differential heating). Because the equator receives more solar energy than the poles, the equatorial region is warmer. However, this isn’t the only factor contributing to horizontal differences. There are also differences in the distribution of land and water masses which profoundly affect the overall temperature distribution. You should know that land heats and cools very rapidly, while water heats slowly but retains its heat longer. Extensive snow cover over polar regions reflects much of the incoming solar radiation, further contributing to latitudinal temperature differences.

In addition to the differences in the distribution of land and water masses, convective heating of gases will occur because of air currents produced from contact or radiative heating. These convective currents (circulation) are due to the nonuniform heating of the atmosphere. The currents are key factors responsible for the atmospheric circulation pattern.

In this section, you'll study other elements which should help you understand general atmospheric circulation. We'll look at these four major topic areas:

1. Thermal and three-cell circulation.
2. Circular motion and the forces involved.
3. Forces affecting earth's general circulation.
4. The earth's circulation systems.

004. Thermal and three-cell circulation

You already know that the equatorial areas receive more direct solar radiation than do the polar areas. Also keep in mind that in the lower latitudes, the ratio of land to water is low. Since water loses heat slowly, more heat is received than is lost through radiation. In the polar regions, just the opposite is true. The land-to-water ratio is high and more heat is lost due to radiation than is received. Since the tropics don't become progressively hotter and polar regions colder, there must be a transfer of heat between the two regions or a general circulation. The mechanism that sets up earth's general circulation pattern is called the latitudinal transfer of heat. We'll first discuss thermal circulation. Then, we'll turn our attention to the characteristics of three-cell circulation.

Thermal circulation

If the earth's surface were smooth, uniform, and stationary, atmospheric circulation would be very simple. The atmosphere would act as a contained fluid and movements within this fluid would be the convective currents caused by temperature and density differences. The latitudinal transfer of heat would result in a single circulation cell.

However, unequal heating is the *main* driving mechanism responsible for the earth's atmospheric circulations. The equator receives more heat than the poles. The warm equatorial air rises and expands until it reaches the tropopause. The tropopause stops the upward motion of this air, causing it to spread and move toward the colder polar regions. Simultaneously, cold and dense air at the poles sinks and starts to flow toward the equator, replacing the warm air rising in this region. This creates the single-cell circulation model possible only on a nonrotating earth. Figure 1–7 illustrates this concept.

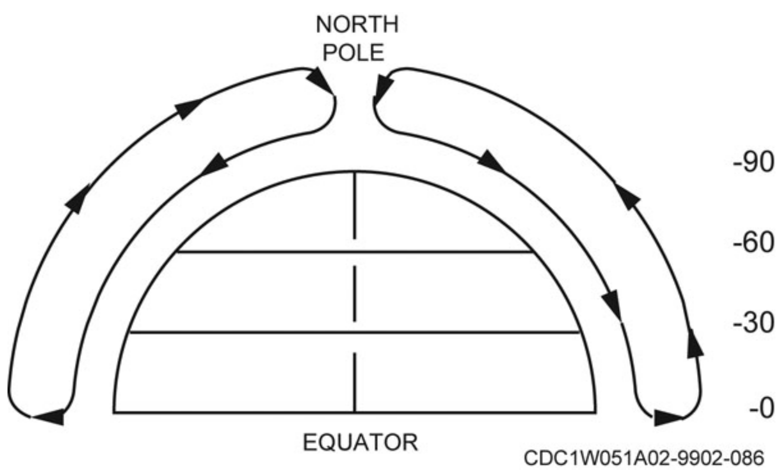

Figure 1–7. Thermal circulation (single cell).

Three-cell circulation

The three-cell circulation theory of circulation best describes the earth's general circulation because it considers the effects of the Coriolis force (CoF) due to the earth's rotation.

NOTE: We'll have an in-depth discussion of the Coriolis force later in the unit.

This circulation model is illustrated on figure 1–8. As you can see, the Northern and Southern Hemisphere are each divided into three cells of circulation, each spanning 30 degrees of latitude. The latitudes that mark the boundaries of these cells are the equator, 30° north and south, and 60° north and south. For our purposes, we consider only the Northern Hemispheric cells shown on figure 1–8. As the drawing illustrates, they are as follows:

- Hadley cell.
- Polar cell.
- Ferrel cell.

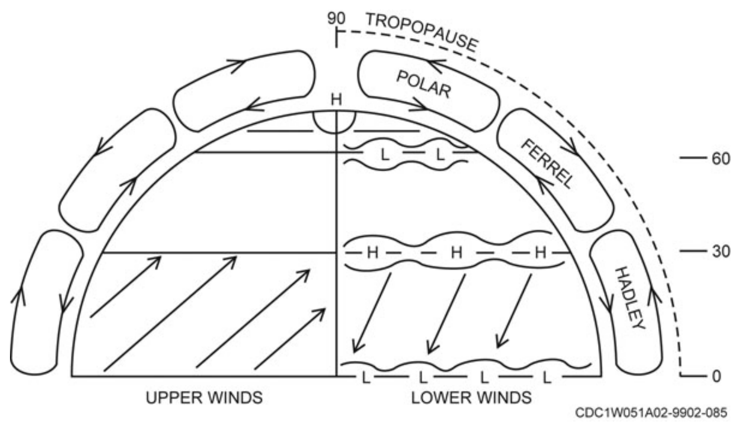

Figure 1–8. Three-cell circulation.

Hadley cell

George Hadley, an English meteorologist, theorized this first circulation cell in 1735. The Hadley cell is the strongest of the three cells of circulation and is formed as warm air rises above the equator and starts to flow northward. The northward flow deflects to the right, due to Coriolis, becoming an upper-level westerly flow. As this air moves northeastward toward the pole, it cools and a portion of it sinks at about 30°N. This sinking air spreads northward and southward as it nears the surface. The southward moving air again deflects to the right, becoming the northeasterly trade winds.

Because of the circulation in the Hadley cell, two pressure belts are created. The first is a belt of semipermanent high pressure that results from the sinking air at 30°. This belt of high pressure is called the *subtropical ridge*. The second pressure belt is a trough of low pressure near the equator. It's called the *near equatorial trough*.

Polar cell

This is the northernmost cell of circulation and its mean position is between 60°N and the North Pole. At the pole, cold, dense air descends causing an area of subsidence and high pressure. As the air sinks, it begins spreading southward. Since the CoF is strongest at the poles, the southward moving air deflects sharply to the right. This wind regime is called the *surface polar easterlies*, although the upper winds are still predominantly from the southwest. Near 60º N, the southeasterly moving air moving along the surface collides with the weak, northwesterly surface flow that resulted from spreading air at 30°N. This colliding air rises, creating a belt of low pressure near 60° N.

Ferrel cell

The midlatitude circulation cell between the Polar cell and the Hadley cell is called the Ferrel cell. This cell is named after William Ferrel, a Nashville school teacher who first proposed its existence. Oddly enough, Mr. Ferrel published his observations in a medical journal in 1856.

The Ferrel cell circulation isn't as easily explained as the Hadley and Polar cells. Unlike the other two cells, where the upper and low-level flows are reversed, a generally westerly flow dominates the Ferrel cell at the surface and aloft. It's believed the cell is a forced phenomena induced by interaction between the other two cells. The stronger downward vertical motion and surface convergence at 30°N coupled with surface convergence and net upward vertical motion at 60° N induces the circulation of the Ferrel cell. This net circulation pattern is greatly upset by the exchange of polar air moving southward and tropical air moving northward. This best explains why the midlatitudes experience the widest range of weather types.

005. Circular motion and the forces involved

Up to this point, we've addressed motion in a straight line. Now, we'll explore the forces involved in circular motion. To do this, we'll first review some useful laws of motion set forth by English mathematician Sir Isaac Newton (1642–1727):

- An object at rest tends to stay at rest and an object in motion tends to stay in motion at the same velocity (be conserved) until it's acted on by an outside force.
- For every action, there's an equal and opposite reaction.
- The acceleration of an object as produced by a net force is directly proportional to the magnitude of the net force, in the same direction as the net force, and inversely proportional to the mass of the object.

Basically, Newton's first law of motion states that a body will remain in a constant state of motion if no force interferes with that motion. Therefore, in order for objects to move in other than a straight line, other forces must act upon the body. The forces we address here are of great importance in understanding the motions in the atmosphere that we'll address later. In this lesson, we'll address these three major forces at work as they apply to circular motion:

- Centripetal force.
- Centrifugal force.
- Coriolis force.

Centripetal force

Most physics textbooks define centripetal force as a center-seeking force causing objects to move in a circular path. This might lead you to believe that to get an object to move in other than a straight line, you need a centripetal generator or some such contraption. There's no such thing as a centripetal generator, but objects in space move in circular orbits for a specific reason—gravity. The force of gravity is an attracting force that affects every object in the universe. It's the "pull" of gravity that causes smaller bodies in space to orbit larger ones. Imagine a comet moving through space in a straight line. If the comet comes close to a body with enough mass (such as the star in fig. 1–9), it will be attracted to the larger mass and begins to move in a somewhat circular path around the star. The force responsible for attracting the comet isn't centripetal; instead, it's gravity.

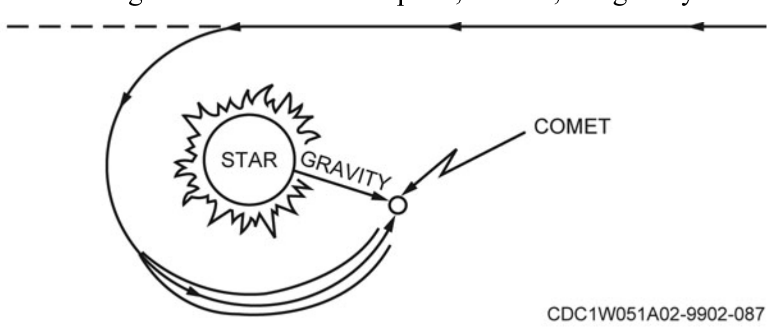

Figure 1–9. Gravity as centripetal force.

Here's another example. Let's say you have a ball on a string; if you hold the string and begin swinging the ball, it will move in a circular path around your hand. The force causing the ball to move in a circle is the result of you holding the string, not of centripetal force.

Any object moving in a circular path has an external force acting on it that's perpendicular to the object's path and pointed toward the axis of circulation. This force may be gravity, you holding a string tied to a ball, or between a planet and its moon. In conclusion, centripetal force is *any* force that deflects an object from a straight path.

Centrifugal force

Recall that Newton's second law states that for every action there's an equal and opposite reaction. If this is true, then for objects moving in a circular path, there must be some force acting against the center-seeking force. This reaction is termed centrifugal force (CeF). Where does it come from? Think about our earlier example of the ball on the string. There is no force pulling the ball outward; instead, the only force acting on the ball is the inward pull of the string. The opposite reaction to that force is the outward pull of the ball against the string. If the string breaks, the ball won't move outward on a curved path—it will move in a straight line perpendicular to the line created by the string at the moment it breaks (fig. 1–10). CeF is inertia resisting the inward force of the string.

Now let's look at CeF from another frame of reference. Imagine that the ball on the string is hollow and is partially filled with water. If the ball isn't accelerating, the water will rest in the bottom of the ball. If the ball is twirled on a string, what happens to the water? It's forced to the side of the ball away from the string or opposite to the center-seeking force shown in figure 1–11. The force responsible for this is CeF.

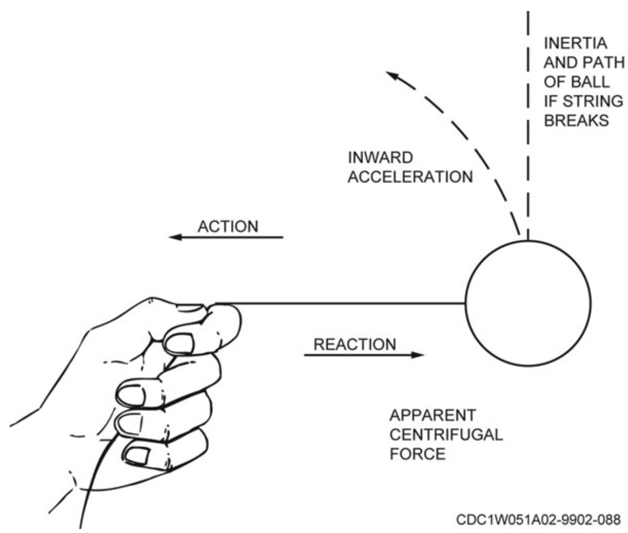

Figure 1–10. Ball and string experiment to show centrifugal force.

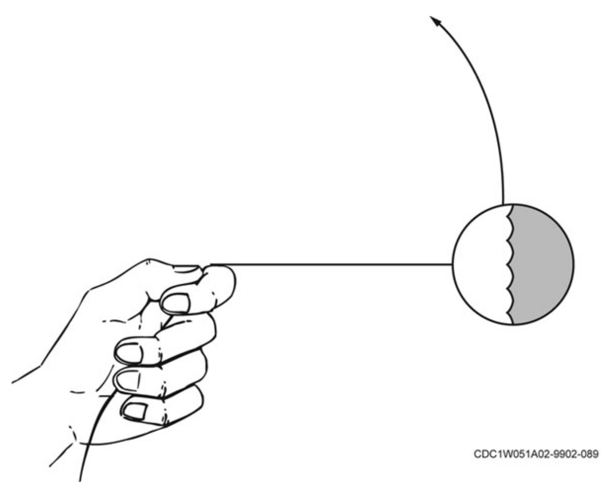

Figure 1–11. Effect of centrifugal force on water-filled ball.

The strength of CeF is dependent on these three things:

- Mass.
- Velocity.
- Radius of rotation.

CeF is directly proportional to mass and velocity, and inversely proportional to the radius of rotation. This simply means that if either mass or velocity increases or decreases, then CeF will do likewise. This force is illustrated every time you exit off the interstate. As you turn the wheel of your vehicle, you begin to feel an outward-pulling force. To compensate for this outward pull, you apply the brakes, thus decreasing your velocity. Now imagine two vehicles traveling at the same speed and entering the exit ramp. One vehicle is a small sports car and the other a tractor-trailer. The tractor-trailer experiences a much greater CeF because of its larger mass.

If an orbiting object is accelerated, CeF becomes stronger than the balancing center-seeking force. For example, if you swing a heavy ball fast enough, it breaks the string. The same is true when gravity is the center-seeking force. This idea was used to get spacecraft out of the earth's gravitational field to fly to the moon.

Coriolis force

Coriolis force (CoF) is a force created by the cyclonic rotation of the earth that acts on any body moving above the earth's surface. This force causes objects in motion to be deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Because Coriolis is a force, you'd expect that it causes an acceleration on a moving object. Actually, the deflection due to Coriolis is an illusion, or often referred to as an apparent force.

To help you understand what the CoF really is, let's consider a parallel event. Imagine you're throwing a rubber ball at a car that's moving from right to left past the point where you're standing. If you throw the ball straight at the car the exact instant it passes you, the ball will miss its target to the right. To hit the car, you must throw the ball at some point ahead of the car before it passes directly in front of you.

Now imagine you're in the car and trying to hit a stationary object with the ball as you pass the object. If you don't throw the ball at some point before you're directly in front of the object, it will again miss its target to the right. Figure 1–12 illustrates the reason. There are actually two forces acting on the ball as it leaves your hand—the force you use throwing the ball and the inertia from the forward motion of the car. The path the ball travels is the vector sum of the two forces following Newton's second law.

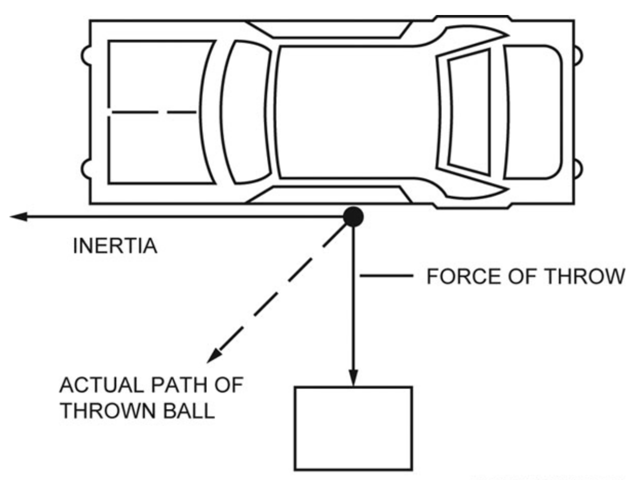

Figure 1–12. Throwing a ball from a moving car at a stationary object.

Now consider two points on the earth, both on the same line of longitude—one at the North Pole, the other at the equator. Both points make exactly one revolution per day around the pole. Still, as a stationary observer in space, you'd see that the point at the equator is moving faster than the point at the pole. Because the earth's circumference is greater at the equator than at the pole, the point at the equator is moving faster. This is similar to two people running around a circular track, one on the inside and the other on the outside. The runner on the outside must run faster to complete a lap in the same time as the runner on the inside—the runner on the outside of the track has farther to run.

Again, from your stationary platform in space, imagine that a missile is launched from the North Pole toward a target on the equator as seen in figure 1–13. As you observe the missile, it flies in a straight line; but, when it reaches the equator, its target has moved out of the way because the target is moving faster than the point of launch. To the person who launched the missile, it would appear as if the missile had veered to the right of its target. This is the same effect you got when you tried to hit the moving car with the ball.

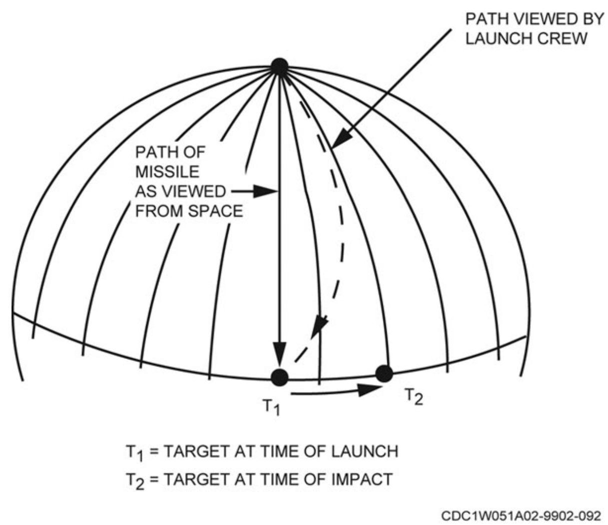

Figure 1–13. Coriolis affecting a southward moving projectile.

If we reverse the missile and target positions (fig. 1–14), the same thing will happen, except this time the target is moving slower than the launch point. This would be the same as when you throw a ball from a moving car at a stationary object. The missile, like the ball, has two forces acting on it, and its path will be the vector sum of those two forces.

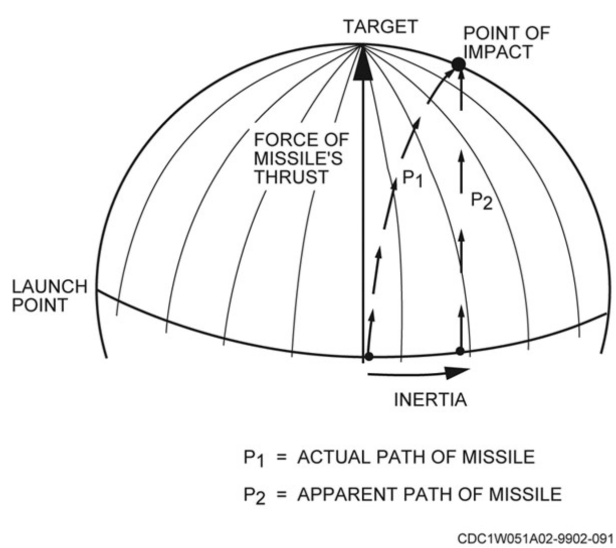

Figure 1–14. Coriolis affecting a northward moving projectile.

006. Forces affecting earth's general circulation

Differential heating sets into motion and largely sustains the earth's circulation pattern. Yet once in motion, the general circulation pattern is acted upon by a combination of the forces described here.

Pressure gradient force

The pressure gradient force (PGF) starts the horizontal movement of air over earth's surface. It is due to differences in atmospheric density that result in horizontal differences in atmospheric pressure. A pressure gradient is a difference in atmospheric pressure from one point to another. The direction of PGF is always toward lower pressure, perpendicular to the isobars. This is why air starts to flow from high pressure at 30°N toward low pressure at the equator. The strength of the PGF depends on the horizontal difference in pressure over distance (i.e., pressure gradient).

For example, a change in pressure of 20 millibar (mb) over 10 nautical miles produces a greater PGF than a change of 10 mb over the same distance. The left diagram on figure 1–15 shows an example of PGF with a change of 10 mb over 200 nautical miles. Now, look at the right diagram of figure 1–15 and notice how the PGF is perpendicular to the isobars and points towards lower pressure.

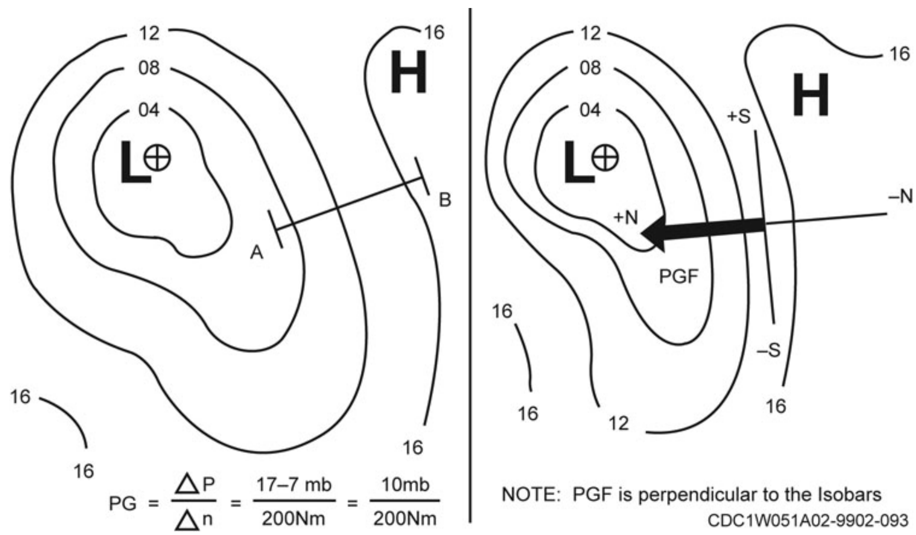

Figure 1–15. Pressure gradient force.

PGF is directly proportional to the strength of the pressure gradient. If the pressure gradient increases (isobars become more closely spaced), then the PGF increases. If the pressure gradient decreases (isobars spread out), then the PGF also decreases.

Contour gradient force

Contour gradient force (CGF) is the force that represents PGF on a constant-pressure product. It's the rate of height change with change in distance on a constant-pressure surface. Much like hills and valleys on the earth's surface, contours correspond to the higher and lower height values, respectively, on a constant-pressure surface.

Differences in virtual temperature and the quantity of mass over a given point in the atmosphere result in changes in thickness between layers. These variations in height result in the contour gradient.

CGF is directly proportional to the strength of the contour gradient. If the contour gradient increases, then the CGF also increases. Conversely, if the contour gradient decreases, then the CGF decreases. Figure 1–16 shows an example of two different contour gradient forces. Notice on the left side of the diagram how the contours are closely spaced (tighter gradient) while the contours spread out on the right side (weaker gradient). The larger arrow indicates stronger CGF on the left while the smaller arrow indicates a weaker CGF on the right. Also notice how both CGF arrows point towards lower heights.

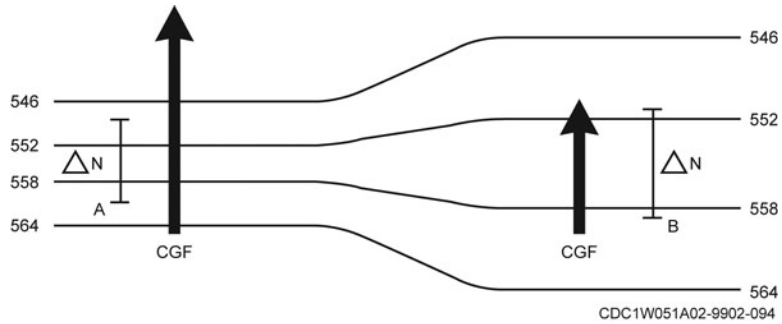

Figure 1–16. Different contour gradients.

Coriolis force

We discussed the CoF in detail earlier. Remember, this is the force that causes any mass moving free of the earth's surface to be deflected from its intended path. The deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Deflection due to CoF is greatest at the poles and is directly proportional to speed of the mass.

Centrifugal force

Also discussed in detail earlier, this force is the equal and opposite reaction to any center-seeking force. Its strength is directly proportional to speed and inversely proportional to the radius of rotation.

Frictional force

Frictional force (Fr) directly opposes and retards the motion of one mass in contact with another. The strength of the force depends on the nature of the contact surface. The more irregular the contact surface, the greater is the Fr. Therefore, air moving over water is least affected by friction; air moving over mountains is most affected by friction. At heights greater than 3,000 feet above the surface, friction has little effect on air movement. Friction always acts opposite to the direction of motion. With an increase in friction, the wind velocity decreases.

Each of the forces we discussed has an effect on the flow pattern in the earth's atmosphere. The two forces having the greatest effect on winds are pressure gradient force and CoF. The PGF initiates the horizontal movement of air. As the air starts its motion toward lower pressure, CoF deflects it to the right. When these two forces are equal, the resulting windflow is at right angles to PGF/CGF—this is called *geostrophic windflow* (fig. 1–17). Notice the flow is parallel to lines of equal pressure (isobars/contours) with lower pressure/heights to the left of the direction of flow (V).

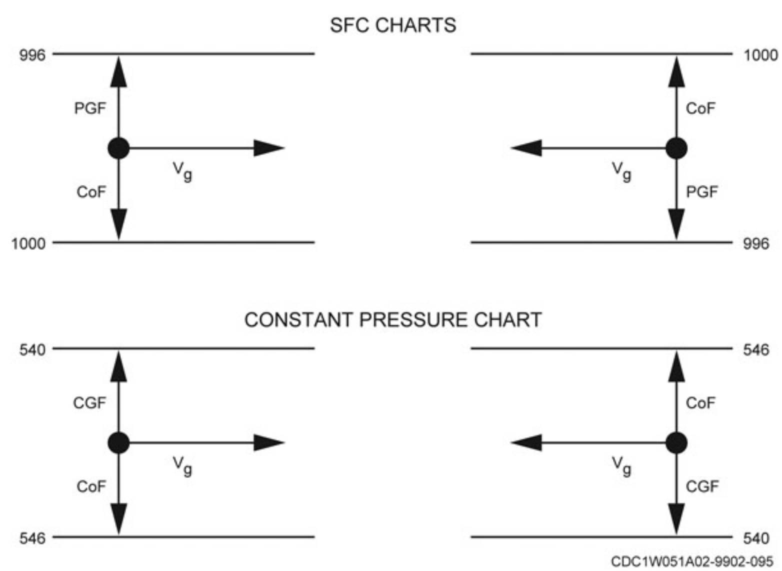

Figure 1–17. Geostrophic windflow.

This is the basis for Buys Ballot's law which states, "If the wind is at your back in the Northern Hemisphere, lower pressure is to your left and higher pressure is to your right."

Wind tends to follow the geostrophic pattern (parallel to the isobars) until an imbalance occurs between PGF and CoF. Friction is one possible cause for this imbalance to occur. When friction is present, speed is reduced, which lessens the effect of Coriolis. Since Coriolis is now weaker than PGF, air moves across the isobars toward lower pressure (V_g). Figure 1–18 shows an example of PGF without friction and an example with friction. Friction causes both CoF and CeF to decrease.

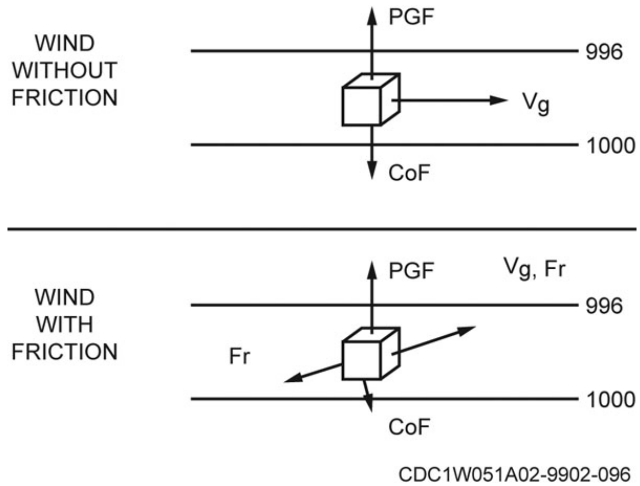

Figure 1–18. Effects of friction on geostrophic flow.

Winds tend to back in direction and slow with an increase in friction. For example, if an aircraft were descending from 2,000 feet down to the surface (using the diagram on the left of fig. 1–19), the winds would change from a west-southwest wind to a southwest wind and decrease in speed due to increased friction. Conversely, if the aircraft were ascending from the surface to 2,000 feet, the winds would veer and increase speed.

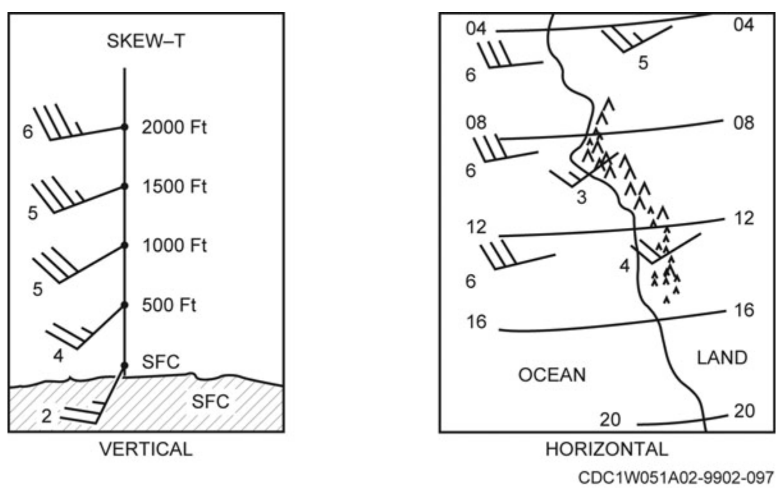

Figure 1–19. Effects of friction in the vertical and horizontal.

Circulation around pressure systems

As we stated before, air tends to flow parallel to lines of equal pressure. This is true even if the isobars are curved around high- and low-pressure centers. Curved flow that's parallel to isobars is called gradient flow. Gradient flow, due to its curvature, has the added effect of CeF. To illustrate gradient flow, we'll examine the balance of forces involved with the two types of pressure circulations.

High pressure

Air moving around high pressure is termed *anticyclonic circulation*. In the Northern Hemisphere, anticyclonic circulation is clockwise rotation. To get anticyclonic gradient circulation (V_a), there must be a balance of CoF against PGF and CeF. This occurs if the pressure gradient is constant around the high-pressure center. Figure 1–20 shows the balance of forces for gradient anticyclonic circulation. Also, notice the equal spacing of the circular isobars in the figure. This represents an equal pressure gradient around the pressure center.

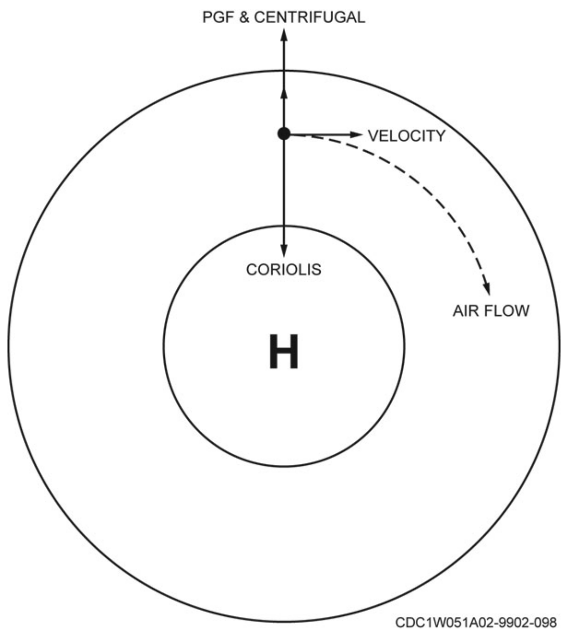

Figure 1–20. Gradient high-pressure circulation.

If friction affects circulation, air speed is reduced, which again decreases the effects of CoF. Then, as with geostrophic flow, the CoF can no longer balance the PGF and air moves across the isobars toward lower pressure. This is divergent flow—it's diverging or flowing away from the pressure center.

Examine figure 1–21 closely to understand the balance of forces associated with divergent anticyclonic circulation.

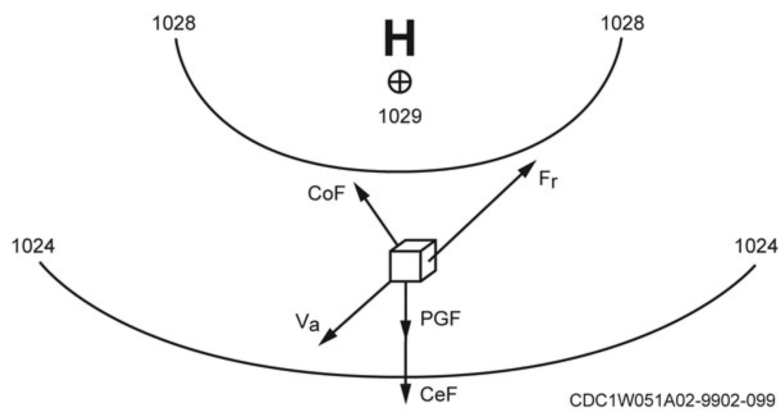

Figure 1–21. Divergent high-pressure circulation associated with friction.

Low pressure

Low pressure, or cyclonic gradient circulation (V_c), is the exact opposite of anticyclonic gradient circulation (V_a), with PGF directed inward. Were it not for CoF, air movement would be directed toward the low center. Still, CoF deflects the movement of air to the right of PGF, resulting in a counterclockwise rotation. As with high-pressure circulation (if we discount friction) the forces balance and a gradient flow circulation results. Notice CoF and CeF now balance against PGF (fig. 1–22).

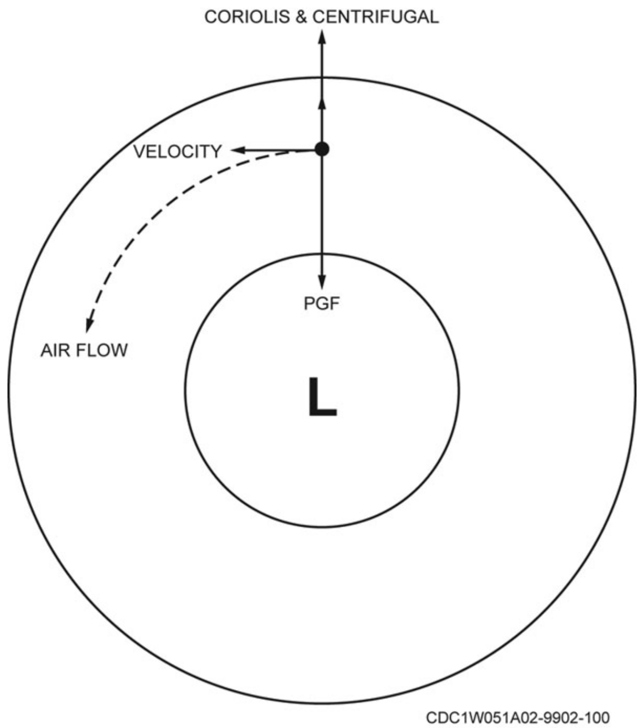

Figure 1–22. Gradient low-pressure circulation.

When we add friction, as we show in figure 1–23, velocity is again reduced, lessening the effect of CoF. The result is the PGF overcomes the CeF and CoF with flow spiraling inward toward the center of the low. This inward directed flow is known as *convergent, cyclonic circulation*.

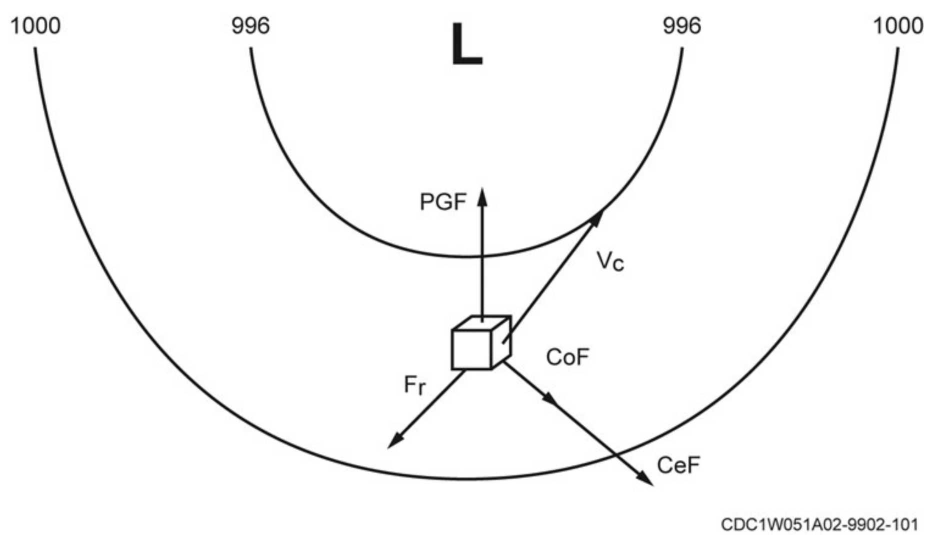

Figure 1–23. Convergent low-pressure circulation.

Confluence

Confluence is the merging of wind flow into a common axis of flow (fig. 1–24). As the flow compresses, the air parcels are constricted to a smaller area. The effect is much like putting your thumb over the end of a garden hose; the velocity of the parcels increase.

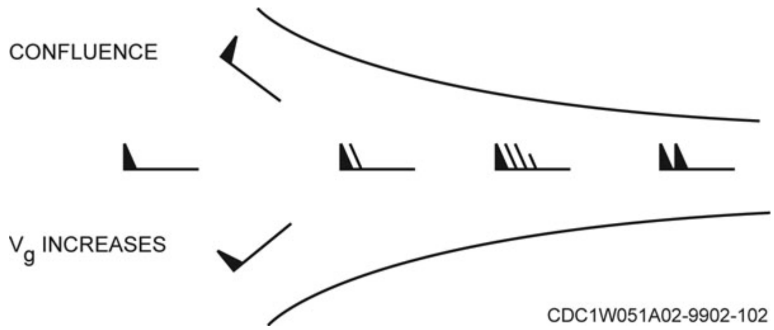

Figure 1–24. Confluence.

Difluence

Difluence is the spreading apart of wind flow from a common axis. The effect is similar to removing your thumb from the end of a garden hose; the velocity of the parcels decrease (fig. 1–25.)

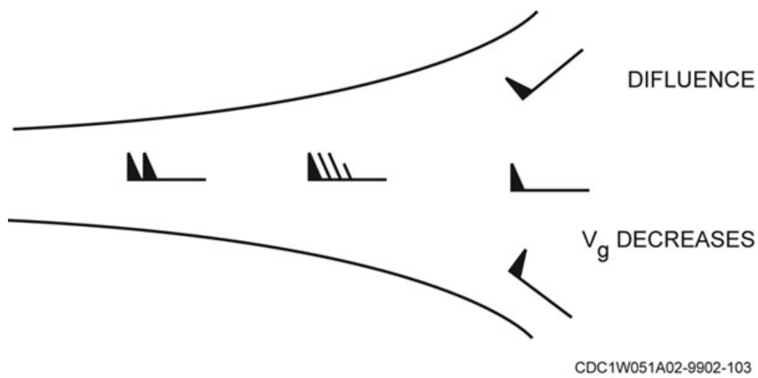

Figure 1–25. Difluence.

Supergradient winds

A supergradient wind (V_{super}) is a wind that's adjusting to a rapid decrease in CGF downstream. The terms upstream and downstream can be a little confusing, so here's an example to clarify them. If you were rafting down a river, upstream would be where you came from, while downstream would be where you're going. Now, let's get back to supergradient winds. The wind temporarily exceeds that of the new, weaker CGF. Supergradient winds occur in areas of difluence where the CGF rapidly decreases. Due to the linear momentum, the air parcels decelerate slowly (temporarily creating an unbalanced situation) and a deflection to the right of flow occurs (figure 1–26, top diagram.) The bottom diagram of figure 1–26 shows a situation where supergradient winds would occur. Notice how the stronger winds upstream are moving into an area of weaker winds.

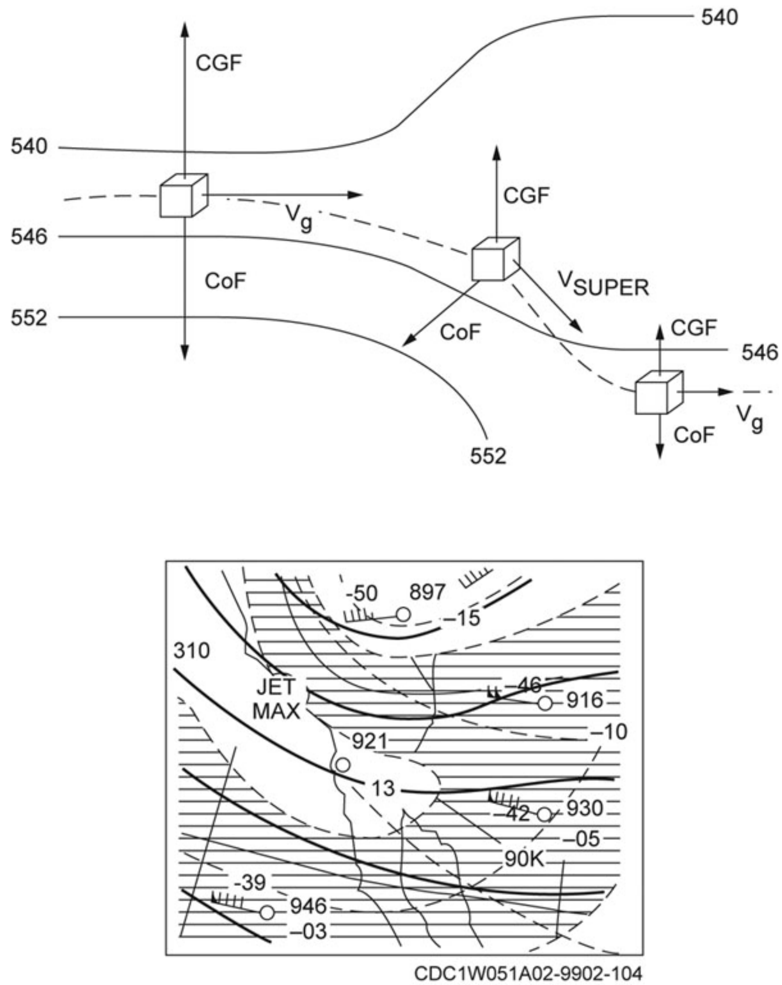

Figure 1–26. Supergradient wind forces.

CoF causes the deflection because it remains stronger than CGF (or PGF if we were looking at surface features) and remains stronger until the winds slow to the new gradient speed. As the winds deflect towards higher heights, the parcels are converting kinetic energy to potential energy.

What do we mean by kinetic and potential energy? First, you must understand the variable relationships of kinetic and potential energy.

Kinetic energy

Kinetic energy is energy resulting from motion and is directly proportional to mass and velocity. If the mass or velocity of an object is increased, the kinetic energy increases. Conversely, if the mass and velocity of an object are decreased, its kinetic energy decreases.

Potential energy

Webster's Dictionary defines potential energy as energy that's the result of relative position instead of motion. It's directly proportional to mass and height. If the mass or height of an object is increased, the potential energy increases. Conversely, if the mass and height of an object is decreased, its potential energy decreases.

In our discussion of V_{super} we stated that parcels are deflected towards higher heights. Therefore, the potential energy increases due to the parcel moving towards the higher heights. We also stated that the winds slow to the new gradient speed. Therefore, kinetic energy decreases due to the velocity decreasing. The process is completed as a conversion and isn't an abrupt change from one energy to the other. This is how V_{super} converts kinetic energy to potential energy.

Subgradient winds

A subgradient wind (V_{sub}) is a wind adjusting to a rapid increase in CGF (or PGF) downstream. It's *temporarily* weaker than that of the new, stronger CGF (or PGF).

Subgradient winds occur in areas of confluence and are in the process of accelerating. Air parcels convert potential energy to kinetic energy by turning towards lower heights or pressure. As potential energy decreases, the kinetic energy increases and the wind crosses contours at slight angles towards lower values. The bottom diagram of figure 1–27 shows slower winds upstream from the stronger winds downstream.

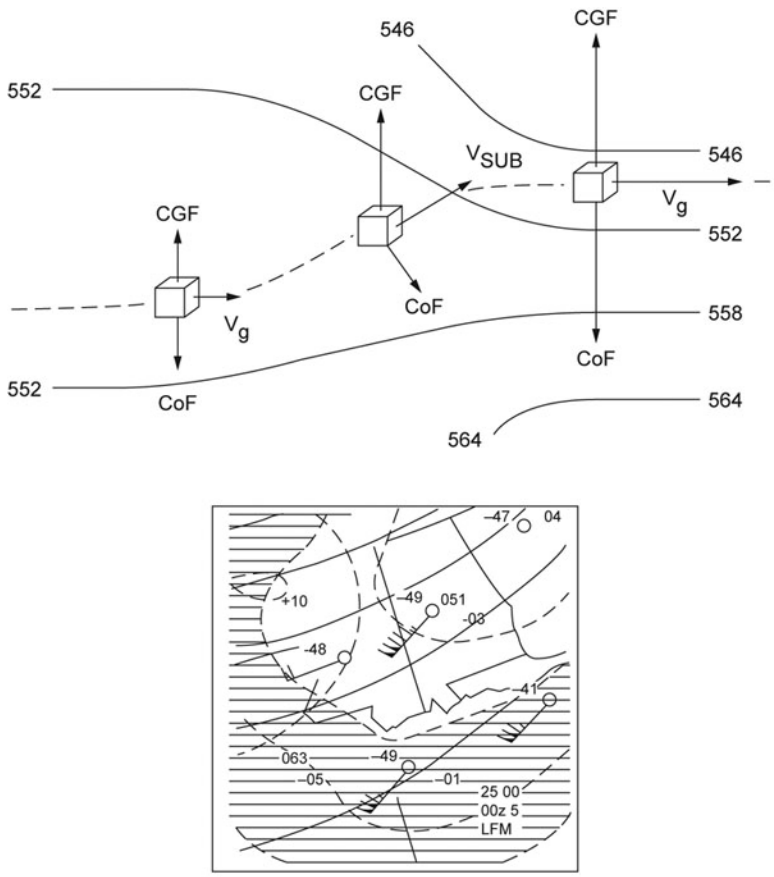

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Figure 1–27. Subgradient wind forces.

The deflection to the left is caused by the increase in CGF (or PGF). CGF (or PGF) increases immediately upon entering a confluent area and remains stronger than CoF until winds increase speed to new gradient speed. While the winds are subgradient, they deflect to the left toward lower heights (or pressure) because of an imbalance of forces between CGF (or PGF) and CoF.

Cyclostrophic wind

Cyclostrophic wind (V_{cyc}) is a wind that occurs when CGF (or PGF) is balanced by CeF. CoF is negligible when compared to CGF (or PGF) or CeF. This circulation can only exist around low-pressure areas where CGF (or PGF) is exactly opposite CeF (fig. 1–28.)

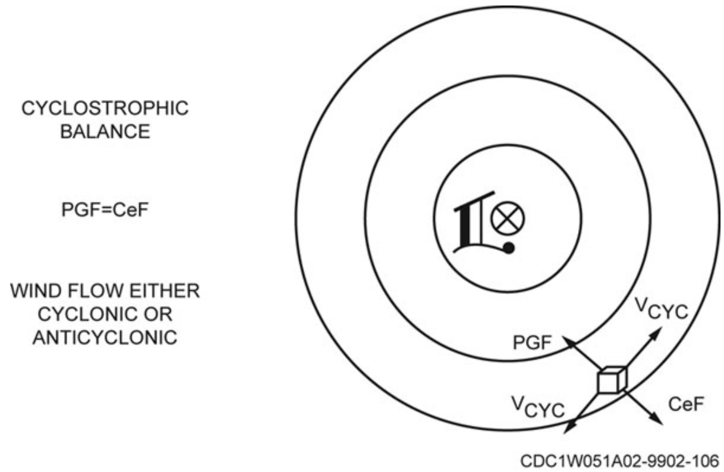

Figure 1–28. Cyclostrophic forces around a low.

This circulation *cannot form* around a high-pressure area because CGF (or PGF) and CeF both point in the same direction (fig. 1–29.)

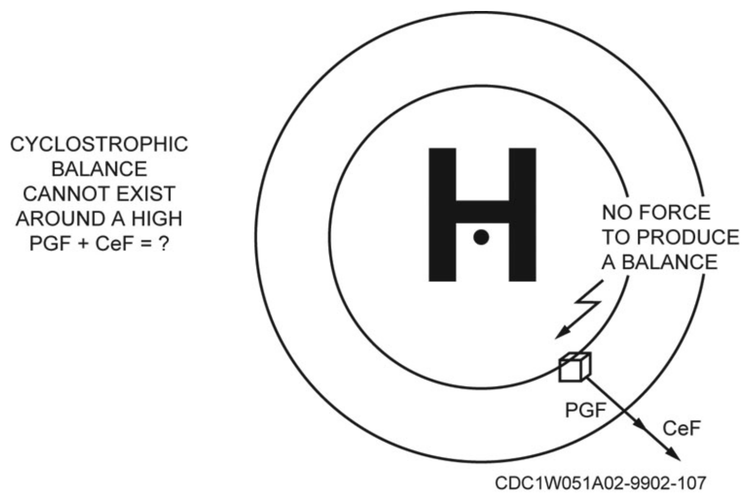

Figure 1–29. Cyclostrophic forces around a high.

Some examples of cyclostrophic wind flow in the atmosphere include hurricanes, tornadoes, dust devils, and waterspouts. Hurricanes occur in the tropics where CoF is small.

In the midlatitudes, some systems with a very strong PGF and a small radius of rotation result in a large CeF, often forming dust devils, tornadoes, or waterspouts. Due to the extreme PGF (or CGF), either clockwise or counterclockwise flow is possible around these small-scale, low-pressure systems. CoF is only a very small part of the balance of forces since CeF and PGF (or CGF) are very large.

The actual wind is difficult to observe accurately because of the many disruptions to the flow (gust, eddies, turbulence, vertical motion, etc.) and our observation limitations.

007. Earth's circulation systems

The distribution of land and water over the surface of the earth alters the general circulation pattern. The unequal distribution of land and water masses amplifies the effects of unequal heating of the earth's surface. In addition, the belts of high and low pressure described earlier are actually broken into separate cells of high and low pressure. Besides these semipermanent features, a system of migratory pressure systems contributes to the general atmospheric circulation system.

Semipermanent pressure systems

Semipermanent pressure systems are largely the result of the general circulation pattern. They're called *semipermanent* because they can usually be found at the same place with the same intensity during the same month each year. You should become familiar with several of these semipermanent high- and low-pressure systems.

Semipermanent highs

The belts of high pressure described as existing near 30° in both hemispheres are actually divided into distinct circulation centers. These centers comprise the subtropical ridge. The locations of the centers of these large anticyclonic circulation's are over the oceans between the continents. There are two in the Northern Hemisphere:

- The North Pacific high, located over the Pacific Ocean to the west of California.
- The North Atlantic or Azores high, located west of the African coast near the Azores.

A third high-pressure circulation exists southeast of the United States near Bermuda. This circulation is an extension of the North Atlantic high. Appropriately, it's called the *Bermuda high*.

There are three subtropical highs in the Southern Hemisphere. Their locations are as follows:

- South Pacific Ocean.
- South Atlantic Ocean.
- Indian Ocean.

Subtropical high-pressure systems exist throughout the year, though seasonal variations in location and intensity do occur. Circulation centers migrate north and south with the seasonal variation in the solar angle of incidence. The maximum intensity of these systems occurs in the summer months for their respective hemispheres. For example, the North Atlantic system is strongest in July, while the South Atlantic system is strongest in January.

There's another persistent type of high-pressure system in the Northern Hemisphere. It's called the Siberian high. Unlike the subtropical systems, this high-pressure circulation system exists only during the winter. It's a low-level feature that forms because of the cold, dense air present in this region during the winter. This pressure system has its mean position at 50°N and is over central Asia.

Another common persistent high-pressure circulation is the Polar high. This high-pressure circulation is also the result of cold temperatures. Though its name suggests a center over the North Pole, its center varies with the seasons—usually displaced toward the area of coldest temperatures. During the winter months, it appears as an extension of the Siberian high. In the summer, it's situated near the cooler air of the North Pacific. This high is also called the *Greenland* or *Canadian high*.

Semipermanent lows

An area of low pressure is along the equator. It's called the *near-equatorial trough* and appears as an elongated area of lower pressure circling the globe. This low-pressure area exists due to strong equatorial warming and convergence of the northeasterly trade winds of the Northern Hemisphere and the southeasterly trade winds of the Southern Hemisphere. Its position varies with the seasonal change in latitude of direct solar radiation. The near-equatorial trough moves northward during the Northern Hemisphere summer and southward during the Northern Hemisphere winter.

There are three other semipermanent low-pressure circulations important to general circulation in the Northern Hemisphere:

- Icelandic low.
- Aleutian low.
- Asiatic low.

Two lows corresponding to the semipermanent belt of low pressure at 60°N are the Icelandic and Aleutian lows.

Icelandic low

The Icelandic low is prevalent throughout the year. It's located in the North Atlantic near Iceland and Greenland. Its existence is due to cold temperatures of the ice caps of Iceland and Greenland contrasting to warmer water temperatures of the North Atlantic Ocean. This contrast is greatest in the winter; therefore, the Icelandic low is most intense during this season.

Aleutian low

The location of the Aleutian low is in the Gulf of Alaska near the Aleutian Islands. Its presence is explained by a contrast in land and water temperatures, as was the Icelandic low. Yet, the Aleutian low is less persistent in the summer because of the influence of the Polar and North Pacific highs.

Asiatic low

The Asiatic low dominates the entire Asian continent during the summer. The mean position of the center of this low is east of the Persian Gulf and just north of 23°. In this place, the Asiatic low dominates the same region as the Siberian high but in different seasons. The Asiatic low forms because increased summer heating in the Southern Asian Deserts replaces the cold Siberian high that dominates Asia in winter.

Semipermanent low pressure in the Southern Hemisphere is located in a somewhat continuous band around the globe at 60°S. This is very close to the pressure pattern suggested by the three-cell theory. This continuous belt exists because of a lack of significant land masses at this latitude in the Southern Hemisphere. In contrast, the Northern Hemisphere has a large concentration of land area in the latitudes around 60°.

The other areas of prevalent low pressure in the Southern Hemisphere are over the continents during January. These low-pressure circulations are the result of rapid heating of these land masses during the summer of the Southern Hemisphere.

Migratory pressure systems

Changes in the circulation pattern discussed thus far have been caused by seasonal changes in semipermanent pressure systems. Yet, the general circulation pattern of the atmosphere includes wind systems that continually migrate. Migratory systems comprise both anticyclonic (high-pressure) and cyclonic (low-pressure) circulations. These circulations are responsible for the rapidly changing weather conditions found especially in the middle latitudes. This occurs because the migratory systems are responsible for the transportation of polar and tropical air into the middle latitude regions. When these two types of air with greatly different characteristics clash, the result is stormy weather, such as violent spring thunderstorms or heavy winter snowfalls. Usually, fair weather is associated with migratory high-pressure systems while stormy weather is associated with migratory lows.

Self-Test Questions

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

004. Thermal and three-cell circulation

1. Briefly describe conditions that cause a single-cell circulation pattern.

2. What's the *main* driving mechanism responsible for the earth's large-scale atmospheric circulations?

3. Match the characteristics of each of the three circulation cells in column B with the name of the cell in column A. Items in column B may be used more than once.

Column A	*Column B*
____ (1) Strong easterly surface winds are found beneath this cell.	a. Hadley cell.
____ (2) A strong Coriolis force sharply alters wind direction.	b. Polar cell.
____ (3) Circulation in this cell is interrupted by the exchange of polar and tropical air.	c. Ferrel cell.
____ (4) The cell most responsible for subtropical belt of high pressure.	
____ (5) Northeast trade winds are found beneath this cell.	
____ (6) The cell dominated by westerly flow at the surface and aloft.	
____ (7) Strongest of the three cells.	
____ (8) Generally located between 30 and 60°.	
____ (9) The cell created by rising equatorial air.	
____ (10) Mean position is between 60 and 90°.	
____ (11) Upper-level flow in this cell is deflected eastward.	

005. Circular motion and the forces involved

1. Which force is any force that causes a body to veer from a straight path?

2. If a bucket of water is swung perpendicular to the ground, what force keeps water in the bucket from falling out even when the bucket is upside down?

3. In circular motion, what force acts against the inertia of the moving object?

4. What force causes objects to veer from their intended path on a rotating platform?

5. What force is the equal and opposite reaction to center-seeking forces?

6. What force appears to cause objects in the southern hemisphere to be deflected to the left of their intended path?

006. Forces affecting earth's general circulation

1. What are the four forces that dictate the general circulation pattern?

2. Which force starts the horizontal movement of air over the earth's surface?

3. Describe contour gradient force.

4. What kind of flow is created when PGF equals CoF?

5. What does Buy Ballot's law state?

6. What can a pilot expect the wind direction and speed to do as the aircraft descends into an area with rough terrain? Why?

7. What's the difference between confluence and difluence?

8. Where do winds deflect towards in a supergradient wind condition and what's happening to the air parcels?

9. A hurricane is an example of what kind of wind?

007. Earth's circulation systems

Identify each of these described pressure systems as semipermanent or migratory and as a high or low type of pressure system.

1. Dominates the Atlantic Ocean west of Africa.

2. Most prevalent over the oceans of the Southern Hemisphere during January.

3. Dominates Asia during the winter.

4. Located in the North Atlantic Ocean near Greenland.

5. Dominates the continents of the Southern Hemisphere during January.

6. In the summer it's located east of the Persian Gulf near 23°N.

7. The influence of the Polar high makes it less prevalent in the summer.

8. Fair weather is associated with this system in the middle latitudes.

9. Lack of land masses near 60°S allows this to remain as a somewhat constant belt of pressure.

10. Found as an extension of another feature and is located off the Southeast Coast of the United States.

1–3. Jet Stream

The discovery of the jet stream during the closing days of World War II introduced insights on the general atmospheric circulation. Jet streams now are recognized as potent forces for inducing cyclogenesis, tornadoes, squall lines, mountain waves, and turbulence. Regions where jet streams aren't present tend to have more uniform and less violent weather.

The awareness of the jet stream also improved our understanding of the midlatitude climatic structure and enabled forecasters to strengthen the reliability of both short- and long-range prediction procedures. In this section, we'll explore these five major subject areas:

1. General structure of jet streams.
2. Polar front jet (PFJ) stream.
3. Subtropical jet (STJ) stream.
4. Life cycle of a jet stream.
5. Jet stream and migratory pressure system relationships.

008. General structure of jet streams

In this lesson, we'll give you the description of a jet stream, discuss jet stream wind shear and major jet stream systems, and describe the origins of jet streams

Description

The jet stream is a narrow belt of strong winds in the upper troposphere with speeds of 50 to 200 knots. In the Northern Hemisphere, these winds usually have a westerly component. Jet stream position varies between different latitudes and elevations around the earth. It even varies in latitude and elevation within a small geographical area. The jet stream may appear as a continuous band around the earth, but more often it gradually diminishes at one or more points and then reappears farther downstream. From the term downstream, we can compare the basic structure of a jet stream as a river of air flowing horizontally through the atmosphere.

Jet streams are normally thousands of miles in length, hundreds of miles wide, and a few miles deep. Wind speeds in a jet stream vary along each of its dimensions. Figure 1–30 is a graphic illustration of a typical jet stream. By looking into the spiraling column in the direction of flow, you can see the change in speed across the jet stream width and through its height. The greatest wind speeds through the jet stream are in the core within the interior of the jet stream. Notice the displacement of the core to the left and top within the jet stream. From this, you can see a change in wind speed over distance (speed shear) is greatest above and to the left of the jet stream core as you look in the direction of flow.

The core of the jet stream is found along its length. The core defines the jet stream axis. This concept is illustrated on figure 1–31. Notice, the axis is drawn along the band of maximum winds when you view it from the top. The figure also shows the variation in wind speed along the length of a jet stream.

Jet maxima (maxes) or jet streaks are intermittent regions of the strongest wind speeds within the jet. The isotachs which define the jet maximum are elliptically (football) shaped; both horizontally and vertically. This is shown in the two views on figure 1–32.

Core speeds of 100 to 150 knots are common within jet maxima, but speeds up to 300 knots have been observed. Jet maxima aren't stationary features; they actually move along the jet stream axis. The importance of jet maxima comes from their link with other atmospheric disturbances.

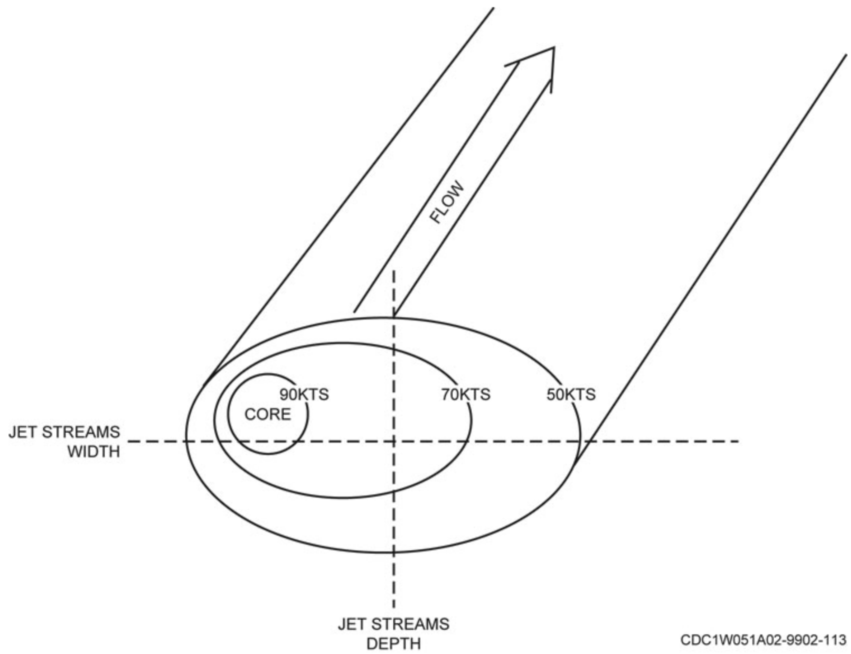

Figure 1–30. Cross section of jet stream.

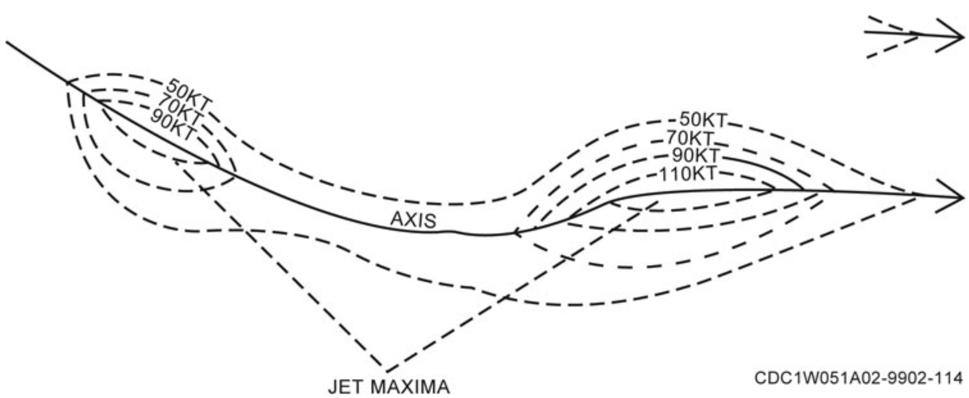

Figure 1–31. Jet stream axis and maxima.

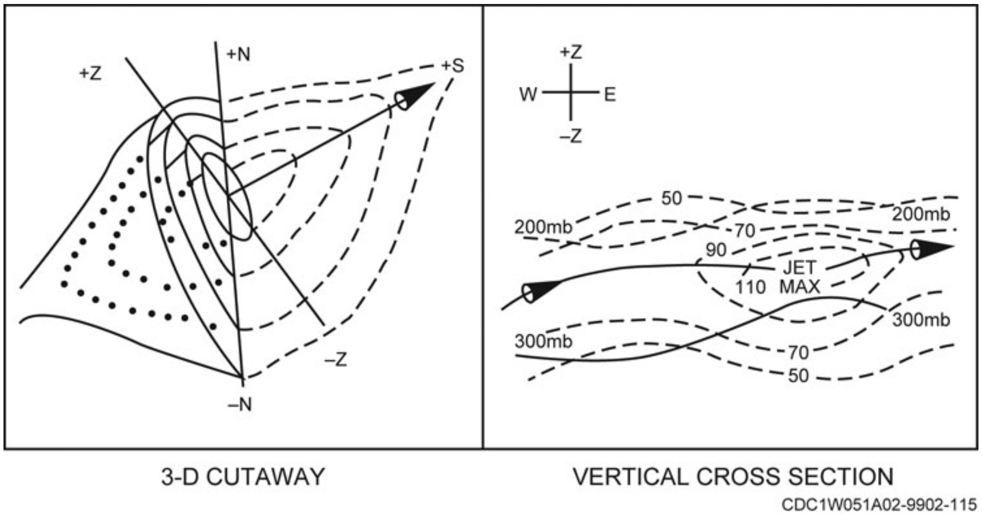

Figure 1–32. Examples of jet maxima.

Jet stream wind shear

The jet stream is significantly wider than it is deep. Though the core may be more than 100 miles wide, it may be at least 10,000 feet thick. This produces stronger vertical wind shears than horizontal shears.

Vertical

Winds in the jet core average 120 to 150 knots during the winter, with frequent occurrences above 200 knots. Since the increase to these high speeds occurs in a comparatively short distance, the shear values are also high. The highest vertical shear usually occurs above 500mb.

Shear above the core is usually greater than the shear below the core because on the average the wind gradient is greater above the core. The wind usually decreases 50 percent within 5 kilometers (km) above the core but takes 5.5 km to decrease 50 percent below the core.

Horizontal

Cross sections of the jet stream also indicate horizontal wind shear. There’s a rapid decrease of wind speed on each side of the jet core. This decrease can be as much as 100 knots in 100 miles on the north side and 100 knots in 300 miles on the south side. This indicates greater horizontal shear and turbulence north (cold air side) of the jet. This concept is illustrated on figure 1–33.

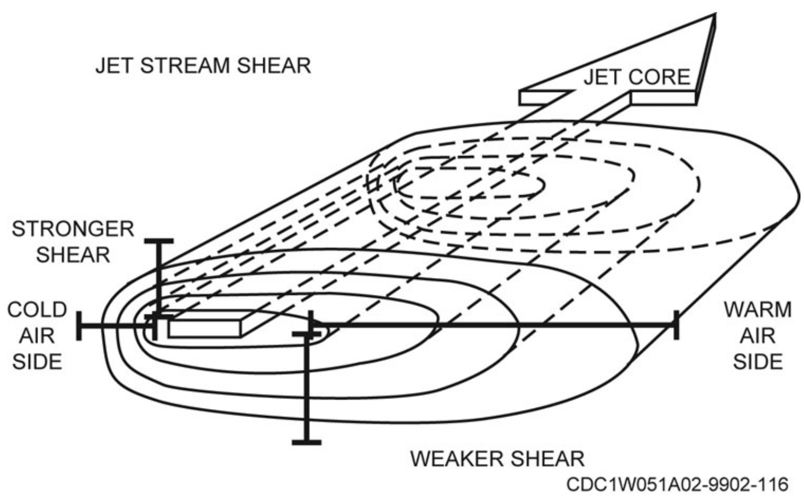

Figure 1–33. Jet stream wind shear.

Major jet stream systems

A jet stream rarely occurs alone in the atmosphere; instead, multiple jets are the rule. Two major jet streams are frequently observed:

1. Polar front.
2. Subtropical.

Figure 1–34 shows the average position of these two Northern Hemisphere jet stream cores with respect to the tropopauses. The jet core depicted between the polar and midlatitude tropopauses is called the *polar front* or *midlatitude* jet. The core between the midlatitude and tropical tropopauses is called the *subtropical* jet.

A third jet stream, called the *arctic* or *polar night jet stream*, is believed to occur during the winter at 75°N around the 300- to 350-mb level near the Arctic circle. This jet stream is most often observed in the winter, hence the name polar night.

In addition to these three jets, the tropical easterly jet is located over southern Asia and northern Africa. This jet is most frequently observed in the summer at the 150-mb level.

NOTE: Later in this unit, we'll have in-depth discussions about the characteristics of the polar and STJ streams.

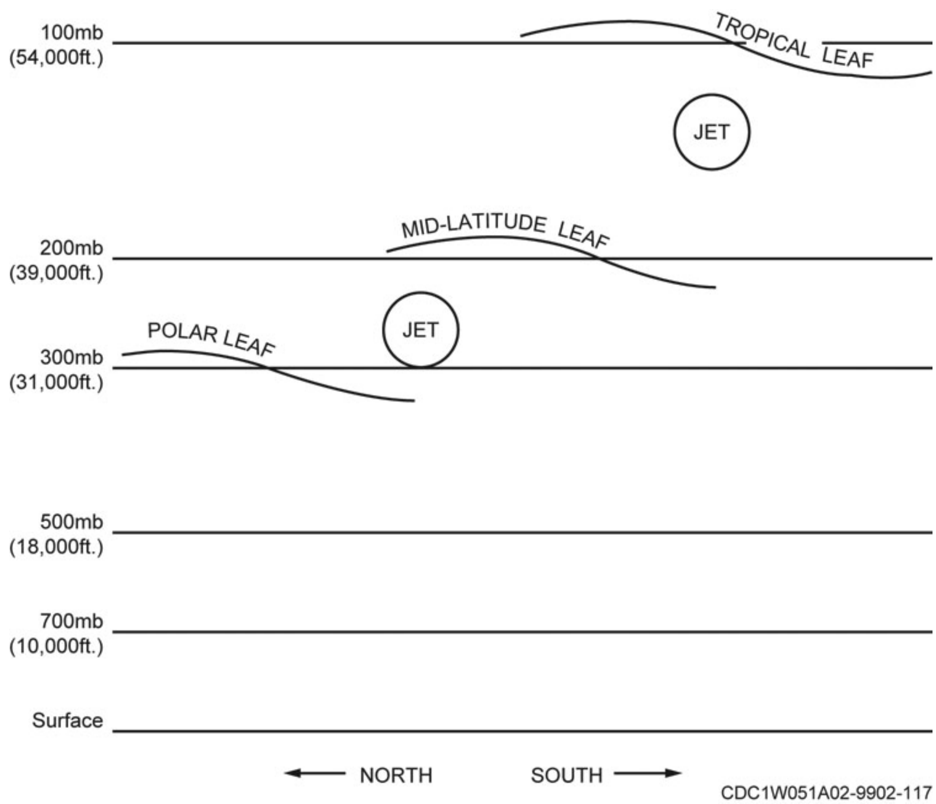

Figure 1–34. Relative position of jet cores.

Origins of jet streams

Jet streams are primarily caused by two origins. One origin of a jet stream is caused by a large horizontal temperature contrast associated with a large thermal wind vector. Jet streams that form by this large temperature contrast have a jet axis that's at the tropopause because this is where the contour gradient force is at a maximum. Above and below the tropopause, the slope of the constant-pressure surface weakens.

NOTE: Horizontal temperature contrast is the primary originator of the PFJ stream.

The second origin of a jet stream is due to the conservation of angular momentum. Recall that Newton's first law of motion states that an object at rest tends to stay at rest and an object in motion tends to stay in motion at the same velocity (be conserved) until it's acted on by an outside force.

Angular momentum is the product of the mass times velocity and radius of rotation for an object, or parcel, following a curved path. It's stated by the formula:

$$AM = m \times v \times r$$

where:

AM = Angular momentum

m = mass

v = velocity

r = radius

Angular momentum is conserved for an object when forces acting upon the body are in balance. If the mass is held constant and the radius decreases, then the velocity must increase in order to conserve angular momentum. This is known as an *inversely proportional relationship*. Our earlier example pertaining to a ball on the end of the string may help explain this. If you have a ball on the end of a string and you spin it around your finger, the string begins to wrap around your finger. As the string wraps around your finger, the length of string decreases (decreasing the radius) and, ultimately, the ball spins faster (increasing its velocity).

Because the atmosphere is a fluid and not attached to the solid earth, it tries to conserve its angular momentum. As a parcel moves poleward, the radius (of the earth) decreases with an increase in latitude, causing its velocity to increase. Conversely, with a parcel flowing equatorward, the radius increases with a decrease in latitude; as a result, its velocity decreases.

009. Polar front jet stream

The PFJ stream is the primary jet stream in the middle latitudes of the Northern Hemisphere. It may appear as a continuous band around the globe, but, more often, it vanishes at one point and reappears downstream. Its disappearance and reappearance is often due to variations in its height. Its existence can be attributed to large horizontal temperature differences (thermal gradient) across the upper and middle latitudes. These temperature differences exist not only at the surface, but vertically to the tropopause. The stronger the thermal gradient, the stronger is the jet stream.

Vertically, the PFJ is at the top of the troposphere near the tropopause zone. In fact, the polar jet divides the tropopause into two segments called *leaves.* This is illustrated on figure 1–35. The PFJ is within the warm air, south of the coldest air. It frequently marks the boundary between cold polar air and warmer subtropical air. In the summer, with temperature differences lessened by warming in the middle and higher latitudes, the PFJ is found at higher latitudes. When the middle and higher latitudes cool in winter, the polar jet migrates farther south. Winter is also the time of greatest horizontal temperature differences between latitudes and the time when the PFJ's speeds are greatest.

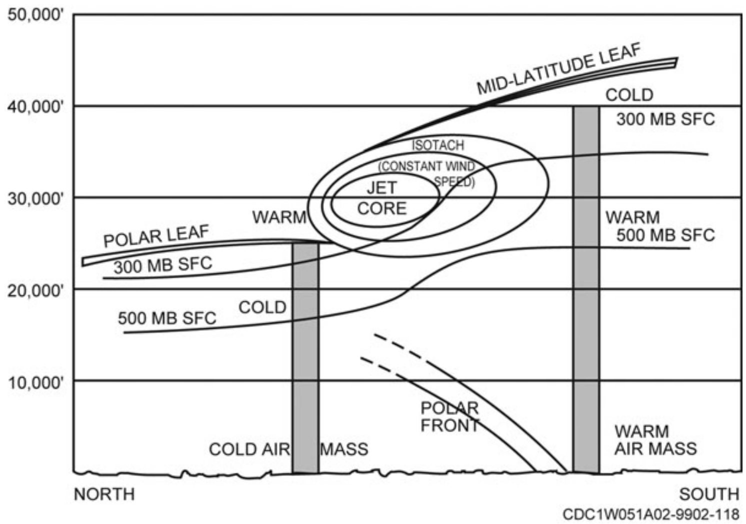

Figure 1–35. Polar jet and tropospheric leaves.

The PFJ follows a generally eastward track; still, if you view the entire pattern outlined around the globe, you should notice large north to south undulations. The undulations are called *long waves.* This concept is illustrated on figure 1–36. As you can see, the long waves consist of large low-pressure troughs and high-pressure ridges. The series of waves the jet stream follows is the long-wave pattern. Observations reveal the series of long waves created by this flow number as few as three and as many as seven around the globe—the average number found is five.

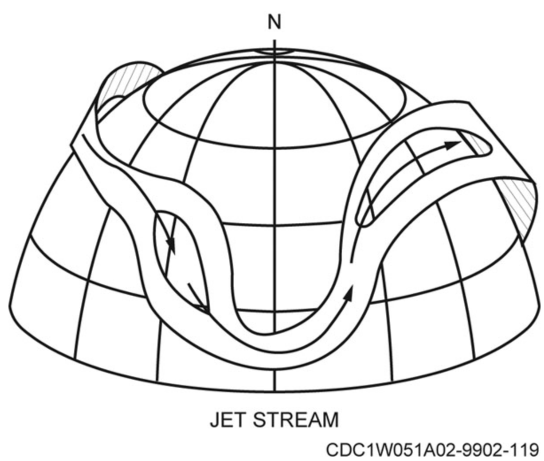

Figure 1–36. Long-wave pattern.

Jet maxima in the PFJ are of prime importance to the meteorologist because of their link to midlatitude migratory pressure systems. PFJ maxima cause and steer the major midlatitude tropospheric pressure systems. Low-pressure systems are found ahead of the maxima while high-pressure systems move behind them.

A valuable clue in locating the PFJ is the relationship between the jet stream and the polar front. The core of the PFJ is directly above the isotherm ribbon associated with the frontal zone (thermal concentration) on the 500-mb surface. This relationship has a high degree of correlation, since troughs and ridges slope very little between 500 and 300mb and have almost no slope from 300 to 200mb.

There are several methods of locating the 500-mb frontal zone. The simplest is to locate the position of the –17°C isotherm. The polar front usually cuts the 500-mb surface along the –17°C isotherm. The concentration of isotherms are located to the warm side of the transition zone, with the largest concentration between –17 and –20°C.

The other methods usually depend on locating areas of highly concentrated isotherms. A jet stream is usually associated with a thermal concentration of more than 10°C in 200 miles. Warm air is south of the concentration and cold air is north. Thus, with your back to the wind (facing downstream), the cold air is to your left and the warm air to your right.

Since the temperatures reverse above the tropopause, in a north-south direction, the jet stream will be just below this temperature reversal. This reverse occurs above 200mb. Thus, the jet stream core is usually located between 300 and 200-mb levels.

From these observations come the following three rules:

1. The 500-mb position of the jet stream is located in the –17 to –20°C isotherm ribbon, with the most frequent occurrence along the –17°C isotherm.
2. The jet stream core has been found between 8,960 meters (29,400 ft) and 9,236 meters (30,300 ft) more than 82 percent of the time during the winter.
3. The width of the jet core is about equal to the width of the 500-mb zone having the maximum temperature concentration (isotherm ribbon).

010. Subtropical jet stream

Another jet stream important to the Northern Hemisphere is the STJ stream. It exists independently, but frequently exists with the PFJ.

Two main factors contribute to the formation of the STJ. The first is the convergence of the Hadley and Ferrel cells as previously discussed in atmospheric circulation. The second factor is the horizontal temperature contrast due to the interaction of warm, moist subtropical air and cold dry polar air.

As shown in figure 1–37, the STJ position is south of the PFJ and is normally observed between 25 and 30°N, with a mean latitude of 28ºN. The STJ extends from the Hawaiian Islands eastward to southern Florida. Conservation of angular momentum is a factor that accelerates the subtropical jet, while CoF causes recurvature to the right. Remember, as the jet moves northward, radius decreases and velocity increases. At the same time, CoF is also increasing as it moves closer to the poles, thereby increasing curvature.

You can trace the STJ's axis around the entire globe. In contrast to the polar jet stream, the STJ persists at the 200- to 150-mb level. Though the STJ persists at higher altitudes than the polar jet stream, STJ winds as high as 120 knots have been observed as low as 300mb, but rarely below this level. The STJ reaches its maximum speed at its farthest projection poleward. Figure 1–34 shows the relative vertical and horizontal positions of the STJ and PFJs. Notice the STJ also divides the tropopause into leaves with the tropical leaf to the south and the midlatitude leaf to the north.

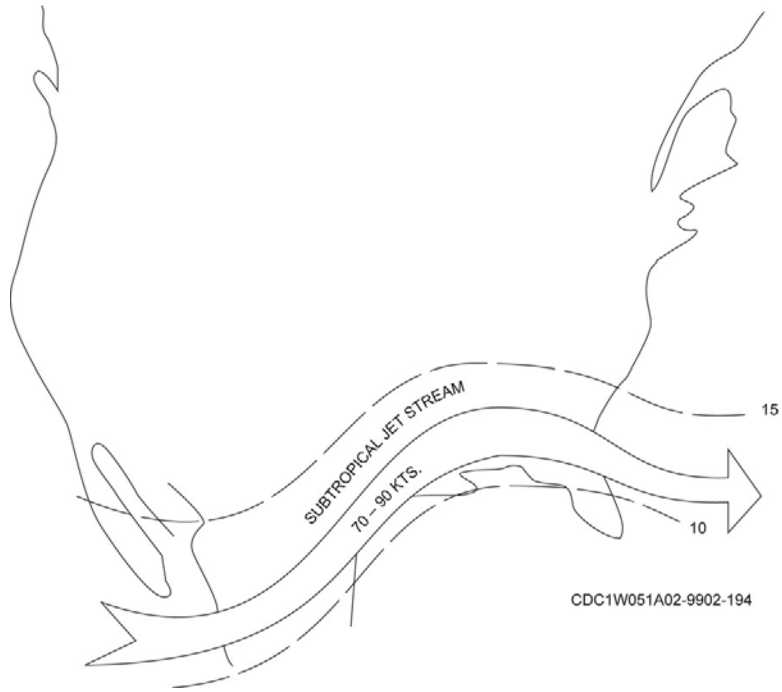

Figure 1–37. Typical position of the subtropical jet stream.

The STJ typically flows more directly from west to east. At times, the STJ moves northward and merges with the PFJ but is more often observed as a separate feature. However, when these two jet streams do interact, favorable conditions for extreme turbulence exist. This condition also favors the development of severe thunderstorms, as satellite imagery often confirms severe weather activity as a characteristic wedged-shaped cloud pattern in this region. The STJ is very important in transporting warm moist air poleward to interact with cool dry polar air. This produces strong temperature, contour, and isotach gradients associated with severe weather.

It's in the region of confluence of the PFJ and STJ streams where the STJ velocities are the greatest. This occurs partly because the STJ axis is at its furthest poleward extent and, consequently, the radius of the earth is smaller (conservation of angular momentum). Also, the confluence of the STJ with air already moving at high speeds in the PFJ results in producing a broader and stronger jet stream.

The STJ is less connected with lower tropospheric pressure systems than the PFJ. Occasionally, it does become superimposed on a lower tropospheric pressure system and can exert considerable influence on the pressure system's development. Just like the PFJ, the STJ also has seasonal variances, being stronger in the winter than in the summer and typically weak during hurricane season.

011. Life cycle of a jet stream

Jet streams undergo a distinct life cycle consisting of a period of organization and one of disintegration. Their horizontal structure, as it appears on upper-level products, depends on the phase of this life cycle. When the jet is well organized, the longitudinal axis follows the pattern of long waves. Also, its amplitude is comparable to that of the 300-mb contours. Thus, the jet stream can provide a clue of the long-wave pattern.

Figure 1–38 shows a 300-mb product with the contours removed to emphasize the horizontal structure of the jet stream. A lack of uniformity in wind speed along the jet stream axis is clear in the isotach analysis. As shown, the wind gradient along the axis is sharp and the wind speed difference between the maximum and minimum approaches 100 knots. Often, the difference between individual maximum centers exceeds 100 knots.

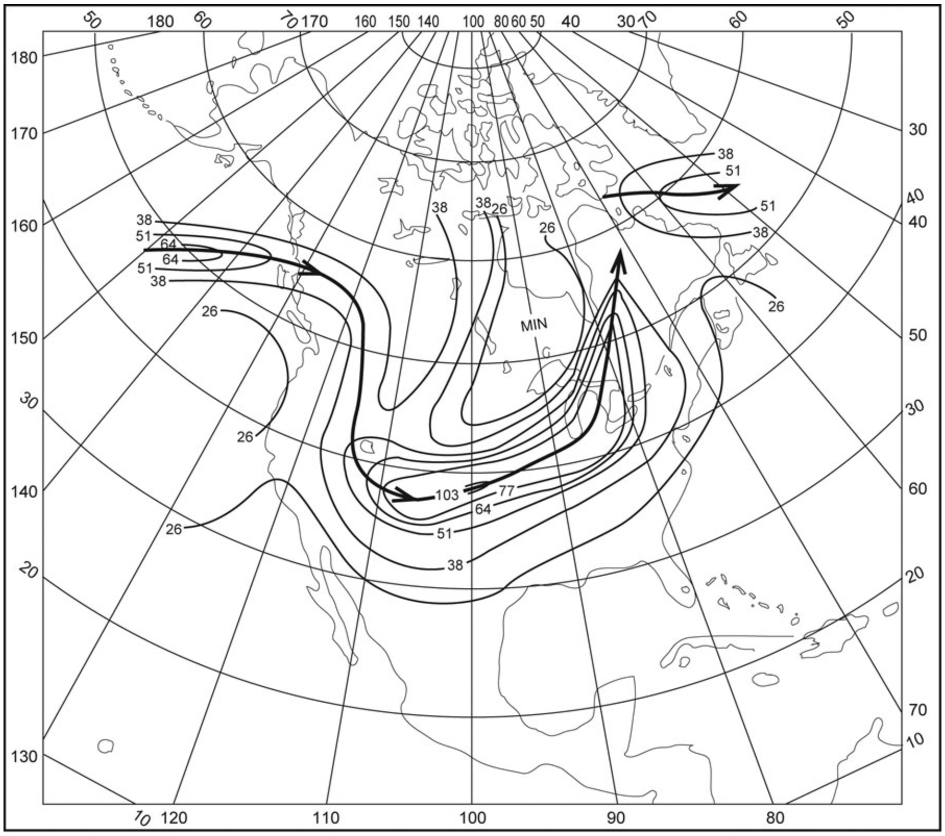

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Figure 1–38. Jet stream configuration on constant-pressure surface.

During disorganized periods, the well-defined wind speed concentrations of 50 knots or more tend to disappear. When this occurs, the westerlies either break into closed circulation patterns or show comparatively uniform velocities throughout the middle latitudes. The westerly wind speed pattern during periods of dissipation is characterized by the development of many, comparatively weak “jet fingers,” often no more than 300 to 400 statute miles apart. Figure 1–39 shows a 300-mb product with “jet fingers.”

The jet stream follows the contours closely. Small departures are often noted, however, such as when the wind speed increases suddenly in a region of comparatively weak contour gradient or decreases suddenly in an area of strong contour gradient.

Diverging contours downstream cause the jet stream to deflect toward greater heights (higher pressure at the same altitude) and converging contours downstream tend to turn the jet toward lower heights.

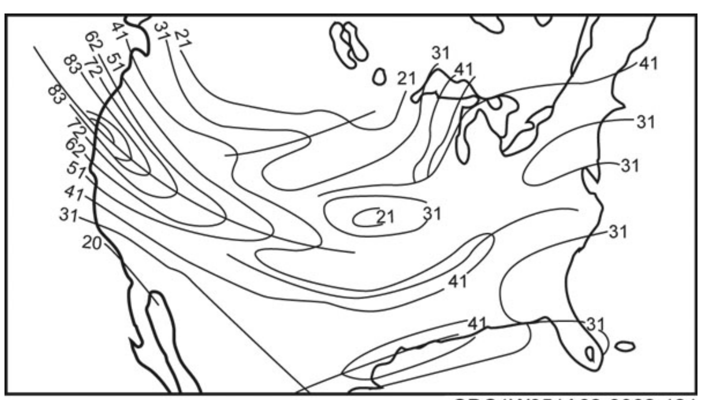

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Figure 1–39. Organized jet stream over western United States displaying jet fingers (disorganized jet) over central and eastern states.

The isotach maxima that occur along the axis of the jet stream are related to the short-wave troughs in the westerlies and their connected surface systems. The maxima usually move near the speed of these waves (average 35 to 40 knots). Distances between maxima vary from about 10 to 25 degrees longitude and speeds may reach 250 knots.

The maxima tend to be closed, symmetrical, lenticular-shaped isotachs that keep their shape over short periods. They undergo definite life cycles with periods of formation and decay, which often may be followed on synoptic products. Figure 1–40 illustrates the life cycle of a fully organized jet stream. Beginning with a balanced symmetrical arrangement of isotachs along the jet stream axis (fig. 1–40,A), the isotachs downstream from the maxima centers are displaced faster than those behind the center (fig. 1–40,B). In addition, high-speed isotachs are displaced faster downstream than slower ones. This increases the area enclosed by the highest valued isotach and packs the isotachs downstream from the maxima centers (fig. 1–40,C). The maxima merge with the low-speed area downstream and the low-speed area disappears (fig. 1–40,D). The resulting pattern leads back to the beginning stage (fig. 1–40,A), and the cycle is repeated. For a jet stream on the decline, the condition shown in figure 1–40, view C, is followed by that in figure 1–40, view E, rather than 1–40, view D. This stage is the first indication of the disintegration of the jet stream and its breakup into jet fingers.

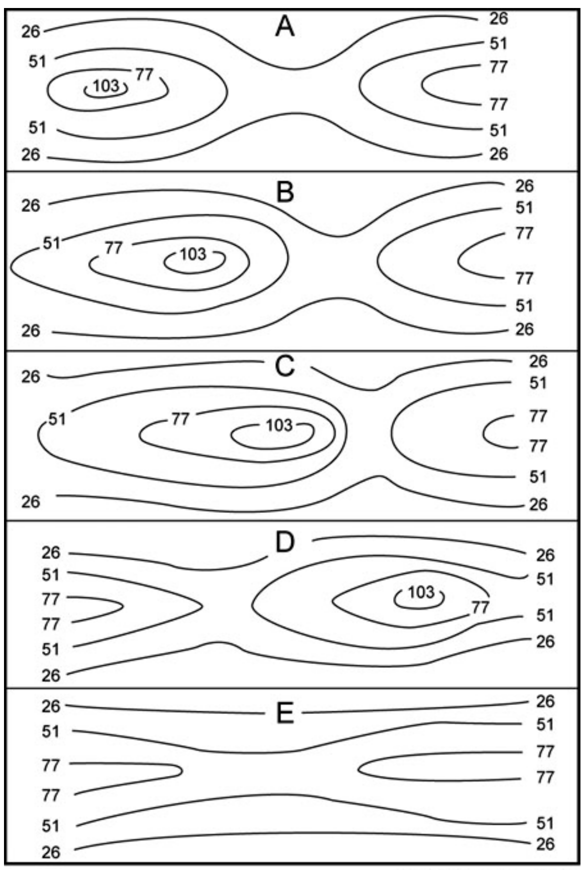

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Figure 1–40. Normal eastward progression of jet maxima and minima.

Occasionally, the jet stream surrounds the earth as a continuous band, but usually diminishes gradually at one or more points. It then reappears in a new location. This new appearance is often connected with migration of the jet stream. The jet stream can migrate due to changing orientation at speeds near 10 knots, but it does not (at any time) move perpendicular to its axis. This movement usually is to the north or south.

Basically, the jet stream follows the migration of the sun. Its mean position advances south in winter and retreats toward the poles during summer. The average latitude of the jet stream in the Northern Hemisphere is 35 to 40°N in winter and 42°N in summer.

In a jet migrating north, a west wind maximum usually emerges from the Tropics and gradually moves through the lower midlatitudes. Another maximum, initially located in the upper midlatitudes, advances toward the Arctic Circle and weakens. The jet stream is usually well organized and the contour pattern shows troughs extending into low latitudes. As the jet progresses northward, the amplitude of the long waves decreases and cut-off lows, south of the prevailing westerlies, dissipate.

The jet stream moves southward until the long-wave amplitude reaches a critical limit. At this stage, the jet usually disintegrates. Also, at this point, cut-off, closed lows frequently form in the southern portion of the troughs. This reestablishes a zonal flow now farther to the north and a new jet stream forms to start another cycle.

012. Jet stream and migratory pressure system relationships

Cyclones and bad weather tend to be associated with jet streams. This is especially true for weather associated with a warm front in the formative and mature stages of a cyclone.

Earlier, you learned that the midlatitude jet usually lies directly above the 500-mb position of the polar front. Using average frontal slopes, this places the jet stream about 300 miles behind the surface cold front and about 600 miles ahead of the surface warm or stationary front.

For instance, the following nine rules apply:

1. The jet stream is perpendicular to an occlusion and to a north-south oriented cold front with no associated warm front.
2. The jet stream remains north of an unoccluded wave cyclone.
3. The jet stream is south (near the point of occlusion) of the low associated with an occluded front.
4. In a series of lows of a cyclone family, each low is associated with a jet stream maximum.
5. Each cyclone associated with westerly flow aloft has an associated isotach maximum.
6. The jet stream parallels the direction of the warm sector isobars of a surface low.
7. The jet stream roughly parallels the isobars around the southern periphery of a cold (slow-moving) surface low.
8. The jet stream roughly parallels the isobars around the northern periphery of a warm (slow-moving) surface high.
9. When a cold, moving, polar high stagnates and begins to warm up, the jet stream usually changes markedly. As the thermal contrast in the southwest quadrant is destroyed and the warm-air advection moves northward, the original jet stream usually dissipates and a new jet forms to the north.

These four cloud patterns are usually associated with the jet stream:

- Lines of cirrus in bands.
- Patches of cirrocumulus or altocumulus castellanus.
- Waves of altocumulus.
- Lenticular clouds in waves.

These cloud patterns often extend from horizon to horizon and move very rapidly. Clouds occur most frequently in the following two positions:

- 10,000 to 15,000 feet below the jet core and 4 to 5° poleward.
- 5,000 to 10,000 feet below the jet core and 4 and 5° equatorward.

By far, most clouds are on the equatorward side of the jet. Most of the time, clouds aren’t observed above the jet core.

The greatest occurrence of precipitation usually straddles the jet axis, with a slight bias toward the poleward side.

Self-Test Questions

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

008. General structure of jet streams

1. Briefly define a jet stream.

2. How does the jet stream vary in latitude and altitude around the world?

3. Two major jet streams are frequently observed. What are they?

4. Of the horizontal or vertical wind shears associated with the jet stream, which is the stronger?

5. Where does the greatest vertical wind shear occur with respect to the jet core?

6. Where does the greatest horizontal shear occur with respect to the jet core?

7. Match the jet stream systems in column B with the characteristics in column A. Systems may be used once, more than once, or not at all.

Column A	*Column B*
____ (1) Jet stream found near the Arctic Circle.	a. Polar front jet stream.
____ (2) Observed in the latitudinal range of 25 to 30°N near the 150-mb level.	b. Subtropical jet stream.
____ (3) Jet stream which divides the tropopause into the polar and midlatitude leaves.	c. Arctic/Polar night jet stream.
____ (4) Summertime jet stream found over southern Asia and northern Africa at the 150-mb level.	d. Tropical easterly jet stream.
____ (5) The jet streams which most affect the Northern Hemisphere.	

8. What are the two primary causes of jet stream formation?

9. What's the primary originator of the PFJ stream?

009. Polar front jet stream

1. In relation to the 500-mb surface, where is the PFJ core usually found?

2. Along what isotherm does the PFJ intersect the 500-mb surface?

3. At what altitude do you usually find the jet stream core during the winter?

4. The width of the jet core is approximately equal to what isotherm ribbon?

010. Subtropical jet stream

1. What factor accelerates the STJ?

2. What might a wedged-shaped cloud pattern seen on satellite imagery over Texas indicate? What causes this pattern to occur?

011. Life cycle of a jet stream

1. For each of the statements below, indicate whether the jet would be organized or disorganized.

   a. The longitudinal axis tends to follow the long-wave pattern.

   b. The wind gradient along the axis becomes very strong (often exceeding 100 knots).

   c. Jet fingers are formed often and about 300 to 400 miles apart.

   d. Well-defined wind speeds of 50 knots or more disappear.

   e. The distance between isotach maxima is 10 to 25° longitude.

2. The following series of statements represent events during the cycle of a jet stream migrating northward. The statements are out of sequence. Arrange them in the proper sequence.

   _____ a. A west wind maximum emerges from the Tropics.

   _____ b. The jet stream is well-organized and often shows troughing into the low latitudes.

   _____ c. A second maximum, located in the upper midlatitudes, moves northward and dissipates.

   _____ d. The amplitude of the long waves decrease and a classical high index situation exists.

   _____ e. Cut-off lows south of the prevailing westerlies dissipate.

012. Jet stream and migratory pressure system relationships

1. How will the jet stream lie in relation to an occlusion and to a cold front oriented north-south, with no associated warm front?

2. Where does the jet stream remain (north or south) in relation to an unoccluded wave cyclone?

3. Does the jet stream lie north or south of the low associated with an occluded front?

4. Will the jet stream parallel the direction of the warm sector isobars of a surface low?

5. Does the jet stream roughly parallel the isobars around the northern periphery of a warm (slow-moving) surface high?

6. When a cold surface high dissipates, what happens to the jet stream aloft?

7. List the four cloud patterns associated with the jet stream.

8. Where do clouds most frequently occur in relation to the jet core and the equator?

9. Where do clouds most frequently occur in relation to the jet core and the poles?

10. Precipitation usually straddles the jet, with a slight bias toward which side of the jet?

1–4. Air Masses

In earlier sections, you learned about migratory pressure systems. Now, we'll explore the air masses that are an integral part of each migratory system. We'll begin by stating that a migratory pressure system is categorized by the air mass with which it's associated. For instance, the air mass that constitutes a low pressure system may be categorized as tropical or polar depending on its properties.

Air motions within the atmosphere are normally very complex. However, sometimes a large portion of the atmosphere comes to rest (stagnates) or moves slowly over a uniform surface. When this occurs, the air within this portion of the atmosphere assumes the same thermal and moisture properties as the surface beneath it. These properties tend to be uniform over the entire mass of air.

When the mass of air is set in motion, it maintains these characteristics until it's modified by the surfaces over which it moves. Therefore, you must know where these air masses form, how they're formed, how they're modified, and the flying conditions they produce. To provide this information, we'll explore five subject areas:

- Heat transfer.
- Air mass formation.
- Air mass structure.
- Air mass modification.
- Geographical distribution of air masses and their effect on weather.

013. Heat transfer

One of the thermodynamic laws that control the behavior of heat in our universe is that heat will always flow from hot to cold—as long as there is a temperature difference between them. But, how is this energy transfer process accomplished? In the atmosphere, heat is transferred in four ways:

1. Radiation.
2. Conduction.
3. Convection.
4. Advection.

We'll now examine each of these mechanisms individually to get an understanding of how they affect our weather.

How radiation contributes to heat transfer

Perhaps you've had the opportunity to stand in front of a huge bonfire on a bitterly cold evening. If you have, you've probably noticed how warm and flushed your face feels while the surrounding air remains quite cold? Somehow, energy from the fire is being transferred through the air with little effect upon the air itself. In addition, your face is absorbing this energy and converting it to heat energy. Thus, you feel warm.

This energy is known as radiant energy, or radiation. It travels in the form of waves that release energy when they're absorbed by an object. These waves have magnetic and electrical properties, which is why we call them electromagnetic waves.

Electromagnetic waves don't need a medium through which to travel. An example of a medium might be a telephone line or cable. Electromagnetic waves can even travel through a vacuum such as outer space. This can be proven by the fact that electromagnetic waves are emitted from the Sun, travel through space, and are felt here on earth. These electromagnetic waves also travel very fast (approximately 300,000km per second). Let's now explore some characteristics of electromagnetic waves.

NOTE: For the remainder of this text, we may refer to electromagnetic waves with the words "waves" and "wavelength."

Wave characteristics

Wave characteristics include low points or valleys (troughs) and high points or crests (ridges). By definition, wavelength is the distance between two points of corresponding phase in consecutive cycles. Figure 1–41 depicts two crests that are of corresponding phase and that are occurring in a consecutive cycle. We measure wavelengths from crest to crest or trough to trough in terms of micrometers (μm), which we symbolize with the Greek symbol lambda (λ). Now that we've discussed some characteristics of wavelength, let's see how they apply to radiation and temperature.

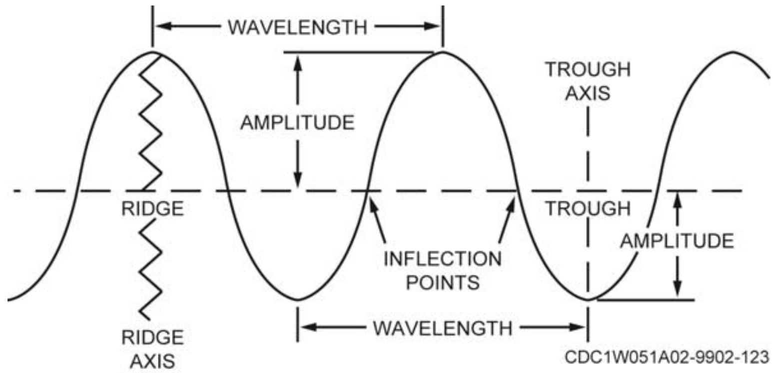

Figure 1–41. Electromagnetic wave characteristics.

Radiation and temperature

All things, no matter how large or small, emit some form of electromagnetic radiation in wavelengths. The temperature of an object primarily decides the object's wavelength. The higher the temperature, the faster the associated electrons vibrate and the shorter the wavelength emission.

By completing the following experiment, you can visualize this theory more clearly. Attach a 10-foot piece of rope to a pole. If you shake the free end of the rope rapidly (simulating high temperature and fast electron vibration), many "shorter" waves travel or propagate along the rope. Conversely, if you shake the free end of the rope slowly (simulating lower temperatures and slow electron vibration), "longer" waves appear and propagate on the rope. This experiment supports the fact that the higher the temperature of an object, the shorter is the wavelength and vice versa.

Two scientific laws explain the relationship of radiation, temperature, and energy emitted. They are:

1. Stefan-Boltzman law.
2. Wein's law.

Stefan-Boltzman law

Stefan-Boltzman law states as the temperature of an object increases, the object emits more total radiation. The sun is approximately 20 times warmer than the earth (using Kelvin degrees).

Here's a question for you. How much more solar radiation will the sun emit than the earth? Answer: The amount of radiation emitted is proportional (comparative relation in amount) to the fourth power of the temperature. In other words, if the sun is twenty times warmer than the earth, it will emit 20^4 times more radiation.

Wein's law

Wein's law states that the wavelength at which the maximum amount of energy is emitted by an object is inversely (directly opposite) proportional to the temperature of the object. Simply put, where the temperature of the object is high, such as the sun, the wavelength is shorter. Also, where the temperature of the object is cooler, such as the earth, the wavelength is longer. The sun emits a great deal more energy than the earth; its temperature is approximately 6,000°K. The wavelength of the sun's emitted energy is shorter than that of the earth's emitted energy.

Long-wave/short-wave radiation

Because the earth emits longer wavelengths, its emission is called *long-wave radiation*. It's also referred to as *terrestrial radiation*, meaning, "from the earth." Conversely, the sun's shorter wavelength emission is often called *short-wave radiation.* Radiation from the sun is also referred to as *solar radiation,* meaning, "from the sun." You know that the sun radiates at a maximum rate at one wavelength; however, it also emits some radiation at almost all other wavelengths. When looking at the emission of radiation at each wavelength, you can obtain the sun's electromagnetic spectrum or electromagnetic wavelength range. We'll now briefly examine the sun's electromagnetic spectrum.

Electromagnetic spectrum

The three areas of the sun’s electromagnetic spectrum we’ll discuss are as follows:

1. Visible.
2. Ultraviolet.
3. Infrared.

The visible spectrum is obviously the spectrum visible to the naked eye and is comprised of the colors violet to red. The range of this spectrum is between 0.4 and 0.74 µm (fig. 1–42.) Violet corresponds to the shortest of these wavelengths (0.4 µm) with red corresponding to the longer of these wavelengths (0.7 µm). The sun emits nearly 44 percent of its radiation in the visible spectrum.

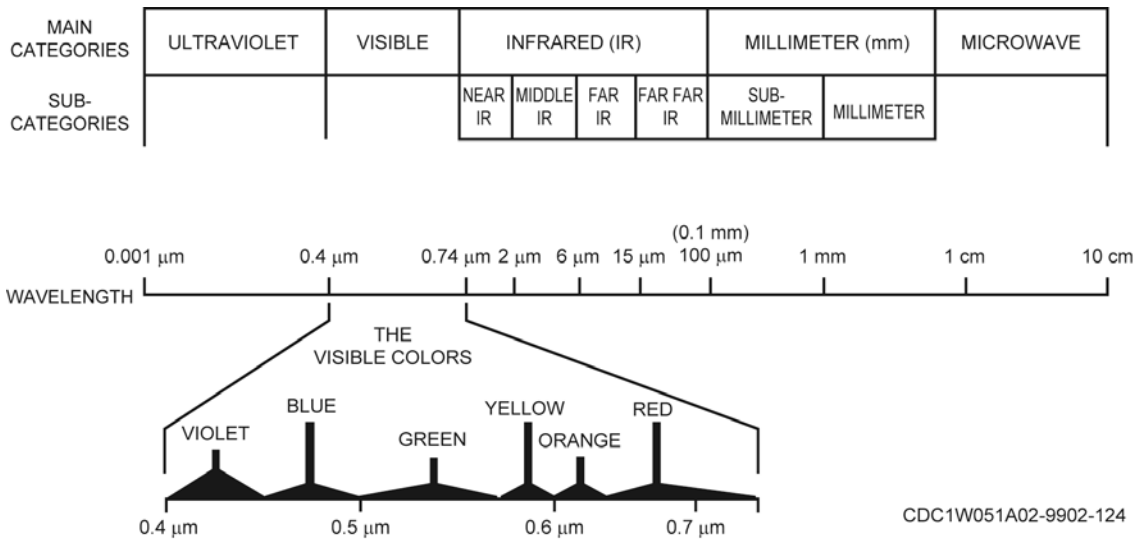

Figure 1–42. Electromagnetic spectrum.

Wavelengths that are shorter than the visible spectrum (< 0.4 µm) are known as *ultraviolet* (fig. 1–42). X-rays and gamma rays fall into this category. Seven percent of the sun’s total energy is emitted at the ultraviolet region.

Wavelengths longer than 0.7 µm are called infrared. We use sensors, like those on weather satellites, to detect these wavelengths as the infrared spectrum can’t be seen by humans. Since clouds always radiate infrared energy, we can process satellite imagery using infrared sensors during both day and night. This nighttime coverage is an invaluable tool for the forecaster, especially during nocturnal severe weather events. Roughly 37 percent of the sun’s energy is radiated in the near-infrared 0.7 to 1.5 µm range.

Conversely, the relatively cool earth emits almost all of its energy as far-infrared, 4 to 25 µm. This is understandable since infrared wavelengths are longer, which would be associated with cooler temperatures. Remember, part of Wien’s law states: Where the temperature of an object is cooler, the wavelength is longer.

With these thoughts in mind, let’s look at some other factors associated with radiation that help explain some of earth’s atmospheric and weather peculiarities. These factors include absorption, emission, and albedo.

Absorption, emission, and albedo

All objects not only radiate energy, they absorb it as well. Absorption is the process by which incident radiant energy is retained by a substance. This explains why the earth doesn’t get progressively colder through the emission of long-wave radiation because it’s absorbing radiation simultaneously. If the earth radiates more energy than it absorbs, it gets colder. An example of this is a clear night. If the earth absorbs more energy than it emits, it gets warmer. An example of this is a clear sunny day. When an object emits and absorbs at the same rate, its temperature remains constant. Three important factors come into play when dealing with radiation:

- Black body.
- Radiative equilibrium temperature.
- Kirchoff's law.

Black body

Temperature, moisture, texture, and color are surface characteristics that strongly affect the rate of absorption and radiation. Let's examine the color characteristic and how it affects absorption and radiation. Have you ever walked barefoot on a black asphalt road on a hot sunny day? The asphalt obviously gets very hot—much hotter than surrounding surfaces such as dirt and grass. This is because the black asphalt is a good absorber of short-wave radiation. The asphalt converts energy from the sun into internal energy, usually increasing its temperature.

Good absorbers are also good emitters of radiation. Let's go back to the asphalt example. Once the sun sets, the asphalt cools quickly and is usually cooler than the surrounding dirt and grass by the time morning arrives. Objects that are perfect absorbers (absorb all the radiation that strikes them) and perfect emitters (emits all possible radiation) are called black bodies and are considered ideal objects. The earth and sun are assumed to be black bodies.

Radiative equilibrium temperature

If we say that the earth and sun are black bodies and therefore absorb and emit radiation equally, a state of radiative equilibrium is achieved. The average at which this occurs is called the radiative equilibrium temperature. Because of the earth's distance from the sun (93 million miles), its radiative equilibrium temperature is about –4°F! This is much lower than the observed average temperature on earth of 59°F. What could cause this difference?

One factor that we haven't discussed yet is the earth's atmosphere and its ability to absorb and emit radiation. Unlike the earth and sun, the atmosphere doesn't behave like a black body. Instead, it is considered a selective absorber because it selectively absorbs and emits radiation. In other words, it's a good absorber at certain wavelengths but may not be a good absorber of all wavelengths. There's a law in physics that speaks specifically about this phenomenon. It's called Kirchoff's law.

Kirchoff's law

Kirchoff's law says that good absorbers of a certain wavelength are good emitters at that wavelength. Some gases in the atmosphere are selective absorbers. The ozone is a good example of this. We've all heard of ozone depletion due to added chlorofluorocarbons (CFC) in the atmosphere. The reason why the ozone is so important is that it protects the earth from harmful ultraviolet radiation because it selectively absorbs this wavelength.

Carbon dioxide (CO_2) and water vapor are also both selective absorbers and emitters. These gases are abundant in the lower atmosphere of the earth. As the earth radiates its energy at far-infrared wavelengths, CO_2 and water vapor absorb a large portion of this radiation. This absorption increases the kinetic energy (energy of motion) of these gases. These gases (CO_2 and water vapor) collide with oxygen and nitrogen molecules that increase the average kinetic energy of the air. The net result of this absorption is that the lower atmosphere warms.

As CO_2 and water vapor selectively absorb infrared radiation, they also selectively emit infrared radiation in all directions. Some of this energy is radiated back toward the earth's surface where it's absorbed and heats the ground. At this point, the process of the earth radiating infrared wavelengths continues.

This cycle is popularly known as the *greenhouse effect*. The addition of cloud cover enhances this effect as the associated tiny cloud droplets are also selective absorbers of infrared wavelengths. This is why calm, cloudy nights are usually warmer than calm, clear nights whereas calm, cloudy days are cooler than calm, sunny days. If these selective absorbers weren't present in the atmosphere, earth's mean radiative equilibrium temperature would be closer to the –4°F stated earlier!

We've just shown that CO_2 and water vapor are both good absorbers and emitters of the infrared wavelength. By doing so, we've proven Kirchoff's law.

Conduction

The second method of heat transfer is conduction. This process involves the transfer of energy by molecular motion. During this process, the transfer of heat is always from hot to cold. Two important factors are involved in conduction:

1. Heat conductivity.
2. Atmospheric significance.

Heat conductivity

Heat conductivity is the ability of a substance to conduct heat. It's related to the molecular structure (heat capacity) and density of a substance. Air is a poor conductor of heat. In contrast, soil conducts heat approximately 10 times better than air. Furthermore, water conducts heat approximately 100 times better than air. Because air is such a poor conductor of heat, it's important as a heat transfer process only in the molecular boundary layer (i.e., the lower levels of the atmosphere). This brings the question, what's so important about a process that only affects a small area of the atmosphere? Let's look at this question from the point of atmospheric significance.

Atmospheric significance

You already know that the surface of the earth heats and cools by the absorption and emission of solar radiation. Conduction links the surface to the air directly above it. This in turn influences low-level stability and air mass modification. Conduction is important in transferring heat between the surface and the atmosphere, but as we stated previously, only affects the molecular boundary layer. Other mechanisms that we discuss, such as the processes of convection and advection, must act along with conduction to transport heat to and from other places of the atmosphere.

Convection

Convection is the process that vertically transports the atmospheric properties of heat and moisture. This is a much more efficient method of heat transfer than conduction. It's associated with upward vertical motions that can be initiated by any individual or combination of the following:

- Surface heating.
- Low-level convergence.
- Orographic effects.
- Frontal lift.

The vertical extent of convection is influenced by the stability of the atmosphere. A stable atmosphere suppresses convection while an unstable atmosphere enhances convection. Convection occurs in the troposphere as the stable stratosphere acts to "cap" convection.

Advection

Advection is the last of the heat transfer processes that we discuss. It's an important type of heat transfer mechanism because it involves the horizontal transport of atmospheric properties such as heat by the wind. Sometimes referred to as *temperature advection*, you can evaluate its strength by examining the direction and speed of the wind flow. The more closely spaced the isobars/contours, the stronger is the pressure/contour gradient and, therefore, the stronger is advection.

014. Air mass formation

Three factors are considered necessary for the formation of air masses. They are as follows:

1. There must be a surface with comparatively uniform properties, especially the properties of temperature and moisture. This surface may be water or land.
2. The air must stagnate over the uniform surface.
3. There must be a large divergent flow as shown in figure 1–43. This flow inhibits temperature contrasts and produces uniform characteristics throughout the air mass.

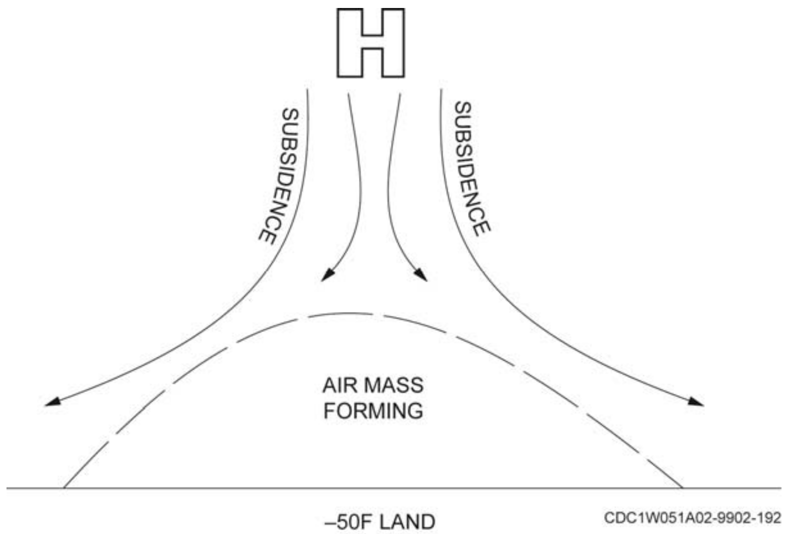

Figure 1–43. Air mass formation.

The stagnation of air over a surface of uniform conditions permits a condition of temperature equilibrium between the ground and air, actually the formation of the air mass. This is accomplished by these three processes:

1. Turbulent-convective transport of heat upward into the higher levels of the air.
2. Cooling of air by radiation loss of heat.
3. Transport of heat by evaporation and condensation processes.

By far the most effective and fastest transfer is the turbulent-convective transport of heat upwards. The least efficient and slowest is radiation. The equilibrium layer created by the turbulent-convective process has the greatest vertical depth, but layers resulting from radiation cooling are comparatively shallow. Evaporation and condensation contribute in each of the other two processes, but are very ineffective in creating air masses by themselves.

Anticyclonic systems are excellent for air mass formation. They're usually stagnant or slow moving, which allows the air time to adjust its heat and moisture content to that of the underlying surface. Anticyclonic systems have a divergent airflow, which spreads the properties horizontally over a large area, with turbulence and convection distributing these properties vertically. These factors have the effect of creating uniformity throughout the air mass. Examples of the anticyclonic systems are as follows:

- Bermuda warm high.
- Pacific warm high.
- Siberian cold high.

Due to higher thickness aloft, warm highs extend to great heights. They also produce an air mass of comparatively large vertical extent. In contrast, cold highs have a moderate or shallow vertical extent and thereby produce air masses of moderate or shallow height.

Cyclonic systems aren't conducive to air mass formation. Instead, they are characterized by strong wind speeds that prevent the air mass from remaining over a surface long enough to become modified. Cyclonic systems also have a convergent windflow that creates temperature contrasts. The convergence creates ascending air, which causes the air to lose contact with the surface. Without this contact, the air can't acquire the properties of the surface.

NOTE: The world's *belts of convergence* resemble the cyclonic systems. They also aren't conducive to air mass formation.

Certain characteristics decide the types of air masses forming in various areas of the world (source regions). Tropic regions are covered predominantly by water surfaces, thus providing the uniform surface. In contrast, air masses in the polar regions form slowly through the loss of heat by radiation. The surface of the polar region is almost completely covered by snow, thus providing the needed uniform surface.

The temperate latitudes provide neither of the two factors needed for air mass formation and rarely serve as source regions. Steep gradients of temperature and a large variance in the land/water distribution prevail in temperate latitudes. Instead of the temperate zone being a source region, the temperate zone is a transition zone. Here, air masses with different properties move in, clash, and produce a variety of weather.

There are two notable exceptions where air masses form despite the lack of proper factors:

1. Air masses that form maritime polar air over the North Pacific between Siberia and North America.
2. Over the Atlantic water off Labrador and Newfoundland.

Both areas have active and strong windflow, yet the air masses still form. This formation, however, is related to the characteristics of the region rather than the system.

015. Air mass structure

The *Glossary of Meteorology* defines an *air mass* as a widespread body of air having properties identified by its horizontal composition—particularly temperature and humidity. Also, vertical temperature and moisture variations are about the same throughout the horizontal extent of a given air mass. Air masses should *not* be classified as a part of the secondary circulation; however, air mass properties, particularly along the boundary zone between contrasting air masses, cause circulation phenomena that are secondary in scale. This is the way we examine air mass properties.

If atmospheric circulation is strictly zonal, you'll find only two different air masses:

1. Polar.
2. Tropical.

The boundary, or zone, between them is the polar front. This distribution is clear-cut in the middle and upper troposphere.

At the surface, perturbations in the zonal flow confuse the picture. Horizontal turbulence, variations in insolation, and the motion of the air masses themselves invoke complicated transitional effects.

The considerable differences in the earth-atmosphere parameters influence the different sections of the atmosphere. Radiation rates, convection, evaporation, condensation, insolation, and turbulence are all functions of the earth's surface over which a given air mass may spend a large part of its time. Air masses over oceans acquire maritime characteristics such as mild temperatures, considerable moisture (at least in the lower layers), moderate lapse rates, and a degree of conditional instability. Polar air masses tend to be very cold and dry with temperature inversions in the lower levels and fairly uniform lapse rates aloft. These examples serve to show that we should examine some source regions of the principal air masses. This examination will enable you to better understand their characteristics.

Air masses are classified with letter identifiers according to their source region and characteristics. The source region is considered the most useful criterion. Thus, the primary air mass identifier refers to the source region. For example:

- A capital *P* represents a polar air mass.
- A capital *T* represents a tropical air mass.
- On occasion arctic air mass properties are important; a capital *A* identifies these air masses.
- Equatorial air masses are represented by a capital *E*.

To further classify the source region, we make a distinction between land and water areas since these regions give decidedly different characteristics to overlying air masses. Use the letter "c" for continental or "m" for maritime surfaces; for example, cP indicates a continental polar air mass and mT a maritime tropical air mass.

The stability characteristics of air masses often depend on the temperature differences between them and the surfaces over which they may travel; thus, a third category identifies whether the air is colder (k) or warmer (w) than the surface it's moving over. Therefore, "w" designates those that are warmer and "k" those that are colder than the surfaces over which they're found. This classification is relative and frequently difficult to determine. The several possibilities can be listed are as follows:

mPw	mPk	CPk	cA
mTw	mTk	CTw	E

We can explain the system as follows. mPw represents a polar maritime air mass that's warmer than the surface over which it moves. cPk describes a continental polar air mass that's colder than the surface over which it moves, and so on. We don't need all the categories for arctic air masses; instead, they're usually continental in character and colder than the surface over which they move.

Continental and maritime designations signify the influence of the surface on air mass properties. Arctic, polar, and tropical primary designations suggest the importance of latitude of the source regions.

Typical k air masses have convective clouds, turbulent gusty winds, and good visibility. Heating of these air masses, by a warmer surface, increases the lapse rate and vertical eddies that favor convective transport.

In contrast, w air masses have their lower layers cooled by the colder surface. This tendency toward an inversion of the lapse rate increases stability. When present cloud systems are stratiform and vertical turbulence is suppressed, visibility is frequently poor with low-level radiation and advection fog the rule.

However, a word of caution is in order. The interaction of air masses with the surface isn't the only mechanism that produces observed weather effects. Convergence, divergence, and subsidence are very important in finding lapse rates and stability. In addition, dynamical processes frequently dominate the thermal effects.

We use a fourth letter to represent the stability of the air mass. An "s" signifies a stable air mass and "u" specifies an unstable air mass.

By combining these letters, we can obtain a relatively complete description of the air masses. A classification of cPws signifies a stable, continental polar air mass that's warmer than the surface it's moving over. We'll now turn our attention to some discussion of specific air masses. We'll limit our emphasis to those air masses and source regions that influence the Northern Hemisphere weather. They are as follows:

- Wintertime continental polar (cP) air masses.
- Summertime continental polar (cP) air masses.
- Wintertime maritime polar (mP) air masses.
- Summertime maritime polar (mP) air masses.

- Wintertime maritime tropical (mT) air masses.
- Summertime maritime tropical (mT) air masses.
- Continental tropical (cT) air masses.

Figure 1–44 shows the air mass and source regions for January. In contrast, figure 1–45 shows the air mass and source regions for July.

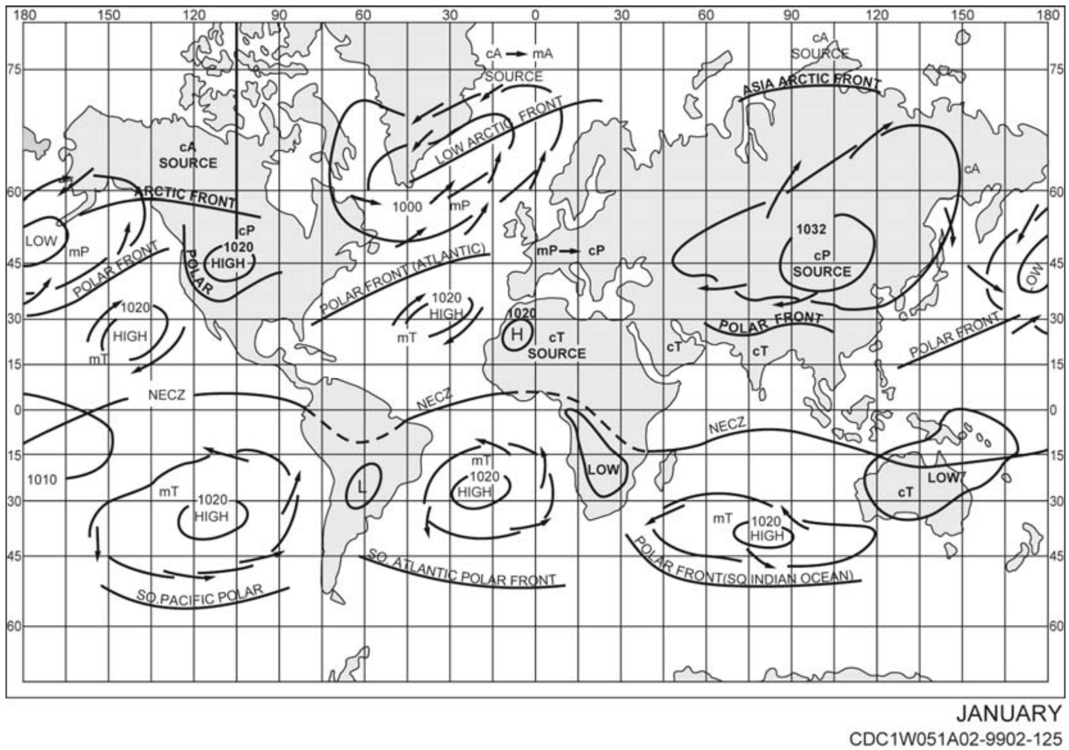

Figure 1–44. Fronts, pressure centers, and air mass source regions (January).

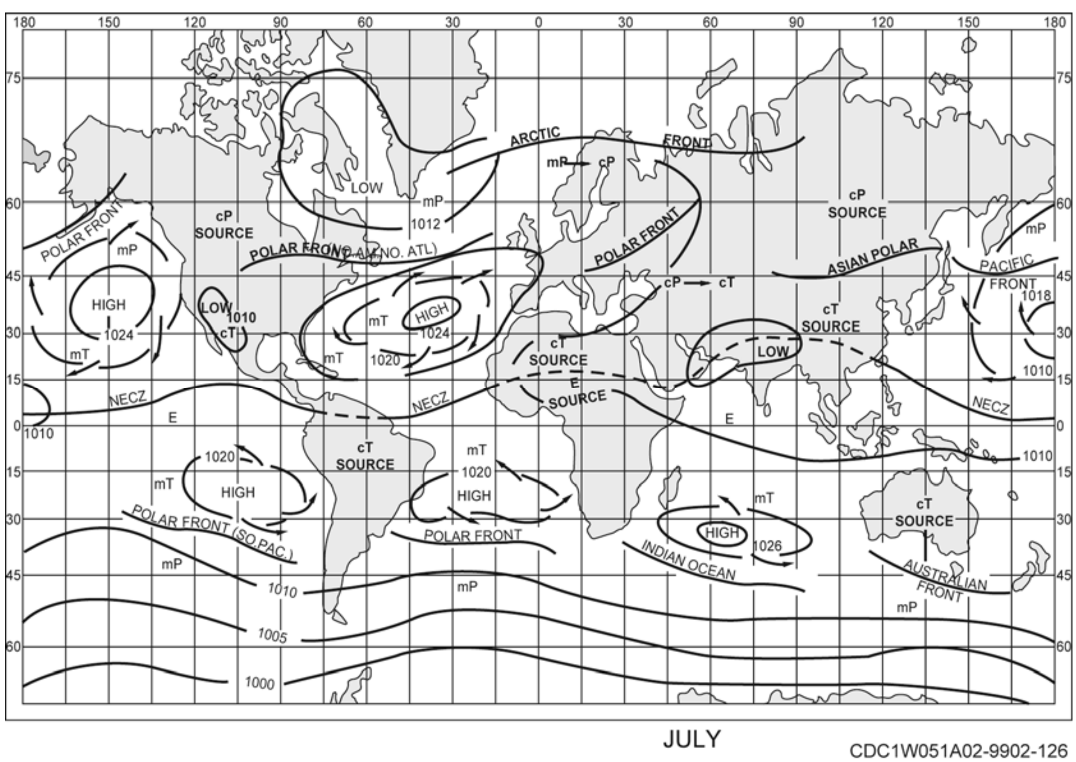

Figure 1–45. Fronts, pressure centers, and air mass source regions (July).

Wintertime cP air masses

These air masses have their source regions in central Canada and Siberia. Next to the extreme Arctic and Antarctic regions, these are the coldest places on earth. This is because of the intense radiational cooling and lack of insolational heating in these source regions. The surface is completely frozen and is mostly snow and ice covered. Air masses in these regions are extremely cold and stable and are, therefore, very dry. Clouds are uncommon in the cP air masses, although ice fogs frequently form when temperatures fall below –40°F.

Summertime cP air masses

These air masses have their source regions in the central portion of high-latitude continents. Insolational heating warms the surface substantially, even to the point of dissipating the snow cover and thawing the ground to a noticeable depth. Summertime cP air masses are cool and dry, but not necessarily stable. They're fundamentally changed forms of the wintertime cP air masses that have experienced insolational heating in their lower layers. Generally, their lapse rates are less steep than their wintertime counterparts.

Wintertime mP air masses

These air masses are found over open oceans at high latitudes. The cool, moist air masses indigenous to this locale are mainly cP air masses that have moved out over the open ocean. The lower layers are substantially changed by the comparatively warmer, yet still cold, water surface. The mP air masses tend to be moist and unstable in the lower layers, and cold and dry aloft.

Summertime mP air masses

These air masses start in basically the same regions as the wintertime mP air masses. For the North American continent, these regions are the Gulf of Alaska and the North Atlantic. Summertime mP air masses resemble their wintertime counterpart; that is, they're cool and moist in the lower layers, and cool and dry aloft. The overall temperature is somewhat higher than in the winter mP air masses. Instability is more general in the lower layers and abrupt change in the lapse rates is found in the moisture discontinuity aloft. An inversion at this level isn't uncommon.

Wintertime mT air masses

These air masses are found over the open oceans, just south of the polar front, near 30°N latitude. In fact, mT air mass source regions extend southward across the near-equatorial trade wind convergence zone (NETWCZ) into the Southern Hemisphere. In North America, the source regions are the semipermanent subtropical high-pressure centers in the southwestern Caribbean. The mT air masses in winter are warm, moist, and unstable. In the lower levels, the lapse rates frequently approach the dry adiabatic rate, and the lapse rates are usually steep up to the tropopause. Moisture is fairly well diffused to comparatively high levels.

Summertime mT air masses

These have the same source region as the wintertime mT air masses. Additionally, they, originate in the Caribbean Sea. Those that influence North American weather originate in the semipermanent anticyclones. In the summer season, the mT masses are centered near 15°N latitude. In the summer, mT air masses are also very warm and moist, and markedly unstable. Convective instability frequently occurs.

cT air masses

These air masses are relevant to North America only during the summer. The source region is confined to Northern Mexico and the extreme southwest deserts of the United States and generally centered near 30ºN latitude. The cT air masses are hot, dry, and very unstable. Because of this inherent instability, large-scale upward vertical motions occur. This in turn causes surface pressures to fall, forming weak surface circulations called *heat lows*. The upper levels (usually 700mb and above) are dominated by subsiding air associated with the convergence between the Hadley and the Farrel cells.

016. Air mass modification

If the various types of air masses remained permanently in their source regions and vertical motions were nonexistent, there would be little or no change in weather. However, air masses migrate. Because of this, definite changes in their structure (called *air mass modifications)* are a natural result. When an air mass moves out of its source region, several factors act on the air mass to change its properties. These changes are as follows:

- Advection.
- Radiation losses.
- Thermodynamic modification.
- Turbulent mixing.

NOTE: Thermodynamics and turbulent (mechanical) mixing is based on the process causing the changes.

The table below lists the major processes and also shows the changes in stability you should expect with these processes.

The Process	How it Happens	Results
1. Heating from below.	Air mass passes from a cold surface to a warm surface, or surface under air mass is heated by sun.	Decreases stability.
2. Cooling from below.	Air mass passes from over a warm surface to a cold surface, or radiational cooling of surface under the air mass takes place.	Increases stability.
3. Addition of moisture.	By evaporation from water, ice, or snow surfaces; moist ground; raindrops; or other precipitation that falls from overrunning saturated air currents.	Decreases stability.
4. Removal of moisture.	By condensation and precipitation from the air mass.	Increases stability.
5. Turbulent mixing.	Up drafts and down drafts.	Tends to result in a thorough mixing of the layer where turbulence exists.
6. Sinking.	Movement down from above colder air masses or descent from high elevations to lowlands, subsidence, and lateral spreading.	Increases stability.
7. Lifting.	Movement up over colder air masses, over elevations of land, or to compensate for air converging at the same level.	Decreases stability.

Advection

As air masses move away from their source regions they carry with them physical characteristics acquired from their long residence in the region. We can describe these initial conditions of vertical temperature profiles (lapse rates) using other terms such as:

- Moisture content and its distribution.
- Mean temperature.
- Vorticity.
- Wind velocity.

Most of the variations in the parameters are found in the vertical distributions.

Air masses are strikingly homogeneous throughout their horizontal extent. The horizontal motion of the air mass doesn't directly affect changes in temperature and moisture content or lapse rates. This is exclusive of the equatorial regions. However, the climatic characteristics of the terrestrial regions over which the air masses pass can and do change them. An example is illustrated in figure 1–46.

Modifications include such factors as subsidence or uplift, changes in radiation rates, addition of moisture, or the loss of it. Also, if vertical wind shear characterizes the original structure of the air mass, advection may significantly alter the vertical structure. For example, if the winds aloft cause a decrease in the upper-level temperature with time, the overall stability of the air mass decreases. Thus, the frequency of the incidence of convective motion may increase. Changes of this type are associated with the thermal wind.

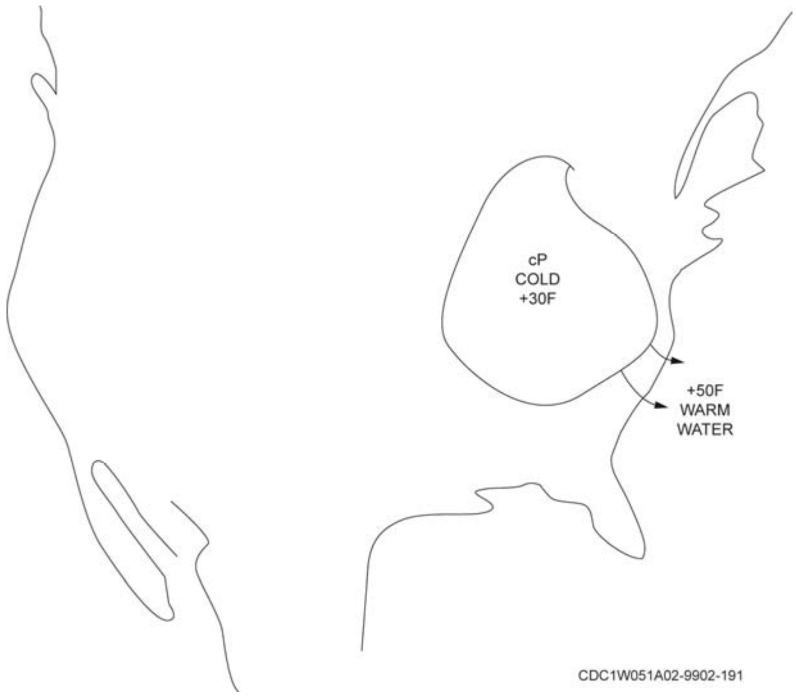

Figure 1–46. Advection of an air mass.

Radiation losses

Subsidence of discrete layers of the air mass often creates inversions in temperature, both on the ground and at intermediate levels aloft. Heat losses through radiation out to space inhibits these inversions locally. In contrast, rising air, moving upwards as the result of mechanical or convective stimulus, usually destroys previously established inversion layers by increasing the incidence of local radiative losses. In either case, temperature changes in air masses resulting from radiative exchange proceed very slowly. Thermodynamic processes may, and often do, affect more rapid temperature changes when turbulent mixing extends to great heights.

Radiation and advective changes are nonadiabatic, but, if the air ascends or descends while being advected, comparatively rapid adiabatic temperature changes are of controlling importance.

Thermodynamic modification

Thermodynamic processes, other than adiabatic heating or cooling that cause *air mass* modification, include such effects as heating from below (diabatics). This warming increases lapse rate and instability in the air mass. The warmed air rises and significantly alters the characteristics of the air mass. This concept is shown in figure 1–47.

If the air mass is warmer than the surface, the air cools from below and an inversion occurs, increasing the stability. One air mass that can't be modified through surface heating is the cT. Surface heating is the mechanism that forms the air mass. The only modification possible is when the air mass begins to move away from its source region; its new classification becomes cTw (warmer than the underlying surface). Therefore, the modification cTk isn't possible.

Changes in *air mass* stability indicate the cloud types that will form and the type of precipitation that could occur. Stability changes also indicate that you should expect lower-layer turbulence and visibility.

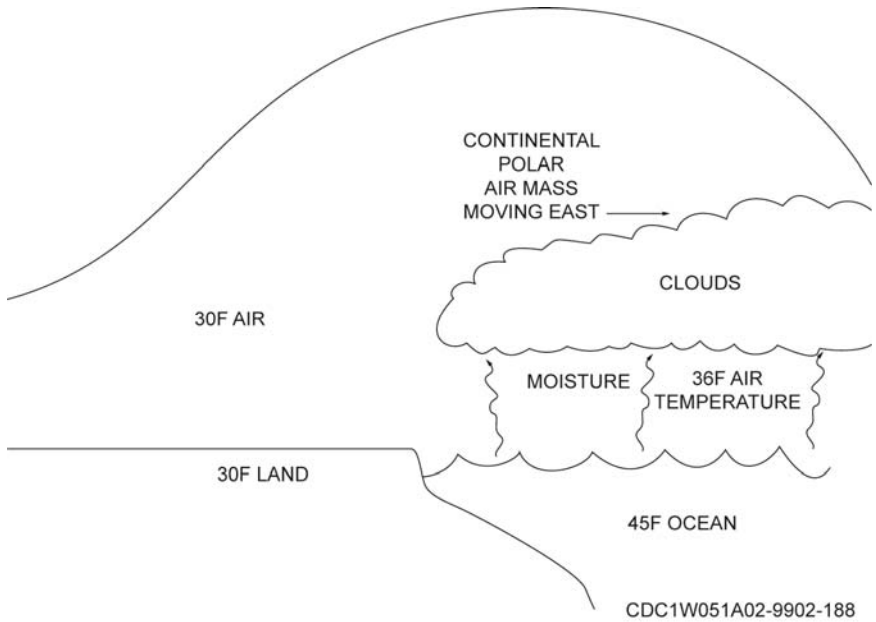

Figure 1–47. Air mass modification.

Nocturnal radiation is a common cause of modification. Radiation from the surface or the top of a low-level cloud layer cools the air immediately above the radiating surface. This decreases the lapse rate in the cooled air layer and increases the chances for condensation. The cooling above the cloud layer may create an unstable condition of cold air above the warm air that lies below the clouds. This is favorable for nocturnal thunderstorm development.

Air mass moisture content may be modified either by the addition of moisture by evaporation from a surface water source or by the removal of moisture by condensation and precipitation. Generally, when an air mass moves over a water surface, the moisture content of the lower layers increases. The stability of the air mass and the convective currents within it determine how high the moisture is carried and how much the air mass changes.

Turbulent mixing

Convective turbulence, a thermo-dynamically induced effect, and mechanically-induced turbulence (frictional effects) at the surface may result in thorough mixing, often to considerable heights. The result of these modification mechanisms changes the character of the air mass, either in discrete layers or throughout. However, once you identify an air mass from its source region characteristics, later modification often can be predicted, or at least expected.

As the air mass moves across uneven terrain and open water, turbulent mixing can occur in the lower levels. The mixing causes the temperature lapse rate in the turbulent layer to approach the dry adiabatic. This gives a decrease in temperature at the top of the mixed layer and an increase at the surface. Moisture also mixes throughout the layer. When the mixing is complete, the mixing ratio for all the points in the layer is equal, as shown in figure 1–48.

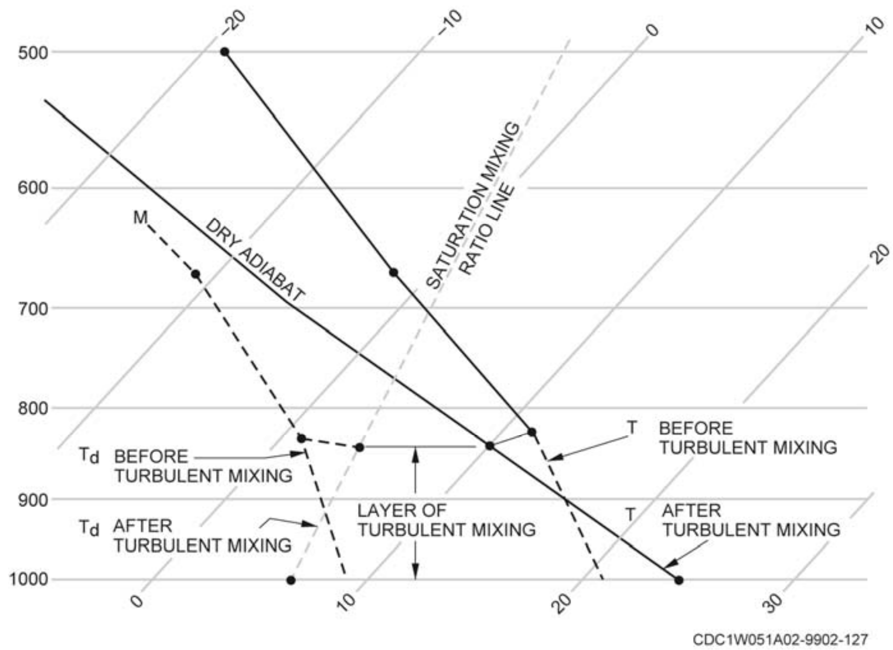

Figure 1–48. Sounding with turbulent mixing.

The example shows the effects of turbulent mixing in the simplest case (no cloud formation, convergence, evaporation, radiation, or advection effects). When the air is well supplied with moisture, the dew-point curve on the Skew-T shows only a few degrees of separation from the temperature curve. Turbulent mixing brings the air to saturation at some elevation above the ground, with clouds forming throughout the saturated layer.

Cumulus clouds are usually formed by turbulent mixing. When heating takes place in the layer near the ground, it’s often the triggering action that pushes the cumulus through the inversion, resulting in thunderstorm development. A check of the stability and the maximum temperature forecast usually reveals whether the thunderstorms will occur. However, the turbulent mixing in combination with normal heating often overrides a stable condition in the atmosphere and result in showers and thunderstorms.

Despite the significance of turbulence, the degree of air mass modification produced by this mechanism isn’t as significant as extensive horizontal convergence or divergence. These horizontal motions set up ascending and descending currents with their resulting effects upon stability. Divergence near the surface is associated with subsidence, dynamic warming, and increasing stability. In contrast, convergence near the surface is associated with ascending currents, a steepening lapse rate, and decreasing stability. These horizontal and other motions produce modifying effects on air masses that deserve our attention.

When you exercise reasonable care and good judgment in air mass analysis, you can track the movement of air masses over large distances for several days with considerable reliability.

General air mass modification

Now, we’ll investigate the modification of several air masses as they move away from their source regions and interact with the different surface regimes. More specifically, we’ll look at the following:

- Winter cP.
- Summertime cP.
- Wintertime mP.
- Summertime mP.
- Wintertime mT.
- Summertime mT.

Winter cP

Winter cP air masses, upon moving out, almost invariably become cPk because the surfaces in their paths are warmer than those of their source regions. Warming, with its accompanying decrease in stability, characterizes the lower levels, with turbulence steadily penetrating higher levels. Cumulus and stratocumulus clouds are features of the migration of these systems. The cPk air that moves over water surfaces, such as the Great Lakes, frequently brings heavy snow showers—called *lake-effect snows*—to areas on the south shores. The cPk air masses that reach the Gulf of Mexico are only slightly modified and produce strong instability. Violent squalls and thunderstorms are frequently the rule.

Summertime cP

The cP air that moves into the United States in the summer has much different properties than its winter counterpart. The source region remains the same, but it's now characterized by long summer days that melt snow and warm the land. The air is only moderately cool and surface evaporation adds water vapor to the air. A summertime cP air mass usually brings relief from the oppressive heat in the central and eastern states. This happens as cooler air lowers the air temperature to more comfortable levels. Daytime heating warms the lower layers, producing surface instability. With its added moisture, the rising air may condense and create a sky dotted with cumulus humulis (fair weather cumulus). Summertime cP air follows a modification similar to that of the wintertime cP air masses with the following exception—cP air that moves over the open ocean becomes cPw. What happens is that stability quickly increases; haze, fog, and low stratus clouds may appear after a few days.

Wintertime mP

Wintertime mP air becomes mPk upon reaching a coast, where instability develops in the lower layers. Showery, squally weather occurs when the air ascends over coastal mountains. As these air masses cross the Sierra Nevada or Cascade ranges, active cumulonimbus clouds, snow showers, and severe aircraft icing conditions are the rule. On moving farther inland, mPk air becomes mPw. Contact with the surface quickly cools the lowest layers, promoting stability through surface inversions. Yet, enough moisture to produce precipitation is retained, provided that lifting of several thousands of feet occurs.

Summertime mP

The surface layers of summertime mP air masses are usually cooler than the air aloft. Upper warming is a result of subsidence, not a feature of the source region. Inversions are frequently present at the surface.

An interesting situation is found with summertime mP air reaching the central coast of California. A very strong inversion, caused by an unusual air-ocean interaction, characterizes the lower layer. The circulation around the Pacific high-pressure center nearly parallels the Pacific Coast. Friction with the water surface produces a gentle north-northwest to south-southeast current. The CoF deflects the ocean current slightly, but surely, seaward. Near the coast just north of San Francisco, an upwelling of deep, cold water replaces the surface water. This upwelling is usually noticeable from June through September; the strongest upwelling occurs in August. Contact with this cold water produces a strong surface inversion in the air mass.

Wintertime mT

Wintertime mT air masses are important in weather phenomena because of their large moisture content. The moisture is released as precipitation when lifting occurs along the coast, though the air becomes mTw as it moves over the land. Most of the Pacific Coast wintertime storms involve mT air.

Summertime mT

Summertime mT air rarely reaches the Pacific Coast. When it does, it produces showery conditions known locally as *Sonora* weather. The mT air from the Gulf of Mexico is much more important for the United States summer weather. As it moves inland it becomes mTk, although the temperature differences between land and air at the surface aren't considerable. The thermal heating at the surface accentuates the inherent instability of the mTk air. The mT air penetrates as far north as Canada, flowing along the east side of the Rocky Mountains, then through the Mississippi River drainage basin.

017. Geographical distribution of air masses and their effect on weather

In this lesson we again look at air masses, but now we concentrate on their locations and the weather they produce.

Air masses

An air mass is identified by the horizontal similarity achieved by the stagnation of air over large areas of land or water in certain regions of the general circulation system. These areas are the air mass source regions. The surfaces (either land or water) of these source regions impart characteristics to the air masses, but once the air mass begins to move, the original characteristics are modified. The following are all causes of the modifications:

- General circulation effects.
- Air mass source regions and classification.
- Air mass characteristics.

General circulation effects

In the typical general circulation patterns, persistent cells of high pressure exist in the zone between 25 and 30° latitude in both hemispheres.

Zones of lower pressure appear at or near the equator and between latitudes 45 and 60°. These zones don't remain stationary throughout the year, but shift northward and southward within the latitude ranges of 45 and 60°. This is indirectly associated with the seasonal differences in the solar heating of the earth's surface. In other words, the zones follow the seasonal migration of the sun. In each hemisphere, all zones reach their most poleward extension during their respective summer months and are nearest the equator during their respective winters. Besides these semipermanent regions of high-pressure cells over the oceans, there are high-pressure cells over the continents in both hemispheres during the winter season. These cells are produced by the cooling of all extensive land surfaces.

Air mass source regions and classification

Earlier, we discussed air mass source regions and air mass classification. In this lesson segment, we'll provide additional information on the following air masses:

- Continental polar (cP).
- Arctic (A).
- Maritime tropical (mT).
- Continental tropical (cT).
- Equatorial (E) and maritime polar (mP).

Refer back to figures 1–44 and 1–45, which show the world air masses, fronts, and centers of major pressure centers in January and July, respectively.

Continental polar

The cP air of winter originates in the persistent cold barotropic high-pressure area that develops over Asia during the winter and in the winter high that frequently forms over the tundra region of North America. From the Asiatic high, winter cP air flows seaward as monsoon winds over the North Indian and Southwest Pacific Oceans. It also flows into the North Pacific to form the polar front. The Asiatic cP air sometimes moves to the Arctic Siberian coast; occasionally, the extensive and deep Asiatic high spreads sufficiently to bring cP air to the coasts of Europe. In North America, cP air masses move southward over the Great Plains and seaward to the North Atlantic.

In summer, the continental highs dissipate over the heated land and the sources of cP air move farther north.

Arctic

Arctic air in the winter, although sometimes indistinguishable from polar air, is commonly colder. It originates over the Arctic ice cap and in the great polar high over the Greenland ice cap. Arctic air accounts for the severest cold waves of the midwestern winters. Arctic air masses and cP air masses are similar in the American area. Frontal activity between Arctic air and cP air does occur in this region. However, it also occurs along the Arctic coast of Eurasia and over the North Atlantic and North Pacific Oceans, where mP air predominates.

Maritime tropical

The mT air masses have their sources in the maritime subtropical high-pressure cells, which are areas located over the oceans at about 30° latitude. The mT air masses move outward in most directions. Along the NETWCZ, they encounter tropical air masses of the opposite hemisphere. Along the polar fronts, they meet polar or Arctic air masses.

Continental tropical

The cT air originates in subtropical high-pressure areas that extend over continents associated with thermal lows. During winter in the Northern Hemisphere, the Sahara high sends cT air toward the equator and the NETWCZ, and northward to frontal activity over the Mediterranean, or, occasionally, to frontal disturbances over northwestern Europe. In summer, this course of cT air extends farther north.

During the southern winter, a continental subtropical high over Australia sends cT air northward toward the summer low that develops over Asia, and southward to meet polar air from the South Pacific.

Equatorial and maritime polar

Although semipermanent highs are the principal origins of air masses, air may remain over certain other regions long enough to acquire characteristics of the underlying surfaces. This occurs most frequently in the E and mP regions in the following manner:

1. Over equatorial oceans, stagnated or slowly moving air masses become modified to E air.
2. Over the central and eastern portions of the North Atlantic (south of Iceland) and over the central and eastern portions of the North Pacific (south of Alaska), stagnated or slowly moving air masses become modified to mP air. This mP air may originally have been either continental polar, arctic, or maritime tropical.

Air mass characteristics

The initial characteristics of the air masses at their sources are retained since the course of the air stream is over regions that are similar to the source. A rapid transformation of the air mass occurs if the course is over a surface that presents radically different characteristics concerning temperature and moisture.

The rate of transformation depends upon these three factors:

- The speed with which the air mass travels.
- The nature of the surface over which it moves.
- The temperature differences between the new surface and the air mass.

Our discussion of the modification process will be simplified if you consider the following three classes of conditions:

- *Warm air moving over a colder surface.* Here, the surface layer of air undergoes rapid transformation, but, because of the stability produced by this surface cooling, the effects don't extend to great heights.
- *Cold air moving over a warmer surface.* In such a situation, the surface layer of air undergoes less rapid modification, but the effects extend too much higher levels because of the turbulent and convective exchange of heat and moisture.
- *Warm, dry air moving over a warm, moist surface.* Here, the air mass picks up moisture rapidly, but, because of the small amount of heating at the surface, the effects reach only a limited height. Under such conditions, a shallow layer of unstable air exists at the surface with a stable inversion layer above. This prevents the distribution of moisture to higher levels.

Causes of weather

Without atmospheric circulation, each region of the earth would possess a climate that would depend solely upon the amount of solar heat received and the nature of the underlying surface (whether the surface was land or water). Such a climate is most nearly realized at a source region within a semipermanent high-pressure cell.

NOTE: The study of the weather phenomena of any region is actually a consideration of the modifications (of the fictitious climate) caused by airflow.

The weather characteristics of a particular month in a given locality are governed by these five factors:

- Wind speed over that particular region.
- Origin of the air mass.
- Previous history of the air mass before it arrived over the region.
- Effects of local topography.
- Proximity of the region to a zone of atmospheric convergence.

Zones of convergence and divergence

Over the oceans and other large expanses of water, unsettled weather (widespread cloudiness, precipitation, high surface winds, and turbulence) is essentially associated with rising air currents. Bad weather zones are generally absent in high-pressure areas, which are regions of descending air (regions of surface divergence). When two air streams converge, conditions conducive to the rising and lifting of air masses exist, and bad weather zones result.

Self-Test Questions

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

013. Heat transfer

1. Explain Wein's law.

2. What wavelength of the electromagnetic spectrum allows weather satellites to obtain imagery during both day and night?

3. Why do calm, cloudy nights tend to be warmer than calm, clear nights?

4. Would thick clouds or a forest have a higher albedo? Why?

5. What can be said about the earth in terms of absorption of solar radiation and emission of infrared radiation?

6. Would soil or air have better heat conductivity? Why?

7. What's convection?

8. What can initiate convection?

9. What's advection?

014. Air mass formation

1. Name the three factors needed for air mass formation.

2. Match the methods of creating temperature equilibrium in an air mass (column B) with the method's effectiveness (column A). Column B items may be used more than once.

Column A	*Column B*
____ (1) Most effective method.	a. Turbulent-convective transport.
____ (2) Slowest method.	b. Radiational cooling.
____ (3) Fastest method.	c. Evaporation and condensation.
____ (4) Not effective by itself.	
____ (5) Effective in polar regions.	

3. Why are anticyclonic systems excellent for formation of air masses?

4. Why are cyclonic systems poor for forming air masses?

015. Air mass structure

1. Define air mass.

2. What specific characteristics do air masses acquire when they form over oceans?

3. Based on the given characteristics, classify the following air masses (use the four letter characteristic designations):

   a. Formed just south of Iceland during winter and is now over USSR. It is unstable.

   b. Formed over Canada during winter and is now over US. It is stable.

   c. Formed over Brazil during summer and is now moving toward the equator. It is unstable.

   d. Formed in Atlantic east of Bermuda during winter. Now over SE US; it is unstable.

4. In what air mass can ice fog form?

5. Which air mass is moist and unstable in the lower layers and cold and dry aloft?

6. Which air mass is hot, dry, and unstable?

7. Match the air mass in column B with its source region and characteristics in column A. Column B items may be used more than once.

Column A

____ (1) This air mass is very moist, very warm and noticeably unstable.
____ (2) Found over open oceans at high latitudes.
____ (3) Exceptionally cold, stable and very dry air mass.
____ (4) Lapse rates in the lower levels often approach the dry adiabatic lapse rate.
____ (5) Cool and moist in the lower layers, and cool and dry aloft.
____ (6) Similar to the upper-level sinking air that flows out of the subtropical anticyclones.
____ (7) Cool and dry, but not necessarily stable.
____ (8) Source region in the southwestern Caribbean.
____ (9) Moist and unstable in the lower layers and cool and dry aloft.
____ (10) Source region is the central portion of high-latitude continents.

Column B

a. Wintertime continental polar.
b. Wintertime maritime polar.
c. Wintertime maritime tropical.
d. Continental tropical.
e. Summertime continental polar.
f. Summertime maritime polar.
g. Summertime maritime tropical.

016. Air mass modification

1. In the following situations, diagnose the type of modification most likely occurring and indicate the major cause of the modification:

   a. mPw air mass moving from Oregon to Wyoming.

   b. cPk air mass moving from Canada into central plains of the US.

c. mTw air mass moving from Gulf of Mexico into the southern US.

d. cPw air mass moving across the Great Lakes.

e. cPw air mass stagnant over Illinois at night.

f. mTw air mass moving from South Atlantic Ocean to North Atlantic Ocean.

017. Geographical distribution of air masses and their effect on weather

1. Classify the following air masses and describe their characteristics:
   a. A high pressure moving over the central US from southern Canada.

   b. Air over the southern Atlantic moves over the southeastern US via the Gulf of Mexico.

   c. Air from northern Canada moves over the midwestern US, which is already under cold air.

   d. Air that originally was over the southern North Pacific that has moved over northwestern US by way of the Gulf of Alaska.

   e. Air that has stagnated over equatorial oceans.

   f. Air that moves over the Mediterranean from northern Africa.

Answers to Self-Test Questions

001

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

(7) c.
(8) g.
(9) c.
(10) b.
(11) e.
(12) a, e.
(13) b.
(14) f.
(15) b.
(16) e, g.

002

2. (1) b.
(2) f.
(3) c.
(4) a.
(5) d.
(6) f.
(7) d.
(8) f.
(9) e.
(10) f.
(11) d.
(12) f.
(13) e.
(14) b.
(15) e.

003

1. The manner in which the earth's surface is heated (differential heating).
2. It decreases.
3. Rotation and revolution.
4. The seasons.
5. In both hemispheres.
6. Oxygen and ozone.
7. The shorter wavelengths, particularly where particles are less than 0.5 microns.
8. 20 percent.
9. 3 percent.

004

1. Single-cell circulation is the result of differential heating of a nonrotating earth with a smooth surface.
2. Unequal heating of the earth.
3. (1) b.
(2) a, b.
(3) c.
(4) a.
(5) a.
(6) c.

(7) a.
(8) c.
(9) a.
(10) b.
(11) a.

005

1. Centripetal force.
2. Centrifugal force.
3. Centripetal force.
4. Coriolis force.
5. Centrifugal force.
6. Coriolis force.

006

1. (1) Coriolis.
   (2) Centrifugal.
   (3) Frictional.
   (4) Pressure gradient.
2. Pressure gradient.
3. The force that represents PGF on a constant-pressure product. It's the rate of height change with change in distance on a constant-pressure surface.
4. Geostrophic flow.
5. In the Northern Hemisphere, if the wind is at your back, lower pressure is to your left and higher pressure is to your right.
6. The wind direction will back and the wind speed will slow due to increased friction.
7. Confluence is the merging of wind flow, whereas difluence is the spreading apart of wind flow.
8. Towards higher heights; the parcels are converting kinetic energy to potential energy.
9. Cyclostrophic.

007

1. Semipermanent; high.
2. Semipermanent; high.
3. Semipermanent; high.
4. Semipermanent; low.
5. Semipermanent; low.
6. Semipermanent; low.
7. Semipermanent; low.
8. Migratory; high.
9. Semipermanent; low.
10. Semipermanent; high.

008

1. A narrow belt of strong winds, with speeds of 50 to 200 knots, in the upper troposphere.
2. Jet stream position varies between different latitudes and elevations around the earth. It even varies in latitude and elevation within a small geographical area. The jet stream may appear as a continuous band around the earth, but more often it gradually diminishes at one or more points and then reappears farther downstream.
3. The PFJ and the STJ.
4. The vertical wind shear associated with the jet stream is much stronger than the horizontal.

5. Immediately above the jet core.
6. North of the jet core.
7. (1) c.
   (2) b.
   (3) a.
   (4) d.
   (5) a, b.
8. Large horizontal temperature contrast and conservation of angular momentum.
9. Horizontal temperature contrast.

009

1. The 500-mb isotherm ribbon.
2. –17°C.
3. 29,400 to 30,300 feet.
4. 500-mb.

010

1. Conservation of angular momentum.
2. Severe weather. The interaction of the PFJ and the STJ.

011

1. a. Organized.
   b. Organized.
   c. Disorganized.
   d. Disorganized.
   e. Disorganized.
2. The proper sequence should be: a=1, b=3, c=2, d=4, e=5.

012

1. Perpendicular.
2. North.
3. South (near the point of occlusion).
4. Yes.
5. Yes.
6. The jet aloft usually dissipates.
7. (1) Lines of cirrus in bands, (2) patches of cirrocumulus or altocumulus castellanus, (3) lenticular clouds in waves, and (4) waves of altocumulus.
8. 5,000 to 10,000 feet below the jet core and 4 to 5° equatorward.
9. 10,000 to 15,000 feet below the jet core and 4 to 5° poleward.
10. Poleward side.

013

1. The wavelength at which the maximum amount of energy is emitted by an object is inversely (directly opposite) proportional to the temperature of the object.
2. Infrared.
3. Because clouds composed of tiny water vapor droplets are excellent absorbers/emitters of infrared radiation.
4. Thick clouds because they have a higher reflective capability.
5. The earth absorbs solar radiation only during daylight hours; however, it emits infrared radiation continuously, both day and night.
6. Soil, because it’s a better conductor of heat than air due to its molecular structure and density.

7. The vertical transport of atmospheric properties (heat and moisture).
8. Any individual occurrence or combination of: (1) surface heating, (2) low-level convergence, (3) orographic effects, and (4) frontal lift.
9. The horizontal transport of atmospheric properties such as heat.

014

1. (1) A surface that has comparatively uniform properties.
   (2) Stagnant air over the uniform surface.
   (3) A large divergent flow.
2. (1) a.
   (2) b.
   (3) a.
   (4) c.
   (5) b.
3. They have stagnant or slowly moving air, with divergent airflow and turbulent-convective mixing.
4. They have strong winds, convergent wind flow, and comparatively fast-moving systems.

015

1. A widespread body of air identified horizontally by temperature and moisture characteristics.
2. Mild temperatures, considerable moisture in the lower layers, moderate lapse rates, and a degree of conditional instability.
3. a. mPwu.
   b. cPks.
   c. cTwu.
   d. mTwu.
4. Wintertime continental polar air mass.
5. Wintertime maritime polar air mass.
6. Continental tropical air mass.
7. (1) g.
   (2) b, f.
   (3) a.
   (4) c.
   (5) f.
   (6) d.
   (7) e.
   (8) c.
   (9) b.
   (10) e.

016

1. a. Turbulent mixing, air mass moving over different terrain.
   b. Thermodynamic, cold air moving over warmer land.
   c. Thermodynamic, warm air moving over cooler land.
   d. Thermodynamic, warm air moving over cooler water.
   e. Thermodynamic, nocturnal radiation.
   f. Thermodynamic, warm air moving over colder water.

017

1. a. cP—It's cold (or cool depending on the time of year). The lower portion of the air mass undergoes slow modification and become less stable, with the effects extending through a relatively deep layer.
   b. mT—The characteristics are warm and moist. The amount and type of modification depends upon the time of year. In summer the land will be warmer and the lower portions less stable. In winter the land may be cooler, resulting in the lower portions of the air mass becoming more stable.
   c. A—The arctic air will be cold and dry, with slight warming in the lower levels. In this type of air mass, the stability decreases slowly. However, since the air is dry, good visibility's prevail.
   d. mP—This air mass was probably mT. Now it's modified enough in the lower portions to reclassify it as mP. It will be relatively stable and moist in the lower levels. The amount of modification that takes place as the air mass moves into the northwestern US depends on the time of year. During all seasons, the moisture is lost as it moves on shore and over the mountains.
   e. E—Equatorial air is slightly warmer than mT and slightly less stable.
   f. cT—This air mass is warm and dry. Although the air receives moisture rapidly, it does so only in a shallow layer. Therefore, a shallow layer of unstable air exists under a stable layer.

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 Extension Course Program (A4L).

1. (001) How many degrees centigrade does the temperature decrease per 1,000 meters of altitude in the troposphere?
   a. 3.5.
   b. 4.5.
   c. 5.5.
   d. 6.5.

2. (001) The stratosphere is characterized by
   a. noctilucent clouds, a "D" layer, and excellent flying conditions.
   b. a constantly warming temperature, mother-of-pearl clouds, and generally poor flying weather.
   c. a maximum temperature of 7°C, the strongest concentration of ozone, dense cirrus clouds, and occasionally poor flying conditions.
   d. a temperature that remains isothermal to about 100,000 feet, the strongest concentration of ozone, and excellent flying conditions.

3. (002) Above 13 miles, the radiation from the sun breaks down the oxygen in the atmosphere into
   a. ozone gas.
   b. argon gas.
   c. carbon dioxide.
   d. gaseous nitrogen.

4. (002) Which statement *best* describes water vapor in the atmosphere?
   a. It absorbs ultraviolet radiation.
   b. The most the air can hold is 6 percent.
   c. The more water vapor, the lighter the air will be.
   d. It keeps the earth from becoming too hot by absorbing solar radiation.

5. (003) The two atmospheric gases *most* responsible for the absorption of incoming solar radiation are
   a. oxygen and ozone.
   b. ozone and water vapor.
   c. oxygen and water vapor.
   d. ozone and carbon dioxide.

6. (004) The driving mechanism that is *mainly* responsible for the earth's large-scale atmospheric circulations is the
   a. unequal heating of the earth.
   b. rotation of the earth.
   c. Hadley cell.
   d. polar cell.

7. (005) Which force is described as any center-seeking force?
   a. Inertia.
   b. Coriolis.
   c. Centrifugal.
   d. Centripetal.

8. (005) Which force is the “equal and opposite reaction” to the center-seeking force?
   a. Inertia.
   b. Coriolis.
   c. Centrifugal.
   d. Centripetal.

9. (005) Centrifugal force (CeF) will increase when there is a decrease in
   a. mass.
   b. centripetal force.
   c. the speed of rotation.
   d. the radius of rotation.

10. (005) Coriolis force (CoF) is created by
    a. the cyclonic rotation of the earth.
    b. the anticyclonic rotation of the earth.
    c. an opposing force to pressure gradient.
    d. an apparent force, and therefore, it does not exist.

11. (005) If you throw a ball towards a stationary target from the window of a speeding vehicle, coriolis force (CoF) will cause the ball to miss the target
    a. to the right.
    b. to the left.
    c. downward.
    d. upward.

12. (006) The force that is responsible for starting the horizontal movement of air over earth’s surface is
    a. gravity.
    b. coriolis
    c. centrifugal.
    d. pressure gradient.

13. (006) The balance of forces needed for gradient *cyclonic* circulation is pressure gradient
    a. balanced against friction and coriolis forces (CoF).
    b. and friction balanced against centrifugal force (CeF).
    c. balanced against coriolis and centrifugal forces.
    d. and centrifugal balanced against CoF.

14. (007) The areas of low pressure that correspond to the belt of low pressure at 60°N created by the 3-cell circulation are the
    a. Icelandic and Asiatic lows.
    b. Icelandic and Aleutian lows.
    c. Aleutian and Asiatic lows.
    d. the Aleutian lows.

15. (008) In relation to the jet core, the *greatest vertical* wind shear is *usually* located
    a. above the strongest horizontal shear.
    b. below the strongest horizontal shear.
    c. above the jet core.
    d. below the jet core.

16. (008) In relation to the jet core, the *greatest horizontal* wind shear is *usually* located
    a. above the jet core.
    b. north of the jet core.
    c. south of the jet core.
    d. at the jet core's narrowest horizontal point.

17. (009) The *simplest* method for locating the 500 millibar frontal zone is to
    a. locate the 500 millibar maximum wind band.
    b. locate the position of the –11° Centigrade isotherm.
    c. locate the position of the –17° Centigrade isotherm.
    d. find where the thermal concentration is more than 5° Centigrade in 200 miles.

18. (009) The width of the jet stream core is *approximately* equal to the
    a. width of the 500 millibar isotherm ribbon.
    b. width of the 500 millibar maximum wind band.
    c. distance between the –17° Centigrade and the –20° Centigrade isotherms at 500 millibar.
    d. distance between the –20° Centigrade and the –26° Centigrade isotherms at 500 millibar.

19. (010) Where are the greatest velocities located in relation to the subtropical jet (SJT)?
    a. In the region of difluence of the STJ and polar front jet (PFJ).
    b. In the region of confluence of the STJ and PFJ.
    c. The STJ's eastern most extension.
    d. The STJ'S western most extension.

20. (011) "Jet fingers"
    a. often develop more than 400 miles apart.
    b. suggest that the jet stream is beginning to dissipate.
    c. are somewhat rare and usually occur either singly or in pairs.
    d. are formed during periods when the jet stream is well organized and the core exceeds 100kts.

21. (011) *Converging* contours downstream of the jet stream will cause the jet to
    a. decrease in amplitude.
    b. increase in amplitude.
    c. deflect toward lower heights.
    d. deflect toward greater heights.

22. (012) Using average surface frontal slopes, how far ahead of the surface *warm* front is the jet stream located?
    a. 300 miles.
    b. 400 miles.
    c. 500 miles.
    d. 600 miles.

23. (013) Which heat transfer process involves the transfer of energy by molecular motion from hot to cold objects?
    a. Radiation.
    b. Advection.
    c. Convection.
    d. Conduction.

24. (013) Advection transfers temperature
    a. by electromagnetic waves.
    b. horizontally by the wind.
    c. vertically by the wind.
    d. by molecular motion.

25. (014) What factors must a region possess in order to facilitate air mass formation?
 a. Uniform surface, stagnant air, and large-scale difluent flow.
 b. Must be over water, stagnant air, and large-scale difluent flow.
 c. Any nonuniform surface, stagnant air, and large-scale difluent flow.
 d. Must be over smooth land, stagnant air, and large-scale difluent flow.

26. (014) Which process is *most* responsible for the slow formation of air masses in the polar region?
 a. Loss of heat by radiation.
 b. Loss of heat by conduction.
 c. Transport of heat by turbulence.
 d. Transport of heat by evaporation and condensation.

27. (015) Air mass stability characteristics often depend on the temperature difference between the
 a. air mass and the source region surface.
 b. upper- and lower-level air, and source region.
 c. air mass and the surface over which it is traveling.
 d. air-mass source region surface and the surface over which the air mass is traveling.

28. (015) How would you classify a stable air mass that formed over land in the Arctic has now moved over the ocean's warmer surface?
 a. mAws.
 b. mAks.
 c. cAwu.
 d. cAks.

29. (015) What air mass classification signifies an unstable, maritime tropical air mass that is colder than the surface it is moving over?
 a. mTwu.
 b. mTws.
 c. mTku.
 d. mTks.

30. (015) What air mass classification signifies a stable, continental polar air mass that is warmer than the surface it is moving over?
 a. cPks.
 b. cPws.
 c. cTws.
 d. cTku.

31. (015) What air mass forms over land only during the summer?
 a. mTk.
 b. mPw.
 c. cPk.
 d. cT.

32. (016) As an air mass is heated from below, there will be increased
 a. stability and a decreased lapse rate.
 b. stability and an increased lapse rate.
 c. instability and a decreased lapse rate.
 d. instability and an increased lapse rate.

33. (016) It is winter. A cPk air mass is moving over the Great Lakes. In this situation, the southern shores of the Great Lakes will experience
    a. hail.
    b. heavy snow.
    c. thunderstorms.
    d. freezing precipitation.

34. (016) Which air mass involves *most* of the wintertime storms for the North American Pacific coast?
    a. mP.
    b. cP.
    c. mT.
    d. cT.

35. (017) Continental tropical air masses are *usually* associated with
    a. thermal lows.
    b. thermal highs.
    c. migratory lows.
    d. migratory highs.

36. (017) The rate that an air mass modifies *depends* on the
    a. temperature differences between the new surface and the air mass, the speed with which the air mass travels, and the initial characteristics of the air mass.
    b. initial characteristics of the air mass, the speed with which the air mass travels, and amount of moisture in the air mass.
    c. initial characteristics of the air mass, the speed with which the air mass travels, and the temperature differences between the new surface and the air mass.
    d. temperature differences between the new surface and the air mass, the nature of the surface over which it moves, and the speed with which the air mass travels.

37. (017) The weather characteristics of a particular month in a given locality are governed by
    a. wind speed and moisture content.
    b. effects of local topography and moisture content.
    c. wind speed and proximity to a zone of divergence.
    d. effects of local topography and proximity to a zone of convergence.

Please read the unit menu for unit 2 and continue ➔

Unit 2. Synoptic Scale Systems

IN THE last unit we examined the structure and composition of the atmosphere, atmospheric circulation, the jet stream, and air masses. We covered these individually to try to help you understand how they contribute to weather. Now, we'll put all that information together and apply it to synoptic-scale systems. The synoptic scale refers to systems ranging from 200 to 2,000 kilometers in size.

In this unit, we'll examine the types of synoptic-scale systems you'll deal with on the job. Our lessons will cover many topics. We'll first look at the basic dynamics of the atmosphere. Then, we'll cover the characteristics of cyclones and anticyclones. We conclude the unit with discussions about the characteristics of fronts and how they affect our weather.

2–1. Basic Dynamics of the Atmosphere

Webster's New World Dictionary defines the word *dynamics* as "the science dealing with motions produced by given forces." In meteorology these motions are produced by given forces and are called *atmospheric dynamics*. In our last unit, we covered the atmospheric motions associated with convergence and divergence. At that time, you learned that air flow into the center of a low is convergent, while outward flow from the center of a high is divergent in the low levels.

In this section, we'll introduce you to a more complete definition of these terms and the atmospheric motions associated with convergence and divergence. To understand this concept, you must be able to visualize the motions produced by given forces, or the dynamics within the atmosphere.

We begin by investigating horizontal divergence/convergence and their effects on surface pressure. Then, we examine the horizontal and vertical motions in the atmosphere and how they're produced.

We'll be mentioning vorticity and advection in the second and third sections of this unit. Recall, these dynamics were taught in the initial skills course and were also covered in unit 1. Because of this coverage, we won't have in-depth explanations in these sections.

We'll start by taking a look at divergence and convergence in the horizontal perspective and their relationship with surface pressure.

018. Horizontal divergence/convergence and their effects on surface pressure

Historical evidence shows that surface pressure changes are largely controlled by mass changes in the upper troposphere. This evidence also shows that the troposphere contains 85 percent of the mass and energy transformations in the atmosphere. It's primarily these energy transformations, caused by divergent and convergent forces, which produce our weather. To understand this, you must visualize the motions within the troposphere.

When we speak of the terms *divergence* and *convergence*, you must remember that, in mathematical terms, we're talking only about the measurement of divergent forces. If the resultant is greater than zero, we call it *divergence*. If it's less than zero, we then call it *convergence*.

In this short lesson, we'll cover these three subject areas:

- Horizontal divergence.
- Horizontal convergence.
- Pressure changes.

Horizontal divergence

Horizontal divergence refers to the spreading of air. Figure 2–1 illustrates the fact that when horizontal divergence occurs, the air moves away from the center of the column. This results from the air pushing downward from the top of the column—which adds mass to the column of air. The original column of air then contracts vertically and expands horizontally; that is, the total volume of the air parcel remains constant.

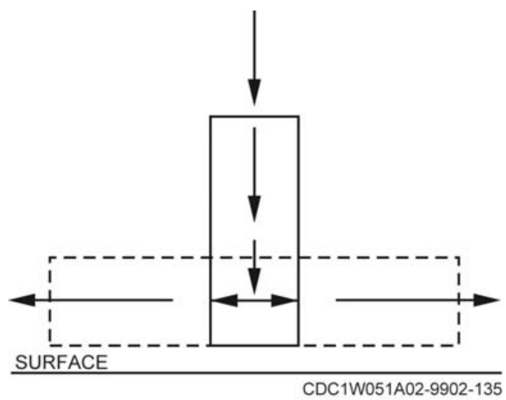

Figure 2–1. Vertical contraction produced by horizontal divergence in the surface layer.

Horizontal convergence

Horizontal convergence refers to the packing of air. Horizontal convergence of a layer near the surface, keeping the volume constant, is shown in figure 2–2. As the air converges horizontally toward the center of the layer, it creates a flow upward toward the top of the layer of air, which contracts the air layer horizontally and expands it vertically; this is caused by upper-level winds taking mass out of the column.

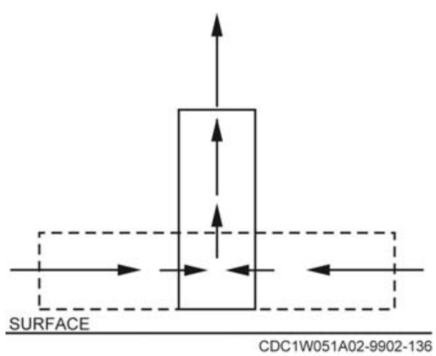

Figure 2–2. Vertical stretching produced by horizontal convergence in the surface layer.

Pressure changes

As indicated, pressure changes at the surface result from changes in mass within the troposphere. The pressure at the surface directly relates to the mass of air in a vertical column above the surface. An increase in this mass increases the surface pressure. A decrease in the mass decreases the surface pressure.

Many separate layers of horizontal divergence or convergence are possible. Therefore, the surface pressure measures the net effect of the convergence and divergence.

019. Horizontal and vertical motions in the atmosphere

As air moves inward toward the center of a surface low-pressure area, it must go somewhere. Since this converging air can't go into the ground, it must rise. Thus, we have some horizontal and vertical motions in the atmosphere. Keep these thoughts in mind as we look at the following three relationships that occur between horizontal and vertical motions in the atmosphere:

- Chimney effect.
- Damper effect.
- Convergence and divergence in the jet stream.

Chimney effect

As you can see in view A on figure 2–3, a surface low pressure develops when the winds aloft forcefully take mass out of a column of air due to upper-level divergence. This causes upward vertical motion and low-level convergence. You have divergence occurring when more air is leaving the atmospheric column than is coming into it (net divergence). When this occurs, the mass decreases within the column. Convergence occurs when more air enters the atmospheric column than leaves it (net convergence).

Divergence can be caused by one of the following:

- Heating.
- Strong upper-level winds.
- Another force that transports air aloft out of our simplistic column.

These outflow/inflow patterns not only produce vertical circulations when they're unbalanced, they produce pressure changes as well. Thus, net divergence can reduce the surface pressure and form lows or destroy highs. View A on figure 2–3 shows net divergence occurring that results in falling pressure as low-level convergence is generated. The vacancy created by the excess diverging air aloft is filled by air from below. As the low-level air converges, it has no place to go but up, thereby forming the chimney effect depicted in view A.

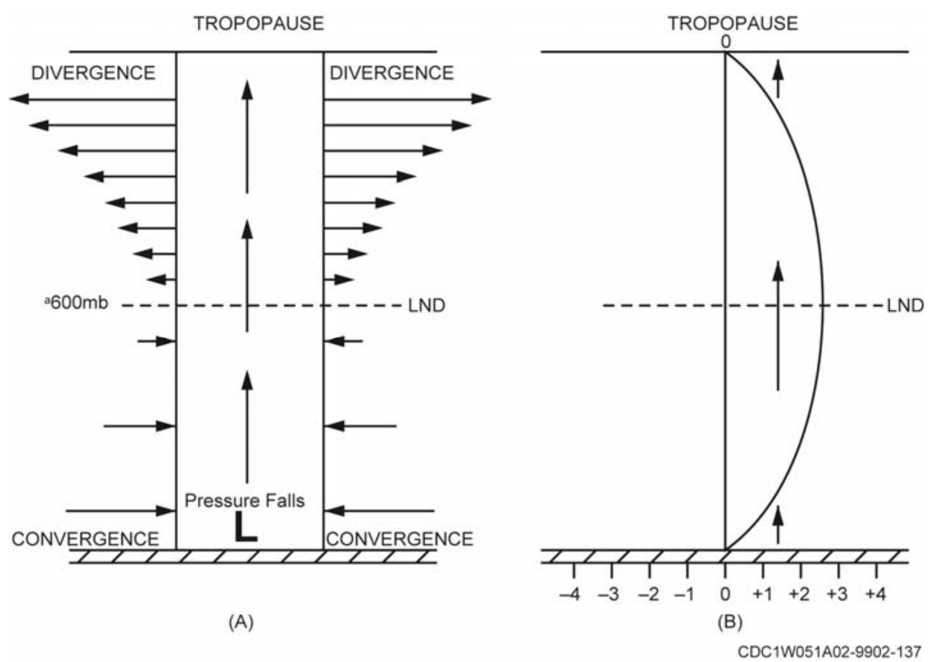

Figure 2–3. Chimney effect: view A shows the relationships of vertical motion with horizontal convergence and divergence, and view B shows the max vertical wind at the level of nondivergence.

You find these divergent wind patterns in association with a developing dynamic low east of an upper-air trough near the 300-mb level. In the middle of the troposphere, about the 600-mb level, a transition zone, commonly called the *level of nondivergence* (LND), is roughly where upper-air divergence changes to convergence in the low levels. Upward vertical motion is at a maximum at the LND and is zero at the surface of the earth and at the tropopause. This is illustrated by view B on figure 2–3. Horizontal wind speeds at the LND roughly average 40 knots, whereas the upward vertical motion is about 0.25 knots. As you can see from these numbers, the changes taking place within the column are small or slow compared to the distance the column is transported horizontally.

When the amount of divergence in the upper troposphere is greater than any convergence occurring near the surface it's referred to as *excess* or *net divergence*. Excess, or net divergence aloft, is the primary cause of deepening a developing dynamic low. Figure 2–4 shows another diagram of the chimney effect with low-level winds converging and excess divergence aloft. Notice how this effect produces clouds and weather due to upward vertical motions and adiabatic cooling.

NOTE: The term *adiabatic* refers to a change that takes place without a transfer of heat between the systems such as an air parcel and its surroundings. In an adiabatic process compression always results in warming and expansion results in cooling.

For a dynamic low, a deepening mechanism does create some minor secondary forces that offset or slow the process to some small degree. Convergence below the LND adds mass to the column of air and, as the air rises, it cools adiabatically. Neither of the effects do anything more than slightly moderate the overall chimney effect in time.

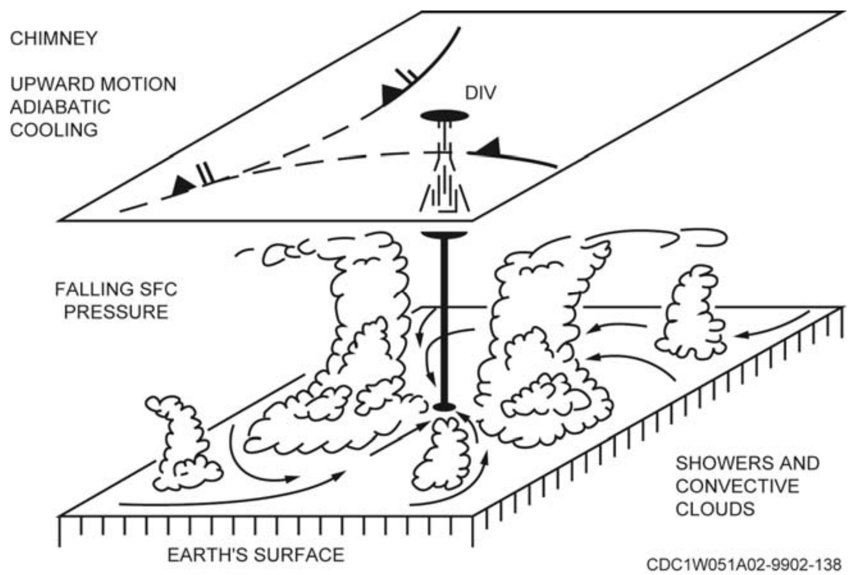

Figure 2–4. Chimney effect with associated clouds and weather.

Damper effect

The damper effect is simply the reverse process of the chimney effect. When the amount of convergence in the lower troposphere is greater than any divergence occurring at the upper levels of the troposphere, it's referred to as *excess* or *net convergence*. When converging winds aloft forcefully put mass into a column of air, it creates net convergence and thereby increases the mass within the column. As the mass increases, the surface pressure rises and, in turn, generates low-level divergence. You can find these converging wind patterns, in association with a developing dynamic high, east of an upper-air ridge near the 300-mb level. As the winds converge aloft, they can only go downward due to the highly stable tropopause. The cold, dense air at the upper levels tends to sink and the convergent upper-level winds help push the air further down the column. This forms the damper effect, as depicted in view A on figure 2–5.

Downward vertical motion is at a maximum at the LND and is zero at both the surface of the earth and at the tropopause. This is illustrated in view B on figure 2–5. It's the excess or net convergence aloft that's the primary building mechanism for a dynamic high. Figure 2–6 shows another diagram of the damper effect with low-level winds diverging and excess convergence aloft. Notice how this effect produces fair weather due to downward vertical motions and adiabatic warming. Some minor secondary forces slow the process. Divergence below the LND removes some mass from the column of the air, and as the air sinks, it warms adiabatically. Neither of the effects do anything more than slightly moderate the overall damper effect in time.

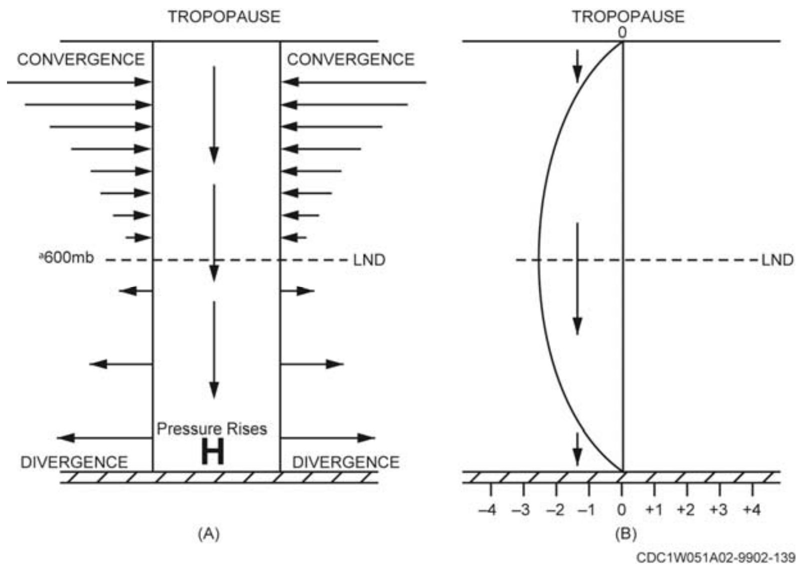

Figure 2–5. Damper effect: (view A) shows the relationships of vertical motion with horizontal convergence and divergence, and (view B) shows the max vertical wind at the LND.

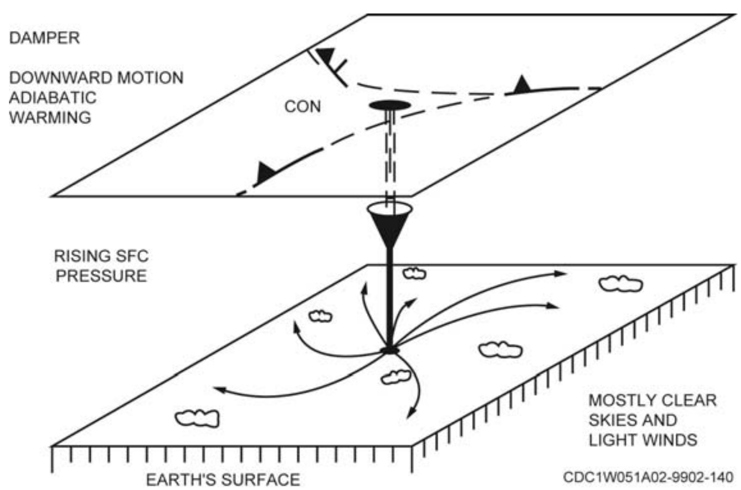

Figure 2–6. Damper effect showing mostly clear skies.

Convergence and divergence in the jet stream

Remember that jet maxima are important because of their association with migratory pressure systems. The horizontal divergence and convergence caused by the jet stream is the driving mechanism for creating, dissipating, and maintaining migratory pressure systems.

When you consider convergent and divergent areas associated with jet maximas, you must first consider the flow of the jet axis. The orientation of the jet axis enhances or hinders the magnitude of convergence and divergence occurring. For example, a jet axis flowing in a straight line, as seen in figure 2–7, actually has four separate quadrants associated with the jet maxima. Notice that the predominate vorticity is an opleth pattern that indicates a shear-jet pattern and that little advection is occurring. This makes sense when you consider that a jet in a high zonal flow pattern is relatively weak and isn't associated with significant surface systems. The left-front quadrant on figure 2–7 is considered to be divergent due to the increasing horizontal shear on the +n side.

NOTE: The term *+n* is from the three dimensional natural coordinate system that moves with the wind flow. It's very useful in understanding the processes involved with atmospheric motions.

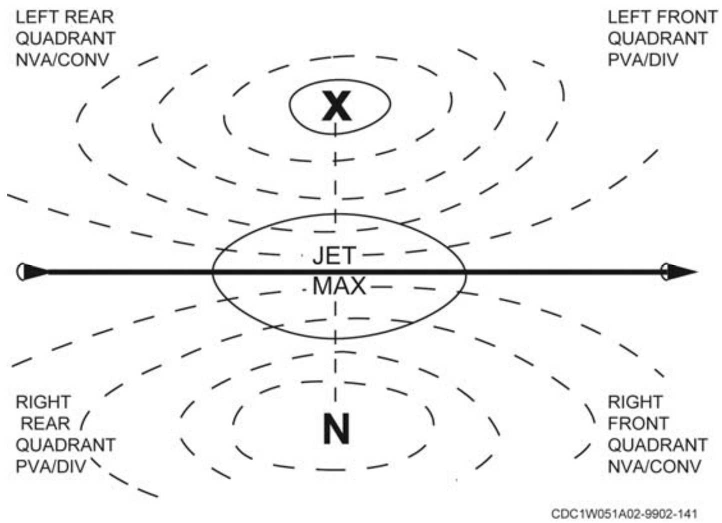

Figure 2–7. Jet max in straight-line flow.

It's a fact that the chances of the PFJ flowing in a straight line for any appreciable amount of time is rare. Therefore, you must be familiar with the modification of the quadrants in cyclonic and anticyclonic flow. When the flow becomes more cyclonic, as seen in figure 2–8, the two quadrants on the +n side of the jet axis increase in magnitude. This is due to the addition of positive curvature vorticity. On the –n side of the jet axis, the positive curvature has a counter affect on the negative shear. This leaves the two quadrants essentially weak or even neutral in magnitude.

Conversely, the right-front quadrant is convergent due to the increasing negative shear. The center of the jet max (denoted by the vertical dashed line) is basically a neutral area where the convergence and divergence cease and reverse for the rear quadrants.

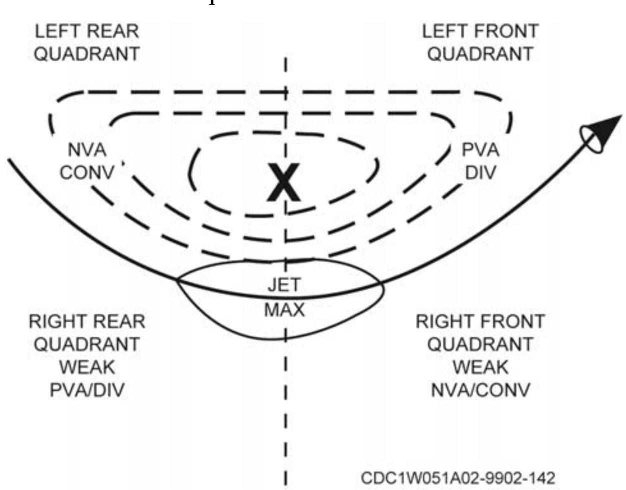

Figure 2–8. Jet max in cyclonic flow.

When the PFJ flow becomes more anticyclonic, the effect on the four quadrants is reversed. This is shown on figure 2–9. The negative curvature vorticity begins to negate the positive shear on the +n side of the jet axis. This leaves the two quadrants weak to neutral in their relative magnitude.

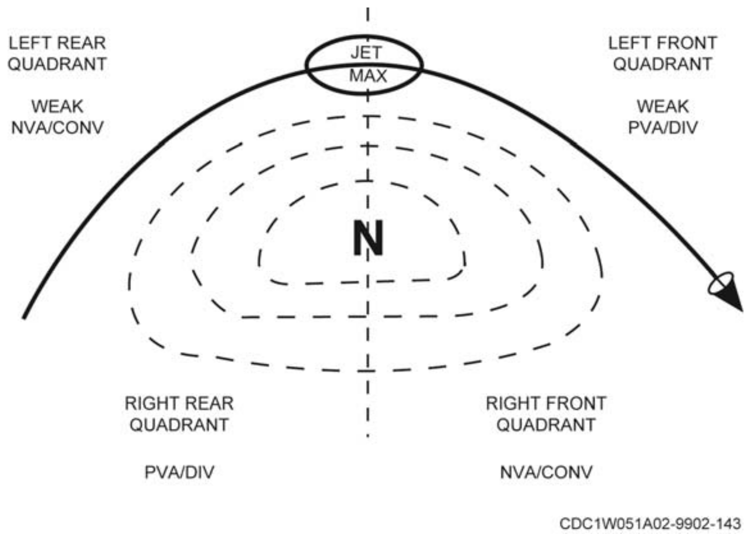

Figure 2–9. Jet max in anticyclonic flow.

Self-Test Questions

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

018. Horizontal divergence/convergence and their effects on surface pressure

1. Define horizontal divergence and horizontal convergence.

2. State how divergence and convergence can change the surface pressure.

019. Horizontal and vertical motions in the atmosphere

1. What mechanism is believed to be the primary cause of the development of high- and low-pressure areas?

2. In the following situations, specify the intensity changes that theoretically should occur:
   a. Cold-air advection into an upper-level low or trough.

   b. Warm-air advection into an upper-level high or ridge.

3. What process creates the chimney effect? What happens to the mass in the column of air?

4. What process creates the damper effect? What happens to the mass in the column of air?

2–2. Cyclones and Anticyclones

Working largely from surface observations, a group of scientists in Bergen, Norway, developed a model explaining the life cycle of an extratropical storm; that is, a storm that forms at middle and high latitudes outside the tropics. This extraordinary group of meteorologists included Vilhelm Bjerknes, his son Jakob, Halvor Solberg, and Tor Bergeron. They published their theory shortly after World War I. It was widely acclaimed and became known as the *polar-front theory of a developing wave cyclone* or, simply, the *polar-front theory*. What these meteorologists gave to the world was a working model of how a midlatitude cyclone progresses through the stages of birth, growth, and decay. An important part of the model involved the development of weather along the polar front. As new information became available, the original work was modified, so that, today, it serves as a convenient way to describe the structure and weather associated with a migratory storm system. In this section, we explore these six subject areas:

- Terms associated with pressure systems.
- Baroclinic instability.
- The development of baroclinic lows.
- Life cycle of a baroclinic low.
- Characteristics of anticyclones.
- Types of pressure systems.

020. Terms associated with pressure systems

Since weather systems are somewhat intricate, you need a thorough understanding of the terms associated with systems. In this lesson, we cover the terms that will assist you in understanding how systems develop and the terms associated with low-pressure systems.

Cyclone

We use the term *cyclone* to describe any area of closed counterclockwise circulation in the Northern Hemisphere. When this closed circulation occurs on a frontal surface, it’s called a *wave* cyclone.

A favorable place for wave cyclone development is on the polar front. This is where cold air from the high latitudes or poles converges with warm air from the low latitudes or tropics. In theory, the polar front is stationary; often, a series of wave cyclones develop along it because of the enormous difference in the air masses. As they develop, the cyclones distort the polar front and follow a distinctive life cycle.

Cyclogenesis

Cyclogenesis describes the formation of a cyclone or the deepening of an existing cyclone. Thus, when cyclogenesis occurs, the central pressure falls more rapidly than surrounding areas. There’s also evidence of an increase in the intensity of the counterclockwise circulation.

Cyclogenesis may begin at the surface and work upward, or begin aloft and work downward to the surface. Waves that work upward are usually associated with surface fronts; waves working downward usually aren’t.

Deepening

When the central pressure of a low decreases, it’s said to be deepening. Deepening usually coincides with cyclogenesis, but the two terms aren’t synonymous.

Cyclolysis

Cyclolysis is the decrease in circulation or dissipation of a low-pressure system. When the counterclockwise circulation area decreases or disappears, the low is undergoing cyclolysis.

Filling

Filling refers to an increase in the central pressure of a low, which may or may not coincide with cyclolysis. Remember, cyclolysis refers to a decrease of the cyclonic circulation.

Anticyclone

An anticyclone is a closed, clockwise wind circulation in the Northern Hemisphere with relatively high pressure. An anticyclone is commonly called a *high*. Semipermanent and migratory anticyclones, as wind systems, never reach the intensity that cyclones do. Even anticyclones of great vertical extent are less organized in formation, shape, and development.

Anticyclogenesis

Anticyclogenesis describes the formation or intensification of an existing anticyclone. Therefore, if the circulation intensity or area of closed *circulation increases*, the high is undergoing anticyclogenesis. The definition of the term anticyclogenesis requires that intensity or the area of *circulation* increase.

Building

When the central *pressure* of the high *increases*, it's said to be *building*. Building usually coincides with anticyclogenesis, but the two terms aren't synonymous. Remember, there must be a change in *circulation* to have anticyclogenesis.

Anticyclolysis

Anticyclolysis is the *decrease in circulation or dissipation* of a high-pressure system. When the clockwise circulation area decreases or disappears, the high is undergoing anticyclolysis.

Weakening

Weakening refers to a *decrease* in the central *pressure* of the high which may or may not coincide with anticyclolysis. Remember anticyclolysis refers to a decrease or weakening of the anticyclonic circulation.

Short waves

Short waves are small atmospheric waves associated with temperature advection (increased baroclinicity). Short waves form as a result of an imbalance in the atmosphere as reflected in thermal instability associated with the jet stream. The long waves move cold air equatorward and warm air poleward in an attempt to balance the energy and heat in the atmosphere. Short waves develop within the long-wave pattern as a result of smaller scale instability which we'll discuss in greater detail later in the lesson.

Short waves are associated with temperature advection and are, therefore, baroclinic. The height contours and isotherms are out-of-phase and cross, resulting in temperature advection. Short-wave wavelengths are generally less than 60° of longitude. Their speeds average about 20 knots per day in the summer while in the winter, the average speed increases to 30 knots per day due to the increased speeds associated with the PFJ. If a short wave supports a baroclinic low, we call it a *short-wave trough*. If a short wave supports a baroclinic high, we call it a *short-wave ridge*.

Major short waves

A major short wave is a short wave that's large enough to distort the long-wave pattern. Major short waves also reflect through a large depth of the atmosphere (visible on more than one constant-pressure level) and support surface features such as baroclinic lows or baroclinic highs.

Minor short waves

A minor short wave is a small short wave that generally moves faster than a major short wave and causes very little distortion to the long wave. They may not be detectable at more than one level.

Terms associated with low-pressure systems

The following terms relate to low-pressure systems:

- Baroclinic low.
- Barotropic low.
- Stable waves.
- Unstable waves.
- Mature waves.
- Decaying waves.

Baroclinic low

When contours and isotherms are out-of-phase, advection is occurring and the atmosphere is described as baroclinic. In other words, a baroclinic low is a low that has temperature advection associated with it. In a baroclinic system the axis tilts with height. The baroclinic low resides in a region of strong thermal contrast. As a result, strong thermal advection exists around the low. The system tilts with height to a short-wave trough. The system stacks down toward the warm air or up to the cold air. The tilted stack of the system causes the wind to change direction with height, thus suggesting thermal advection.

Barotropic low

When contours and isotherms are in phase, the atmosphere is said to be equivalent barotropic—or simply barotropic. There's no temperature advection. Lows located in pockets of cold air are also called barotropic. Barotropic lows are vertically, or nearly vertically, stacked. This means that the upper tropospheric reflection (upper-level low) of the surface system is located directly above the surface system. These systems aren't associated with a contrast in air masses and no fronts accompany them.

You can identify the baroclinic and barotropic lows by noting the relationship between the flow and the temperatures around the system. The two views on figure 2–10 illustrate this concept. Notice the baroclinic situation has isotherms that cross the flow pattern (temperature advection is occurring). In the barotropic situation, the isotherms and flow pattern are parallel to each other (no temperature advection is occurring).

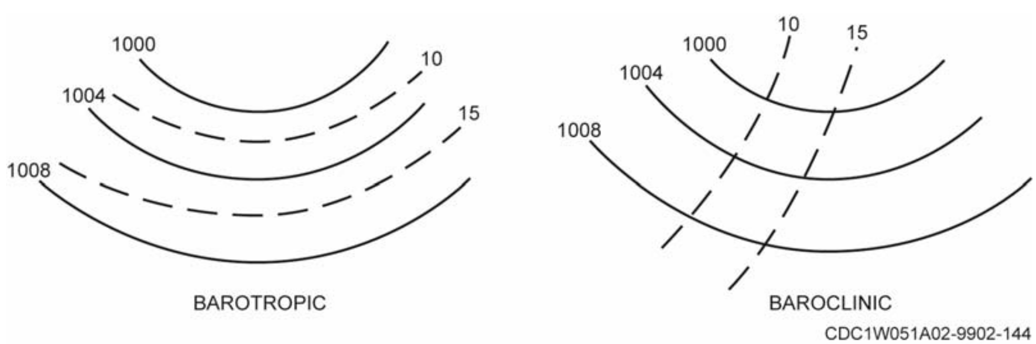

Figure 2–10. Baroclinic and barotropic flow and temperature relationships.

Stable waves

This low-pressure system is usually shallow, seldom seen above the 850-mb level, and lacks upper-air support. It has a closed-cyclonic circulation and is baroclinic. This system doesn't intensify nor does it occlude. As it begins to form, you'll find parallel windflow in the opposite direction on either side of the stationary polar front.

Local convection or heating increases the thickness in the local area, causing pressure falls that induce cyclonic flow and vertical motion. The winds on the cold side turn northerly and the winds on the warm side become southerly. Due to the low-level warm-air advection ahead of the low center and cold-air advection behind it, you'll find height and pressure falls and rises, respectively. Figure 2–11 shows how this stable wave looks—both horizontally and vertically.

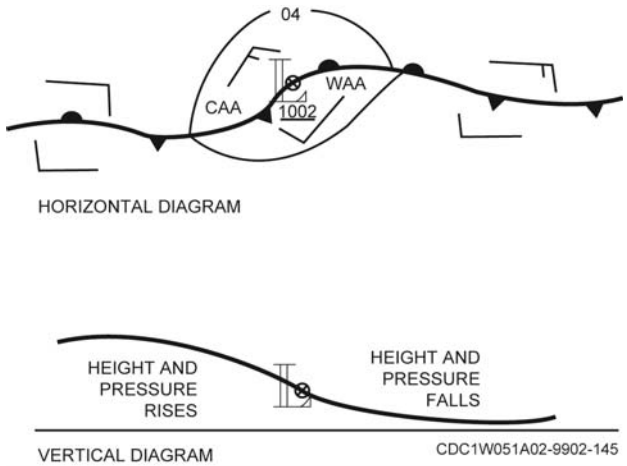

Figure 2–11. Horizontal and vertical diagram of a stable wave.

Stable waves tend to propagate along fronts but without changing intensity. Since there's no upper-air support, the movement of this wave depends on the low-level warm-air advection. The amplitude of the wave is small and the pressure rises and falls are of a comparatively small size. The wave can produce significant weather over a small area and often evolve into unstable waves.

Unstable waves

Unstable waves differ from stable waves in that their amplitude increases over time as they deepen and undergo cyclogenesis. This low is also considered a baroclinic system; however, it's much deeper than the stable wave because it extends higher into the troposphere. This system has upper-air support (i.e., major short-wave trough) that helps intensify it, and may progress into an occlusion (mature wave).

The initial development of an unstable wave is very similar to the stable wave. The upper-air support (an upper-air trough) is the mechanism that intensifies the cyclone. The major short-wave trough and the wave on the frontal system amplify with time as self-development occurs.

The unstable wave is slower moving than the stable wave and the amplitude is much larger. This is depicted in figure 2–12. As the cyclone intensifies, the cold air mass pushes further south and the warm air mass is pushed to the north. The unstable wave also has a large magnitude of pressure rises and falls compared to the stable wave. Because of this, it is an extensive weather producer, especially ahead of the warm front portion of the system.

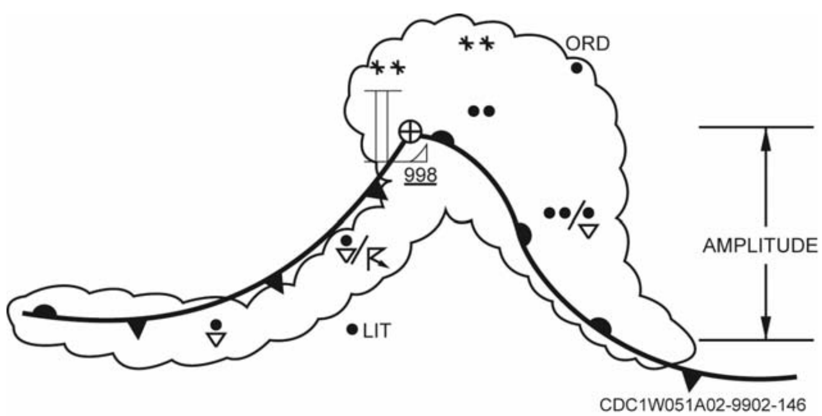

Figure 2–12. Diagram of an unstable wave, weather, and amplitude.

Mature waves

As an unstable wave continues to amplify, the low deepens back into the cold air and migrates to the +n side of the jet axis. The n axis is oriented perpendicular to the flow, with +n oriented to the left of the flow (cold side of the jet) and –n to the right of the flow (warm side of the jet). At this point the unstable wave evolves into a mature wave. The surface low center becomes removed from the surface temperature contrast and separated from the warm and cold fronts.

As the cyclone develops to this stage, you must have three different air masses:

- Cold.
- Warm
- Cool.

For the wave to occlude, the cold front rotates around the cyclone faster than the warm front and eventually overtakes it. As the cold front overtakes the warm front, one front is forced up over the other; this forms an upper front. An occlusion extends from the surface fronts into the cold air mass. The location of the coldest air mass compared with the wave determines the type of occlusion that forms. The point of intersection of the surface warm front, cold front, and occlusion is called the *triple point*. Unfavorable weather conditions occur here. In addition, it’s a favorable area for the development of a new low since cyclonic turning is already occurring at this point.

Decaying waves

Eventually, the warm air in the warm sector associated with the mature wave is pushed aloft. In being pushed aloft, the warm air is cut off from the cyclone. Then the cyclone becomes cold-core or barotropic and the low-pressure center begins to fill. This cold-core system is known as a *decaying wave*. The thermal gradient across the occlusion is gradually lost, washing out the occluded portion of the front. The two views on figure 2–13 show the placement of the air masses as the system washes out or dissipates. The wave then disappears and the stationary polar front reestablishes.

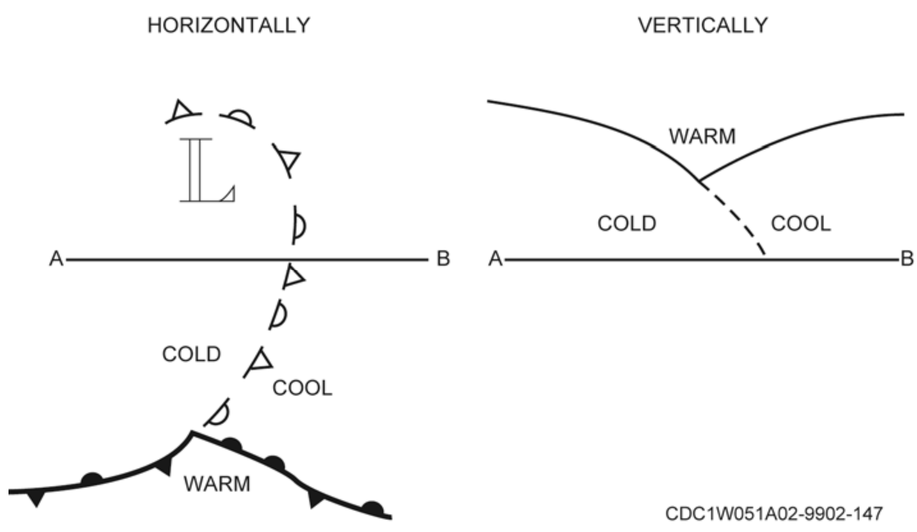

Figure 2–13. Location of air masses as a decaying wave dissipates.

021. Characteristics of baroclinic lows

In this lesson, we explore the general characteristics of baroclinic lows. As our lesson proceeds, we’ll cover three major subject areas:

- Baroclinic instability.
- Development of baroclinic lows.
- Life cycle of a baroclinic low.

Baroclinic instability

Baroclinic instability is a process where short waves amplify (increase amplitude with time) by extracting energy from the north/south temperature gradient. For our study, we'll bypass a formal investigation of baroclinic instability theory; instead, we'll examine these two important implications:

1. Long waves are relatively stable.
2. Major short waves are most likely to amplify due to baroclinic instability. The optimum growth rate occurs for wavelengths of 3,000 to 4,000 km.

Recall that differential heating leads to a strong north/south temperature gradient in the midlatitudes. It also results in the long-wave pattern we discussed previously. This gradient is the source of potential energy which baroclinic systems tap as they strengthen. In order for a major short wave to amplify significantly, there must be a large energy transfer from the temperature gradient to the wave. The potential energy of the north/south temperature gradient is transferred to the major short wave by thermal advection throughout the troposphere.

Aloft, the thermal wave and contour wave are out of phase. Cold-air advection occurs into the contour trough. In contrast, warm-air advection occurs into the contour ridge. This thermal advection, coupled with advection around the well-developed baroclinic system on the surface, causes the upper-level wave to amplify. The greater the amplitude of a short wave, the more energy is associated with it. The well-developed baroclinic system on the surface must be present to ensure thermal advection throughout the troposphere.

As the potential energy is transferred from the temperature gradient to the short wave, the short wave uses the potential energy to develop the low-level circulation of the low or high. The low-level circulation (wind) is kinetic energy. Energy is made available to the system when warm air rises and cold air sinks, as is true in midlatitude systems. Warm air moves northward ahead of a baroclinic low; behind a baroclinic high, it ascends. Cold air moves southward ahead of a baroclinic high; behind a baroclinic low, it descends.

Because the process proceeds without outside influences (if conditions are right), we call it *instability*. As the cyclogenesis/anticyclogenesis proceeds, the thermal advection increases and more energy is transferred to the wave. The energy then strengthens the low-level circulation. The process continues, repeating itself.

You can draw an analogy to an absolutely unstable layer on a Skew-T. Once a parcel is forced to rise, it continues to rise on its own, drawing energy from the unstable atmosphere. Once a short wave begins to amplify, it continues to amplify by drawing energy from the north/south temperature gradient.

Baroclinic instability is the primary mechanism responsible for the development of midlatitude synoptic-scale systems (baroclinic lows and highs). These systems help maintain the global heat balance by transporting warm air northward and cold air southward as the atmosphere attempts to gain equilibrium. Amazingly, only 0.5 percent of the total potential energy of the atmosphere is available for conversion to kinetic energy. Additionally, only 10 percent of that 0.5 percent converts to kinetic energy. Imagine the consequences if the atmosphere was more efficient!

Development of baroclinic lows

Cyclogenesis is favored by certain large-scale conditions. In the long-wave pattern, cyclogenesis typically occurs at, and just downstream from, a long-wave trough axis. Intense cyclogenesis normally occurs in troughs with a negative tilt. Negatively-tilted troughs have an axis that's oriented northwest/southeast. They cause stronger divergence, and, therefore, can support stronger cyclogenesis. In contrast, positively-tilted troughs have an axis that's oriented northeast/southwest and are more often associated with nondeveloping systems.

The relationship between the jet stream and cyclogenesis is also very important. Cyclogenesis typically occurs under difluent flow aloft. Supergradient winds deepen troughs and are associated with difluent flow. Upper-level troughs are likely to intensify when the wind speeds upstream from the trough axis are greater than the wind speeds downstream from the axis. The determination of whether the wind speeds are decreasing downstream or not is based on the contour spacing across the trough axis.

Short-wave troughs strengthen as they move into long-wave trough positions and become negatively-tilted downstream from the long-wave trough axis. Short-wave troughs intensify under difluent upper flow, which is common in negatively-tilted troughs. The negatively-tilted difluent major short-wave trough (six words to describe one trough!) stacks 1 to 3º latitude per standard level to a baroclinic low on the surface. This trough must be present to support the baroclinic low. The baroclinic low must be there to ensure temperature advection occurs throughout the troposphere, in order for the wave to amplify.

Baroclinic instability is the link between the upper- and lower-level features. You know that baroclinic instability converts the potential energy of the north-south temperature gradient to the kinetic energy of the disturbance. The strongest north-south temperature gradient occurs at the polar front. Consequently, baroclinic lows form along the polar front.

We now turn our attention to four important factors that deal with baroclinic lows:

1. Petterssen's rule.
2. Self-development/intensification of a baroclinic low.
3. Braking mechanisms of lows.
4. Dissipation of baroclinic lows.

Petterssen's rule

Petterssen's rule describes the initial interplay between the frontal system and the divergence associated with the major short-wave trough. Thus, all the mechanisms causing cyclogenesis are linked together. Petterssen's rule states that cyclogenesis occurs when and where an area of upper-level divergence becomes superimposed over a low-level frontal zone across which the thermal advection is weak. Figure 2–14 shows the jet (implied by the close spacing of thickness lines), short wave, thermal field, and positive vorticity advection (PVA). All must interact for rapid cyclogenesis to occur. If only some criteria are met, a weak surface low may form, but development is slow.

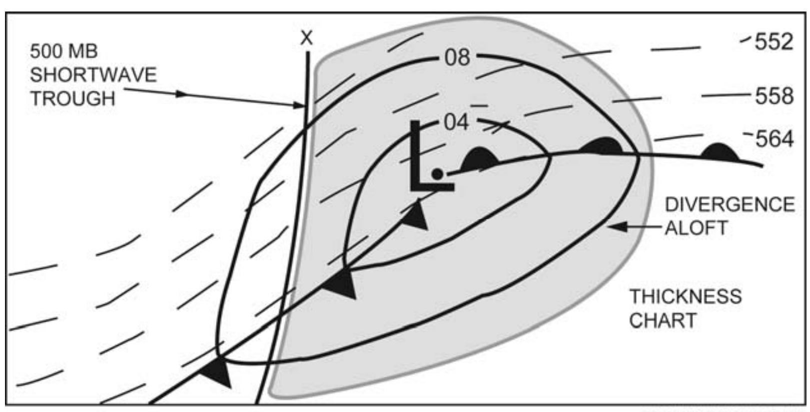

Figure 2–14. Surface cyclogenesis initiated by a short-wave trough.

You can use the application of Petterssen's rule in other synoptic situations. Short waves are constantly moving through the upper levels but only a few lead to the development of strong baroclinic lows. Short-wave troughs approach cold fronts from upstream. If divergence ahead of a short-wave trough occurs behind a surface cold front (in a region of cold-air advection) the tendency is for pressure rises to occur. Conversely, thermal advection offsets the tendency for pressure falls resulting from the divergence aloft. A surface low won't form.

If the divergence ahead of the short-wave trough occurs well behind the surface front (in a cold air mass) intensification at the surface can't occur. Even if a weak surface low begins to form, no thermal advection occurs. Thermal advection throughout the troposphere is required for amplification of the short wave.

The thermal advection required to amplify the upper-level short wave must be supplemented with surface pressure falls. This is initiated by divergence aloft for the entire system to intensify. This combination is most common along the slow moving portion of a surface front. Since winds already turn cyclonically across surface fronts, cyclogenesis more likely occurs along the front than away from it.

Self-development/intensification of a baroclinic low

We can illustrate the process of baroclinic instability further by focusing on the developing system rather than on energy transformations. This is known as *self-development*. Baroclinic lows tilt with height and stack from upper-level short-wave troughs. Because of the tilted stack, the divergence aloft ahead of the short-wave trough is located above the surface low. Divergence over the surface low causes the low to deepen. This coincides with cyclogenesis as the cyclonic circulation strengthens. Here are three rules that apply:

1. The cyclonic circulation strengthens the temperature advection occurring around the low. Stronger warm-air advection occurs into the short-wave ridge located ahead of the surface low. Stronger cold-air advection occurs into the short-wave trough supporting the low. This pattern causes the short-wave ridge to build and the short-wave trough to deepen. This is illustrated on figure 2–15.
2. The ridge becomes more anticyclonically curved so the vorticity minimum in the ridge becomes smaller. (For example, a "10" vort min becomes a "06.") The trough becomes more cyclonically curved so the vorticity maximum within the trough becomes larger. (For example, a "14" vort max becomes a "20." This is shown on figure 2–16.

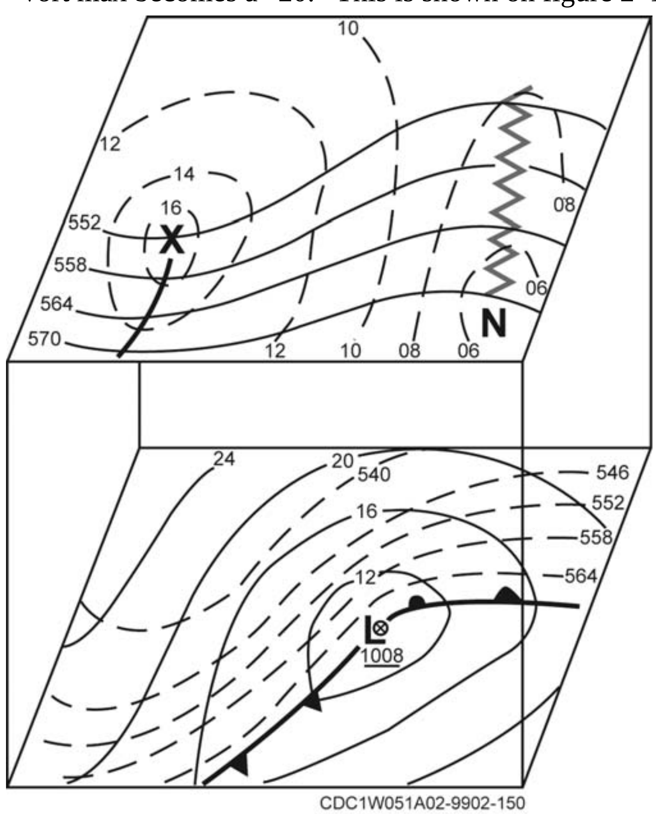

Figure 2–15. Self-development of a low.

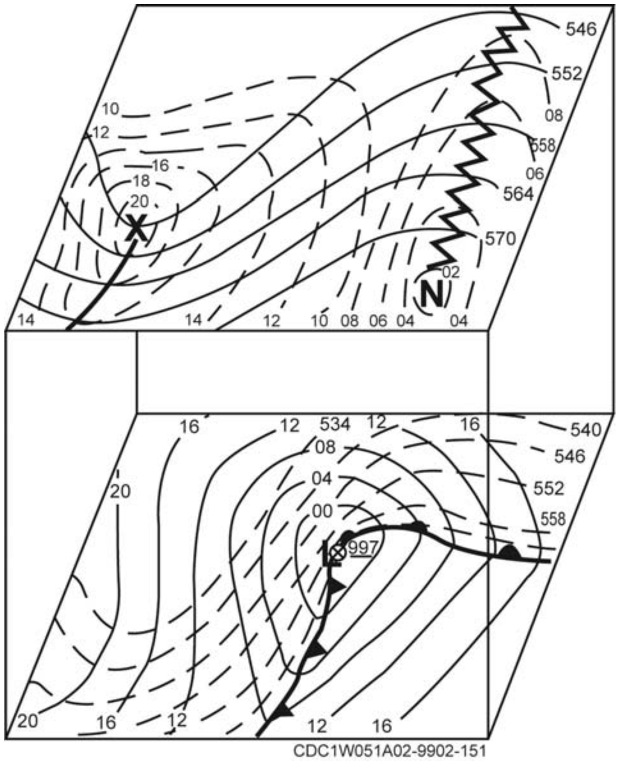

Figure 2–16. Low self-developing process continuing.

3. The divergence ahead of the short-wave trough strengthens. The positive vorticity advection between the trough and ridge is stronger. The stronger the positive vorticity advection, the stronger is the divergence. This divergence causes the surface low to deepen further and further cyclogenesis to occur. This is illustrated on figure 2–17.

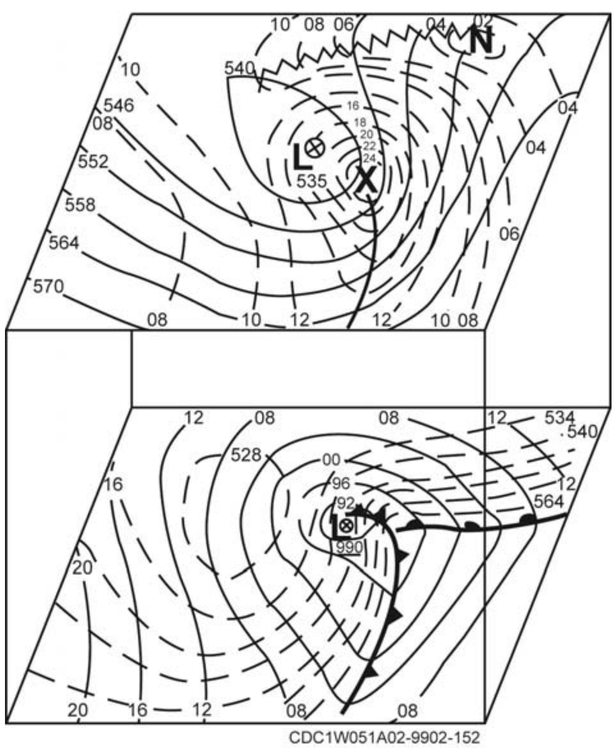

Figure 2–17. Continued development of a baroclinic low.

The above pattern continues as a positive feedback loop by which the upper-level dynamics deepen the surface low and the temperature advection intensifies the upper-level dynamics. This process is called the self-development process because once development starts, it continues until another influence acts to offset or stop it.

Braking mechanisms of lows

Although self-development causes baroclinic lows to strengthen rapidly, the intensification immediately initiates other processes known as braking mechanisms. These mechanisms slow, and ultimately stop self-development and prevent the system from deepening indefinitely.

As a baroclinic low deepens due to divergence aloft, the low-level cyclonic circulation increases. Cyclonically curved flow in the boundary layer causes low-level convergence (a consequence of friction). The addition of mass through low-level convergence offsets the mass removed by divergence aloft. This process, which can act as a braking mechanism, is known as *boundary-layer convergence* (BLC) or *Ekman pumping*.

The vertical motion field around the low intensifies as the low strengthens. Stronger upward vertical motion cools the air adiabatically. The adiabatic temperature change within the developing system acts as a braking mechanism.

Physical processes can affect the development of lows as the ascent of dry, stable air opposes development and extracts significant amounts of energy from the low. Forced ascent of moist stable air opposes the development of the low, although not as strongly as dry stable air. This occurs because some energy is returned to the low by the release of latent heat. As conditionally unstable air is forced to ascend, it can actually enhance the development of the low—if the energy returned by latent heat release is greater than that extracted from the low in lifting the air to its level of free convection (LFC).

Dissipation of baroclinic lows

Baroclinic lows may dissipate in one of two ways. First, the low can lose its upper-level support at any point in its life cycle (the life cycle of lows is covered in the next section). Without upper-level support, the low fills.

Second, the low can proceed through its entire life cycle and become a decaying wave. The system slowly dissipates from the bottom up as boundary-layer convergence adds mass to the column of air and surface pressure rises. Adiabatic cooling of the rising air causes thickness to decrease; the upper low remains, and is situated in a cold pocket. The closed low at 500mb often remains well after the surface low fills.

Life cycle of a baroclinic low

Now that we've examined the development of baroclinic lows through Petterssen's rule and self-development, we'll turn our attention to their classic development patterns. By familiarizing yourself with the symptoms evident on the 500-mb heights/vorticity and surface/thickness products, you can determine how far a particular system has evolved.

When several cyclones are present along a front, we call them a family. These wave cyclones develop in similar patterns and progress through a maturing process, or life cycle, unless the system loses its dynamic support (upper-level trough/low). The families tend to develop in a series, with each succeeding wave cyclone developing similarly to the preceding one. Figure 2–18 illustrates how a typical wave cyclone family appears on the surface product.

A baroclinic low developing along a front is called a *frontal wave cyclone*. Frontal wave cyclones can be divided into these two types—those that amplify and those that don't. A baroclinic low in which the amplitude of the wave along the frontal system doesn't change with time is a stable wave. An unstable wave is a baroclinic low in which the amplitude of the wave in the frontal system increases with time.

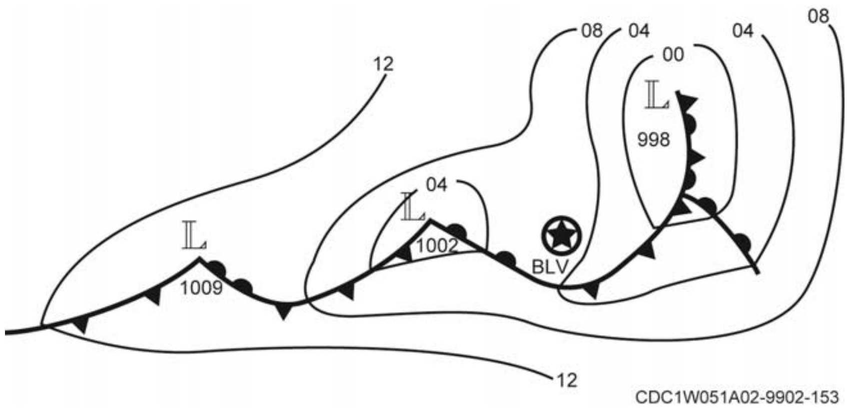

Figure 2–18. A wave cyclone family.

Life cycle of the classic baroclinic low

We'll now explore the evolution of the classic baroclinic low whose life cycle can be arranged in these five stages:

1. Initial conditions.
2. Wave initiation.
3. Wave intensification.
4. Mature wave.
5. Dissipation.

Initial conditions stage

The initial conditions stage consists of a baroclinic zone located downstream from a long-wave trough. The baroclinic zone may be either a slow-moving front or a thermal gradient organizing into a front. On the 500-mb heights/vorticity product you'll notice a short-wave trough approaching the baroclinic zone. Development doesn't occur until the divergence aloft ahead of the trough is over the baroclinic zone. On the surface/thickness product, the baroclinic zone/surface front separates cold from warm air. This front is usually a stationary front or a slow-moving cold front.

Wave initiation stage

During wave initiation, the second stage, the upper-level short-wave trough, causes a wave to form along the front. On the 500-mb heights/vorticity, the short-wave trough nears the surface front. Divergence aloft ahead of the short-wave trough starts to occur over the surface front. On the surface/thickness product, a wave develops along the surface front. Indications of wave formation include winds on the cold air side of the front turning in toward the front and precipitation in the region of the developing wave increasing. Widening of the cloud band at the center of the wave, which is evident on satellite imagery, occurs as the upper-level short-wave trough approaches.

Wave intensification stage

The third stage, wave intensification, is reached as the frontal wave and the upper-level short wave continue to interact and self-development occurs. At 500mb, thermal advection associated with the developing baroclinic low causes the upper-level short-wave trough to deepen. The vorticity maximum in the base of the trough strengthens. On the surface, the frontal wave is unstable and amplifies with time. The surface low is rapidly deepening.

Mature wave stage

As self-development continues to occur, the wave reaches its maximum intensity during the mature wave stage. During this fourth stage, the short-wave trough begins to "close off" at 500mb. In other words, a closed contour forms around a 500-mb low. The vorticity maximum peaks during this stage. PVA, and the corresponding divergence aloft, are intense initially, but weaken rapidly as the vorticity maximum becomes coincident with the closing low at 500mb. At the surface, the low has occluded and is near peak intensity (lowest central pressure). It reaches its zenith just after it occludes. The surface low is removed from the warm air mass at low levels and continues to develop back into the cold air. The vertical tilt between the 500-mb low and the surface low decreases, but the system isn't yet vertically stacked.

Dissipation stage

The fifth and final stage is called dissipation. Here, the system evolves into a cold barotropic low. The 500-mb low develops more closed contours and becomes more circular and less wave-shaped. The vorticity center is still strong, but is centered in the 500-mb low. Weak PVA suggests the lack of upper-level divergence. The surface low is well back in the cold air mass, beneath the upper low. The system is vertically stacked. The surface low is separating from the occlusion and fills and vanishes long before the upper low. The low has now evolved into a decaying wave. This concept is illustrated on figure 2–19.

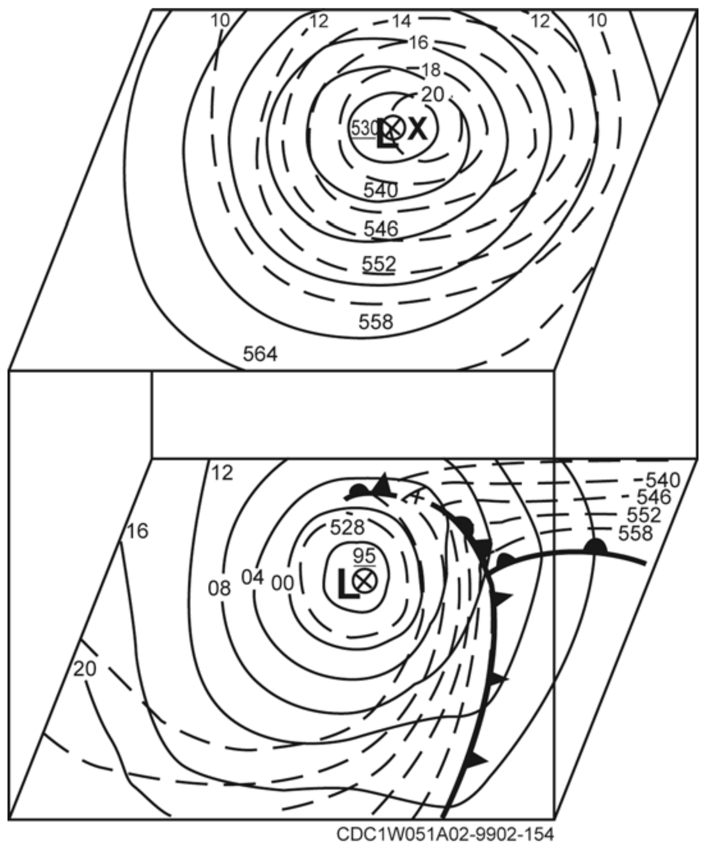

Figure 2–19. Dissipation.

022. Characteristics of anticyclones

Earlier we said anticyclones are closed, clockwise wind circulations in the Northern Hemisphere. We'll now look at some characteristics associated with anticyclones. They include the following:

- Horizontal extent of anticyclones.
- Vertical extent of anticyclones.
- Anticyclogenesis of a baroclinic high.
- Self-development/intensification of a baroclinic high.
- Braking mechanisms of baroclinic highs.
- Dissipation of baroclinic highs.

Horizontal extent of anticyclones

Anticyclones are usually larger than cyclones. An anticyclone may span 2,000 miles or more. Anticyclone size is determined by the size of the air mass with which the anticyclone is associated. Normally, semipermanent anticyclones have a greater horizontal extent than their migratory relatives.

Vertical extent of anticyclones

The vertical extent of anticyclones depends on the characteristics acquired in the air-mass source region and later modification. Subtropical (semipermanent) systems have great vertical extent and are persistent. They're stationary or move very slowly. Still, small parts of these semipermanent features may become separated from the main system and move as a migratory high along the edge of the semipermanent high. The smaller highs of lesser vertical extent provide areas of clearing weather between frontal systems.

The more migratory anticyclones, such as those associated with continental polar air masses, are of far less vertical extent than subtropical highs. They stack to short-wave ridges (the ridge is the upper system). The high itself rarely extends above 10,000 feet. These very cold and dry air masses are forced out of their source regions and cold fronts mark their leading edge. Although slow to change, after 2 or 3 days over warm water, they begin to take on the characteristics of subtropical anticyclones.

Anticyclogenesis of a baroclinic high

As with cyclogenesis, anticyclogenesis is favored by certain large-scale conditions. Anticyclogenesis typically occurs at, and just downstream from, long-wave ridge axes in confluent flow aloft. Upper-level ridges are likely to intensify when the wind speeds upstream from the ridge axis are weaker than the winds downstream, as is common with confluent flow. This describes a subgradient condition (subgradient winds build ridges).

A major short-wave ridge and surface baroclinic high must be present for amplification of the wave to occur. As with baroclinic lows, baroclinic highs derive their energy from the north-south temperature gradient. Baroclinic highs don't form along a front, however, but on the cold air side of the front. The surface divergence associated with baroclinic highs precludes the formation of a front in the region of anticyclonic flow.

Both Petterssen's rule and self-development are analogous for anticyclones. A difference in the application of Petterssen's rule is that the high doesn't develop on the front, but on the cold air side of the front.

Self-development/intensification of a baroclinic high

Baroclinic highs tilt with height and stack from upper-level short-wave ridges. These four rules apply:

1. Because of the tilted stack, the convergence aloft ahead of the short-wave ridge is located above the surface high. Convergence over the surface high causes the high to build. This coincides with anticyclogenesis as the anticyclonic circulation strengthens.
2. The anticyclonic circulation strengthens the temperature advection occurring around the high. Stronger warm-air advection builds the short-wave ridge supporting the high. Stronger cold-air advection deepens the short-wave trough. This pattern causes the short-wave ridge to build and the short-wave trough to deepen.
3. The ridge becomes more anticyclonically curved so the vorticity minimum in the ridge becomes smaller. (For example, a "08" vort min becomes a "06.") The trough becomes more cyclonically curved so the vorticity maximum within the trough becomes even larger.
4. The convergence ahead of the short-wave ridge strengthens. The negative vorticity advection between the ridge and the downstream trough is stronger. The stronger the negative vorticity advection, the stronger is the convergence. This convergence causes the surface high to build further and additional anticyclogenesis to occur.

Braking mechanisms of baroclinic highs

There's a limit on the strength of the anticyclonic gradient wind and self-development doesn't occur as vigorously as for baroclinic lows. Braking mechanisms for highs are much more efficient than for lows. As the high builds, these braking mechanisms begin to strongly oppose further development. As a result, highs never develop or "wind-up" as much as lows. Additionally, the same braking mechanisms that affected the low retard the development of the baroclinic high. Three important factors are involved in the braking mechanism:

- Boundary-layer processes.
- Adiabatic temperature changes.
- Changes in structure.

Boundary-layer processes

As a baroclinic high builds due to convergence aloft, the low-level circulation increases. Within the boundary layer, anticyclonically curved flow causes low-level divergence that "partially" offsets the mass being added to the system aloft. Friction is primarily responsible for this effect. This phenomenon removes mass from the surface high center.

Adiabatic temperature changes

The vertical motion field around the high intensifies as the high strengthens. Since highs are associated with subsidence, the energy used to force the subsidence is taken from the developing high, thus reducing the amount of energy available for anticyclonic development. This is true no matter what the characteristics of the air mass are (dry/stable, moist/stable, conditionally unstable). This factor limits the strength the baroclinic high can attain. Because the air is forced to subside, there's no latent heat energy returned to the system.

Changes in structure

The baroclinic high initially forms under convergence aloft, on the cold air side of the polar front within a polar air mass. Occasionally, the high well north of the polar-front jet may move southward because of a change in the long-wave pattern and evolve into a baroclinic high. As the high moves southeast, where cold-air advection is causing pressure rises, subsidence and diabatic heating cause the air at the center of the high to warm. During its life cycle, the high moves from the +n (cold) side of the polar-front jet to the –n (warm) side. The air mass warms, the contrast across the front ahead of the high diminishes, and eventually the high is absorbed into the subtropical ridge.

Dissipation of baroclinic highs

Baroclinic highs may dissipate in one of two ways. First, the high can lose its upper-level support. This can happen at any point in the life cycle of the high. Without upper-level support, the high weakens.

Second, the high can proceed through its entire life and become absorbed into the subtropical ridge. As the high proceeds through its life cycle and moves southeast, the air in the center warms. The front ahead of the advancing high frontolysizes as the temperature across the front diminishes. The high eventually becomes a warm high and is absorbed into the subtropical ridge.

023. Types of pressure systems

You already know there are at least two types of pressure systems—highs and lows. Highs and lows are categorized further according to individual characteristics, such as vertical extent, vertical structure, and horizontal temperature distribution, which is the key to classification and identification of the systems. We can divide horizontal temperature distribution into these two classes:

1. Baroclinic.
2. Barotropic.

Remember, the term baroclinic describes a horizontal temperature distribution that's nonuniform within the system. Thus, temperature advection is associated with the system. We also stated that the term barotropic describes temperatures *not* changing within the system; that is, there's neutral advection.

Knowledge of the types of pressure systems and their structure will aid your understanding of weather and eventual analysis of these pressure systems. Our study centers on the basic types of pressure systems. They are as follows:

- Cold barotropic high.
- Warm barotropic high.
- Baroclinic high.
- Warm barotropic low.
- Cold barotropic low.

Cold barotropic high

An illustration of the cold barotropic high is shown on figure 2–20. This high is one in which temperatures on a horizontal level decrease toward the center. Thus, the temperature in the center of the high is lower than it is toward the outside of the system. The pressure at the center of these systems on the surface is very high, but the pressure decreases rapidly with height, as does the anticyclonic circulation. A cold barotropic high rarely extends more than 10,000 feet above the surface and is characterized by low-pressure or cyclonic circulation above it. This type of high forms in a source region for continental high-pressure systems and is a migratory pressure system. For example, cP air masses are cold barotropic highs.

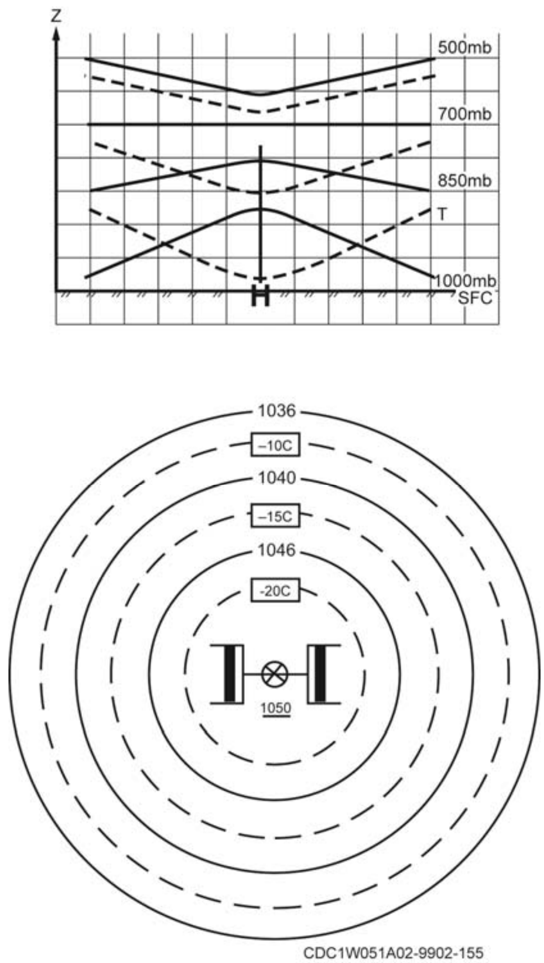

Figure 2–20. Cold barotropic high.

Warm barotropic high

An illustration of a warm barotropic high is shown on figure 2–21. This high is one in which temperatures on a horizontal level increase toward the center. Temperatures in the center of a warm barotropic high are higher than they are on the outside of a system. The pressure at the center is high. Also, the high and anticyclonic circulation increases in intensity with altitude. This type of high has great vertical extent and is usually found over water areas. Semipermanent subtropical high-pressure systems, such as the North Pacific High and the North Atlantic High, are classified as warm barotropic highs.

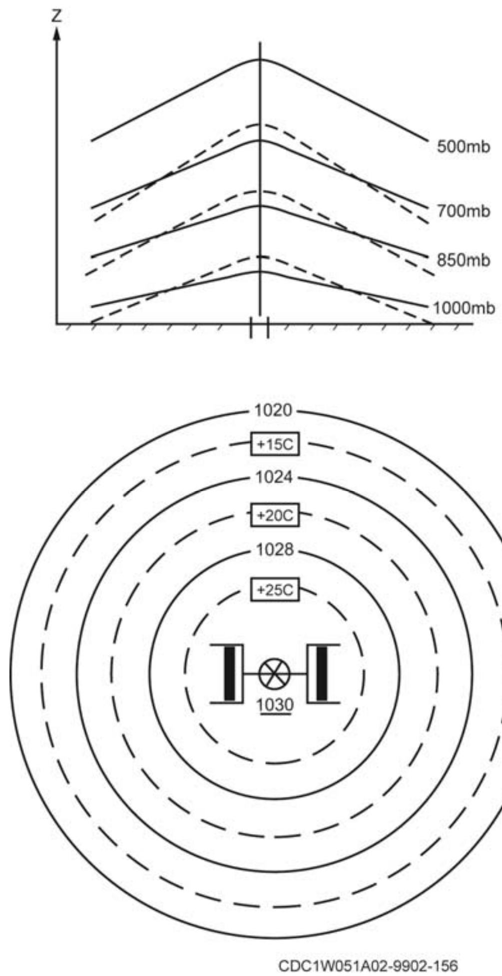

Figure 2–21. Warm barotropic high.

Baroclinic high

Figure 2–22 shows the basic structure of a baroclinic high. This is a high-pressure system with an axis that tilts toward warm air with increasing height above the surface. At upper levels, it exists as a short-wave ridge. Because they're baroclinic, these systems have temperature advection associated with them.

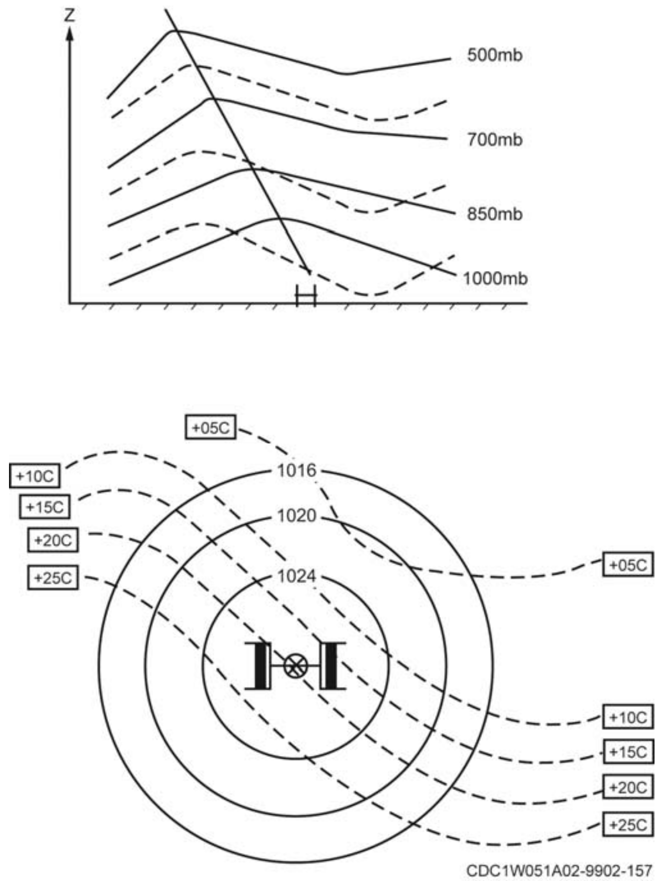

Figure 2–22. Baroclinic high.

Warm barotropic low

An illustration of a warm barotropic low is shown on figure 2–23. This low has horizontal temperatures that increase toward the center of the pressure system. A warm barotropic low is like the cold barotropic high in that it exhibits decreasing intensity and cyclonic circulation with height. A warm barotropic low rarely extends more than 10,000 feet vertically and is often found in warm regions under the subtropical ridge. An example is the "heat" low that forms over the desert southwest of the United States during the summer months. Warm barotropic lows of this nature are considered to be semipermanent in nature.

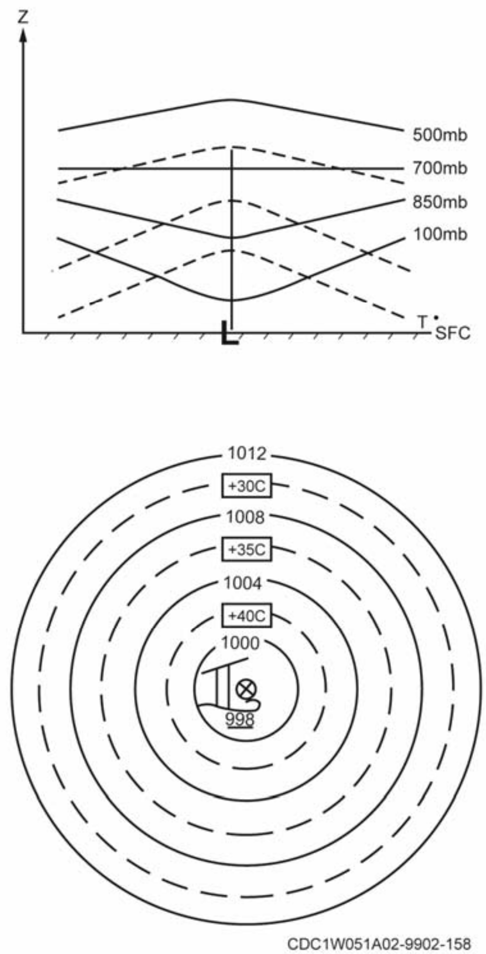

Figure 2–23. Warm barotropic low.

Cold barotropic low

An illustration of a cold barotropic low is shown on figure 2–24. This low has horizontal temperatures that decrease toward the center of the low. A cold low is like the warm barotropic high in that its intensity and cyclonic circulation increases with height. It also has great vertical extent. Cold barotropic lows may be associated with surface occluded frontal systems (as decaying waves).

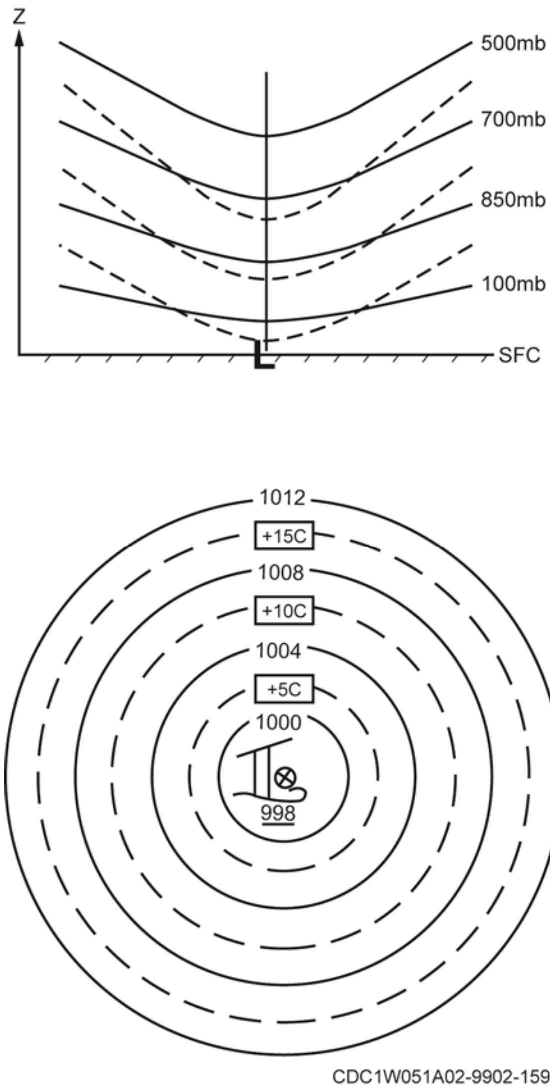

Figure 2–24. Cold barotropic low.

Baroclinic low

This low is shown on figure 2–25. As stated previously, a baroclinic low has an unstable frontal wave and associated upper-air short-wave trough. Also, colder temperatures behind the system and warmer temperatures ahead characterize it. In other words, cold-air advection is occurring to the rear of the low. This type of low is normally in the process of deepening. These lows have a vertical axis that tilts toward the cold air associated with the upper-level short-wave trough.

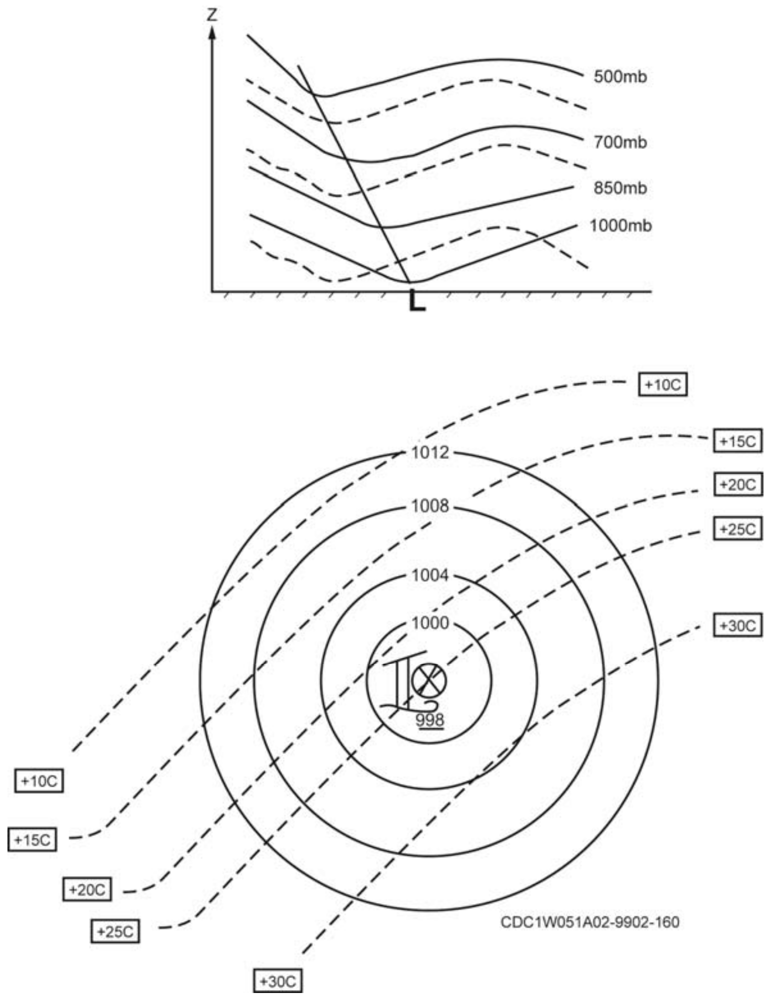

Figure 2–25. Baroclinic low.

Self-Test Questions

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

020. Terms associated with pressure systems

1. Define wave cyclone.

2. Explain the difference between stable and unstable waves.

3. For each of the following characteristics, distinguish whether cyclogenesis or cyclolysis is occurring:

   a. Central pressure within a cyclone is decreasing.

   b. Central pressure within a cyclone is increasing.

c. The deepening of an existing cyclone.

d. The filling of an existing cyclone.

e. Central pressure is rising more rapidly than the pressure around it.

f. Central pressure is falling more rapidly than the pressure around it.

g. Intensity of the counterclockwise circulation increases.

h. Intensity of the counterclockwise circulation decreases.

021. Characteristics of baroclinic lows

1. Where does an amplifying short wave extract energy during the baroclinic process?

2. What causes upper-level short wave amplification?

3. How do short waves convert energy into low-level circulations?

4. What can be said about baroclinic instability and the development of midlatitude synoptic-scale systems?

5. What's the relationship between the thermal wave and the contour wave during the baroclinic process?

6. Where is cyclogenesis favored in reference to the long-wave pattern?

7. What's the significance and orientation of a negatively tilted trough?

8. Under what kind of windflow aloft does cyclogenesis typically occur?

9. Explain Petterssen’s rule.

10. What causes the surface low to deepen during the self-development process?

11. How does boundary layer convergence contribute to the self-development process?

12. During the wave initiation stage, what’s associated with the short-wave trough that causes a wave to form on a front?

13. During the wave intensification stage, what causes the upper-level short-wave trough to deepen? What product supports this?

14. During what stage of low development does an occlusion occur?

15. In what stage of development does a low that’s nearly vertically stacked and is continuing to deepen occur?

022. Characteristics of anticyclones

1. What type of circulation is associated with anticyclones in the Northern Hemisphere?

2. How do the size and intensity of anticyclones compare to the size and intensity of cyclones?

3. Which anticyclones have the greatest horizontal and vertical extent?

4. For each of the following characteristics, distinguish whether anticyclogenesis, anticyclolysis, building, or weakening is occurring.

   (1) The central pressure is increasing.

   (2) An anticyclone’s circulation is weakening.

   (3) The central pressure is decreasing.

(4) An anticyclone is forming.

(5) The clockwise circulation increases.

023. Types of pressure systems

1. Match the pressure system in column B with the characteristics in column A. Items in column B may be used once.

Column A	*Column B*
____ (1) Has anticyclonic circulation that tilts toward the warm air with height.	a. Cold barotropic high.
____ (2) Has the warmest temperatures at its center and rarely extends above 10,000 feet.	b. Warm barotropic high.
____ (3) Has the coldest temperatures at its center and has great vertical extent.	c. Baroclinic high.
____ (4) Combination of a frontal wave and an upper-air short-wave trough.	d. Cold barotropic low.
____ (5) Has the coldest temperatures at its center and rarely extends above 10,000 feet.	e. Warm barotropic low.
____ (6) Has the warmest temperatures at its center and great vertical extent.	f. Baroclinic low.

2–3. Fronts

The basic feature of synoptic meteorology is the air mass. We discussed air masses in the last unit. In this section, we'll concentrate on weather fronts. The term *front* refers to a frontal zone, frontal surface, or surface front (depending on the user). For instance, a frontal zone is a transition zone between two air masses of different properties. A frontal surface is a surface of separation of the two different air masses. A surface front is the line of intersection of either a frontal surface or frontal zone with the earth's surface.

As we explore weather fronts, we look at these major subject areas:

- Frontal zone characteristics.
- Classifying fronts.
- Cold frontal weather effects.
- Warm and stationary frontal weather effects.
- Occluded frontal weather.
- Formation, dissipation, frontogenesis, and frontolysis of frontal systems.
- Frontal intensity.

024. 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 these front characteristics:

- Types of fronts.
- Thermal structure.
- Humidity characteristics.
- Vertical wind distribution.
- Horizontal wind distribution.
- 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, cP air moving out of Canada into the central United States is preceded by a cold front. 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 2–26 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 shows the height, strength, and degree of mixing between the air masses.

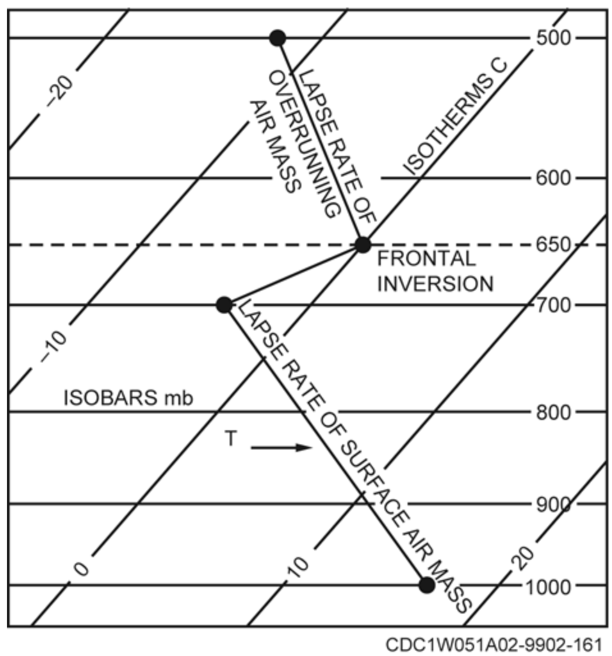

Figure 2–26. Temperature inversion through a front.

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.

Humidity characteristics

Figure 2–27 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 dew-point curve better indicates the frontal zone.

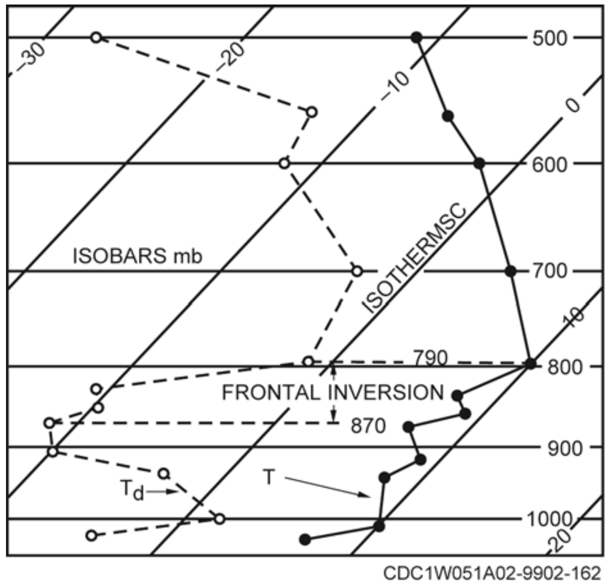

Figure 2–27. Dew-point temperature curves through a frontal zone.

In association with a subsidence inversion, the dew-point 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 2–28 illustrates the characteristics of the dew-point 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 2–28, the dew-point 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.

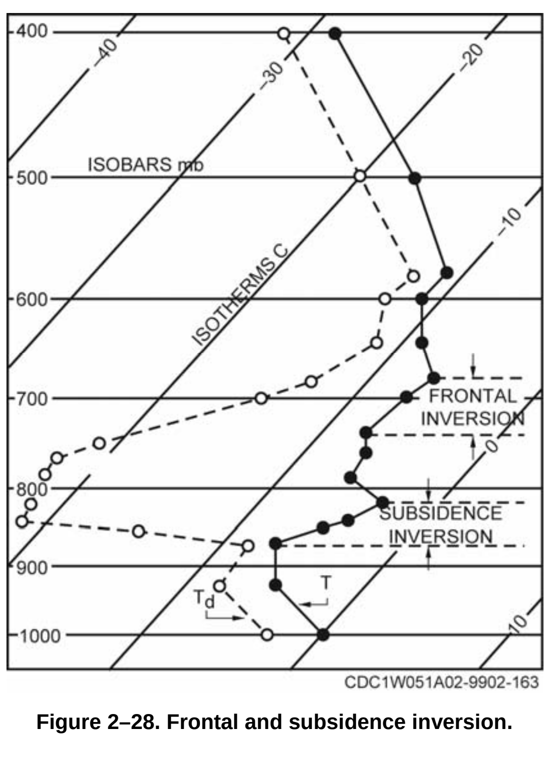

Figure 2–28. Frontal and subsidence inversion.

Vertical wind distribution

The best sources for identifying frontal inversions are a combination of temperature and dew-point curves 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 height through inversions associated with cold fronts and veers (changes clockwise) with height through inversions associated with warm fronts. In figure 2–29, look at the wind changes from the 800- to the 650-mb levels. The winds veer (southwesterly changing to northwesterly) between the two levels; this identifies this as a warm frontal zone.

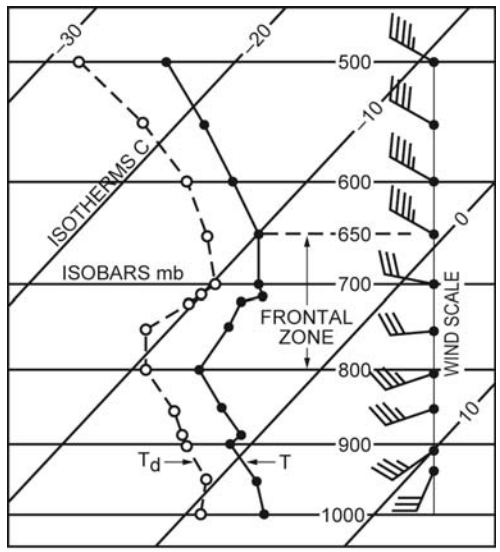

Figure 2–29. Vertical wind distribution through a frontal inversion.

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.

Horizontal wind distribution

Similar to vertical variations, horizontal variation in wind direction and speed also exist across the frontal zone. 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 2–30 illustrates the three different airflow patterns that flow around and through a developing baroclinic low. These different patterns are composed of the following three conveyer belts:

- Warm.
- Cold.
- Dry-air.

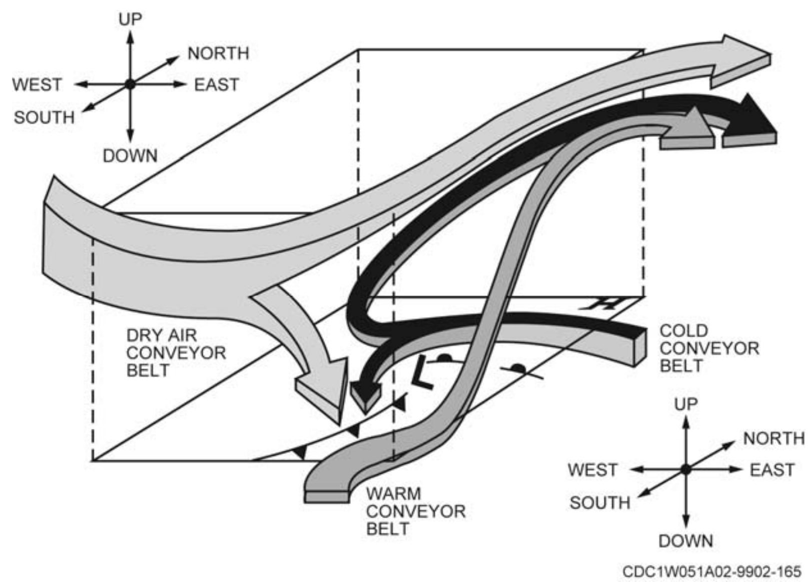

Figure 2–30. Conveyor belts

Warm conveyor belt

The warm conveyor belt is a set of streamlines which 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 (baroclinic zone cirrus is discussed elsewhere) 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 (comma-shaped cloud systems are addressed in the unit on satellite interpretation).

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.

025. Types of fronts and their effects

Fronts are classified by the relative motions of the warm and cold air. The four types of fronts are as follows:

- Cold.
- Warm.
- Stationary.
- Occluded.

Cold

A cold front exists when the cold air displaces the warm air. The windflow in the cold air would be toward the front. The cold air forces the warm air aloft, forming an average frontal slope of 1/30 to 1/100. Cold fronts move faster than warm fronts.

Warm

A warm front occurs when warmer air replaces cold air. Cold air has a horizontal component of flow away from the front. The warm air flows up over the cold air, forming a frontal slope of 1/100 to 1/400. Remember from the resident course that the frontal slope is the ratio of the height of the upper front at some location (rise), to the distance between that location and the surface front (run). In other words: frontal slope = (rise over run).

Stationary

The stationary front shows no significant change between the warm and cold air. Here, the flow of cold air is parallel to the front. This means the front should have no movement.

Occluded

An occluded front occurs when the cold front overtakes the warm front and forces the warm air aloft. There are two types of occluded fronts—warm-frontal and cold-frontal.

Warm-frontal type

The warm-frontal type occurs when the cool air behind the cold front overrides the colder air ahead of the warm front. This results in a cold front aloft.

Cold-frontal type

The cold-frontal type occurs when the cold air behind the cold front lifts the warm front.

Figure 2–31 illustrates these two types of occlusions. The line A-B shows the cold-frontal type and line C-D shows the warm-frontal type.

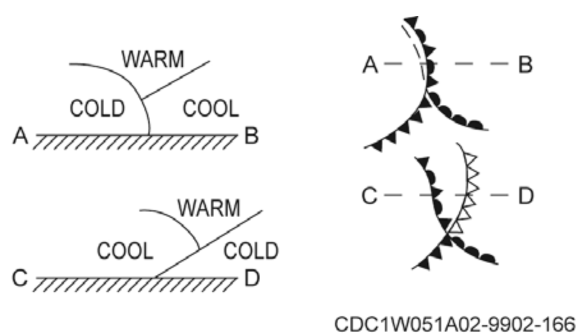

Figure 2–31. Types of occlusions.

Cold frontal weather effects

The type and intensity of frontal weather is determined largely by such factors as the slope of the front, the water vapor content and stability of air masses, the speed of the frontal movement, and the relative motion of air masses at the front. Because of the variability of these factors, frontal weather may range from a minor wind shift with no clouds or other visible weather activity to severe thunderstorms accompanied by low clouds, poor visibility, hail, icing, and severe turbulence.

Weather associated with a frontal passage at one place is frequently very different from that associated with a frontal passage at another place along the same front. This is because when an air mass modifies from below as it moves over a surface, the mixing creates variations in characteristics within the air mass. Therefore, frontal zone characteristics vary along the length of the same frontal boundary.

Because air masses have differing temperature and moisture characteristics, they also have different densities; cold dry air masses are the densest.

When air masses meet, the denser air mass displaces the less dense air mass and pushes it aloft. Thus, less dense warm air slides up over the denser cold air, resulting in a sloping frontal surface. The measure of steepness of the slope is determined by the angle measured between the frontal surface and earth's surface. This concept is illustrated on figure 2–32. The varying slopes of frontal surfaces are due to the dynamics involved with the movement of the associated air mass.

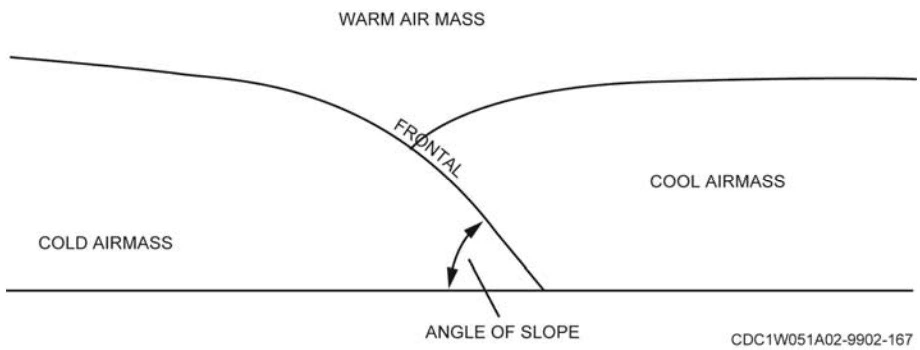

Figure 2–32. Slope of a frontal surface.

We use differences in air-mass properties, such as temperature, moisture, wind, cloud types, and pressure, to find, identify, and track fronts. One of the most easily recognized discontinuities is temperature. At the earth's surface, the passage of a front is most often recognized by a noticeable temperature change. The rate and amount of change are partial signs of frontal intensity. Abrupt and sizable temperature changes accompany strong fronts while gradual or minor decreases in temperature characterize weak or diffuse fronts. With passage of a cold front, the cold air mass displaces a warm air mass at the surface. Cold fronts usually move faster and have a steeper slope than warm fronts.

At the surface, decreases in temperature and dew point, a wind shift, and, occasionally, gusty winds mark a cold frontal passage. Cold frontal weather is found in a somewhat narrow band as opposed to warm fronts, which have widespread weather patterns.

Types of cold fronts

Surface cold fronts are classified as active or inactive, both of which are discussed below.

Active

An active cold front moves at average speeds between 10 to 15 knots. Its slope is about 1/100 miles, as indicated in figure 2–33. The speed and slope are directly related to the perpendicular wind component at the front (between 850 and 700 mb) which decreases with height through the front. Thus, the net flow is upward throughout the front.

This type of front is also called an *anafront* because most of the clouds and weather caused by the upward flow are at and behind the front. The term anafront comes from the word *anabatic*, which means "moving upwards." When the warm air is stable, altostratus and nimbostratus clouds extend several hundred miles behind the front. If the warm air is unstable or conditionally unstable, thunderstorms and cumulonimbus form up to 50 miles or more behind the front.

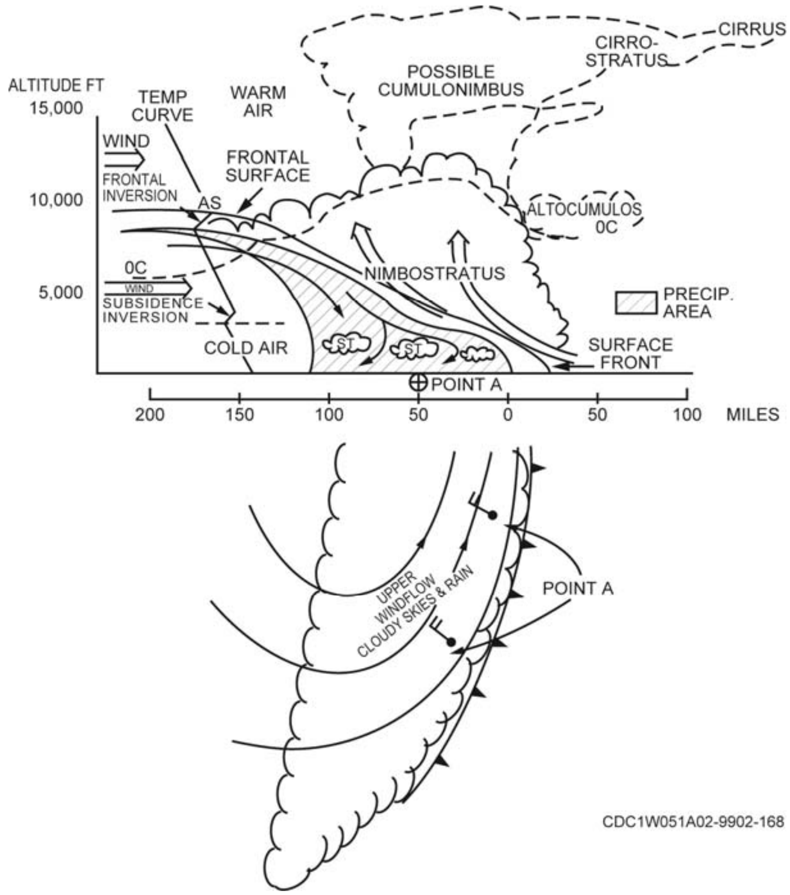

Figure 2–33. Typical vertical structure of an active cold front with upper windflow in back of front.

The ceiling is normally at the height of the frontal surface, since most of the clouds occur in the warm air. The exception is stratus, which may form in precipitation areas near or immediately behind the front in the cold air.

Visibility may be poor near the front, due to fog or precipitation, but usually improves rapidly as the front moves away. Pressures fall steadily before frontal passage and rise weakly behind it. The temperature and dew point drop rapidly behind the front, due to the sharp temperature gradient created by the upward airflow. The wind veers with frontal passage. Maximum velocity occurs at passage of the front and decreases afterward.

Inactive

Figure 2–34 illustrates an inactive type of front. Frontal slope is much steeper than the active front, averaging about 1/40 to 1/80. The wind component perpendicular to the front (between 850 and 700 mb) increases with height which results in a net downward flow of air along the frontal surface. This downward flow helps the front average a movement of about 25 knots. This type of front is called a *katafront*. The term katafront comes from the word *katabatic*, which means “moving downwards.”

The subsidence associated with the downward flow prevents the formation of clouds at or behind the front and visibility increases with frontal passage. However, the downslope air from the front converges with the warm air ahead of the front and produces weather and storms ahead of it. This is discussed more in the section on squall lines. Squall lines usually occur 50 to 200 miles ahead of the front.

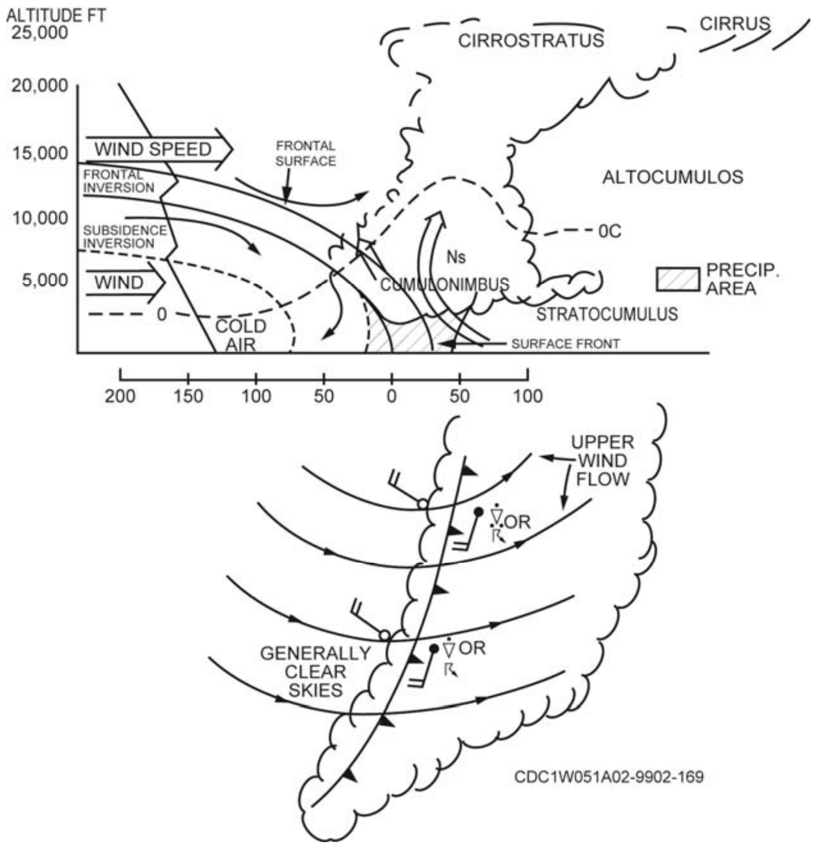

Figure 2–34. Typical vertical structure of an inactive cold front with upper windflow in the front.

The subsidence in the front weakens the temperature gradient across the front. Temperature falls, due to precipitation in the squall line, further weakening the temperature gradient across the front. Maximum temperature drops usually occur far behind the front. Dew point and wind direction are better indicators of frontal passage than temperature. The wind veers with frontal passage and is strong, gusty, and turbulent for a considerable length of time. Dew points decrease sharply with passage. Pressure falls ahead of the front and rises strongly after its passage. Frequently, the squall line is mistaken for the inactive cold front.

Warm and stationary frontal weather effects

As our lesson title implies, we'll be looking at the characteristics of the frontal weather effects for warm and stationary fronts.

Warm fronts

Warm fronts have certain characteristics that distinguish them from other types of fronts. Because of the dynamics associated with retreating colder air, warm fronts have shallow slopes. Two views are shown on figure 2–35. Warm fronts move at an average speed of 10 to 15 knots. A warm front differs from a cold front in that the associated weather pattern may be more extensive. Clouds are usually more stratified and precipitation is continuous.

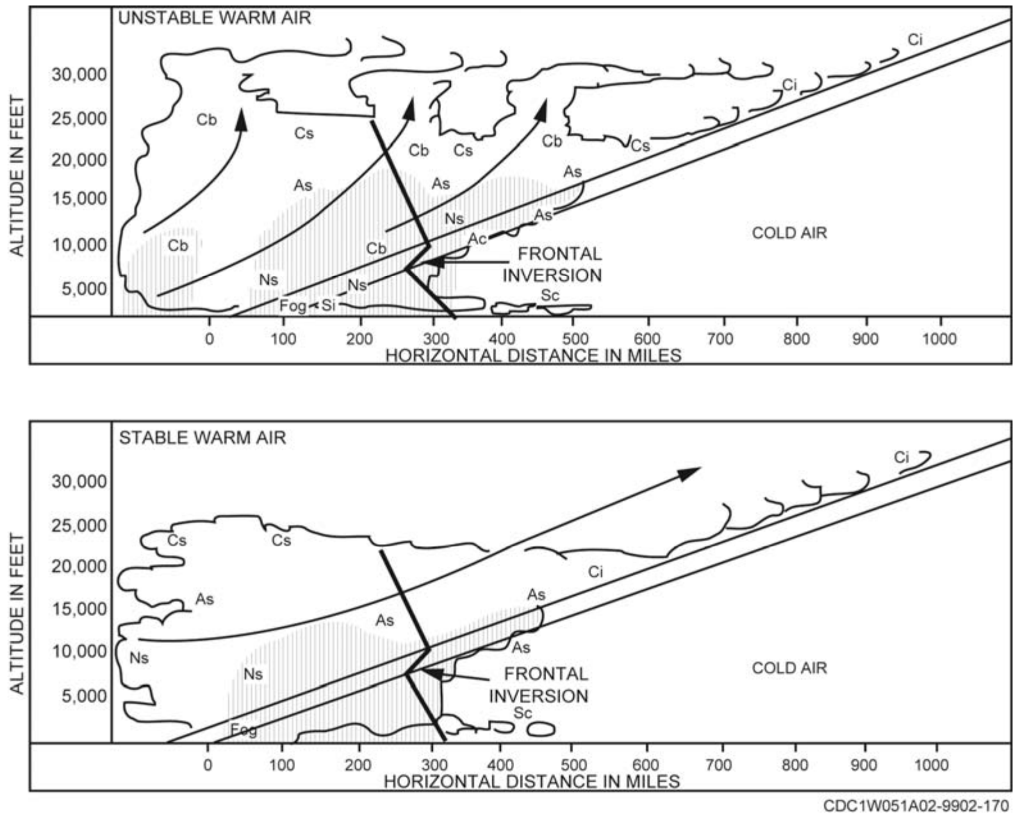

Figure 2–35. Warm fronts with stable and unstable warm air.

Clouds associated with a warm front usually occur in this sequence:

1. Cirrus.
2. Cirrostratus.
3. Altostratus.
4. Nimbostratus.
5. Stratus.

Cirrus and cirrostratus form several hundred miles ahead of the surface warm front. Gradually, the cirrostratus thickens to form altostratus 300 to 500 miles ahead of the front. Closer to the frontal boundary, the altostratus thickens, becoming nimbostratus with its associated precipitation. The amount and type of clouds and precipitation with a warm front vary with the characteristics of the air masses involved.

Warm front with overrunning stable air

When the overrunning warm air is moist and stable, nimbostratus clouds with continuous light to moderate precipitation are found up to 300 miles ahead of the front. The base of the clouds lowers rapidly as more clouds form in the cooler air under the frontal surface. The clouds form by evaporation of the falling rain. When the cooler air mass is stable, the clouds are stratus with fog reducing visibilitys. The clouds in the cooler air are stratocumulus when the cooler air mass is unstable.

The pressure ahead of the warm front usually makes a rapid or unsteady fall with a leveling off after frontal passage. As the cold front approaches, the pressure begins to fall again. Wind speed may increase gradually in advance of warm fronts. The temperature rises ahead of the front and levels off after frontal passage.

Warm front with overrunning unstable air

When overrunning air is moist and unstable, nimbostratus and altostratus clouds frequently have cumulus and cumulonimbus clouds embedded in them. In such cases, heavy rain showers or thunderstorms, which are intense and intermittent, are combined with the continuous precipitation.

Warm front with overrunning dry air

When the overrunning warm air is dry, it must ascend to high altitudes before condensation can occur. Therefore, only high and middle clouds are observed.

Visibility

Visibility, except where the precipitation occurs, is usually good. When the cold air mass is stable, fog may restrict visibility.

Pressure

Pressures ahead of the front usually make a rapid or unsteady fall, with a leveling off after frontal passage. Wind velocity increases ahead of warm fronts, due to the increased pressure gradient. Velocity reaches a maximum just before frontal passage. The wind also veers with frontal passage. Temperature and dew point are usually constant or slowly rising ahead of the front and increase markedly behind it.

On the upper-air products, isotherms are parallel to the front and packed ahead of it. The stronger the packing, the more active is the front.

Stationary front

A front that becomes stationary, or nearly so, is called a *stationary front*. Normally, a front moving less than 5 knots or one that doesn’t move steadily in one direction is considered stationary.

A front becomes stationary when the cooler air mass doesn’t move toward or away from the front. This means the wind blows generally parallel to the front rather than toward or away from it.

Though the front is stationary, there’s still a net inflow of warm air toward the front that causes slow, shallow advancement of air over the frontal surface. As air is lifted and condensed, clouds and precipitation form in the warm air above and on the cold air side of the front. The width of the precipitation band and low ceilings varies from 50 to about 200 miles.

Weather patterns associated with stationary fronts are much like the weather pattern of the warm front. If warm air associated with the stationary front is stable, stratiform clouds predominate with possible drizzle, light rain, or snow. If warm air is unstable, clouds are a mixture of cumuliform and stratiform varieties; precipitation is showery; thunderstorms may occur.

Extensive fog and low ceilings may result within the cold air mass when the warm rain or drizzle falling through the cold air from the warm air mass above saturates the cold air.

Occluded frontal weather

The examples we just discussed involved frontal boundaries with only two air masses. With a third air mass, we must consider other frontal configurations. Remember, these configurations are variations on cold and warm frontal structures we call *occluded fronts*.

Occlusions are the result of an interaction between cold, cool, and warm air masses. They combine weather of the warm and cold front into one extensive system. Specifically, convective activity associated with cold fronts merges with low ceilings and visibilitys associated with the warm front.

Occluded frontal systems are climatologically more common in the northern United States than in the South. The North favors the development of frontal occlusions. They occur frequently in the winter months in the northwest and northeast sections of the country.

There are two types of occluded fronts:

1. Cold occlusions.
2. Warm occlusions.

The type of occlusion is dictated by the relative densities of the air masses. Remember the moisture content of the air mass also affects the air-mass density. If two air masses have equal temperatures, the one with the higher moisture content is the less dense of the two.

Cold occlusions

The views on figure 2–36 illustrate cold-type occlusions. These occur when three air masses are situated so the densest, coldest air is behind the cold front with cool, less dense air ahead of the warm front. Then, as the three air masses come together, the advancing cold front remains on the surface. Thus, it forces both the less dense cool and warm air masses up and over the cold air mass. As the process continues, the surface warm front becomes a warm front aloft. Figure 2–36 also shows a cross section of a cold frontal occlusion that depicts the typical weather and associated cloud patterns. The embedded thunderstorms with the cold occlusion usually occur with passage of the surface occluded front. This is the most common type of occlusion. The name cold occlusion was derived from the front that stays on the surface—here, the cold front.

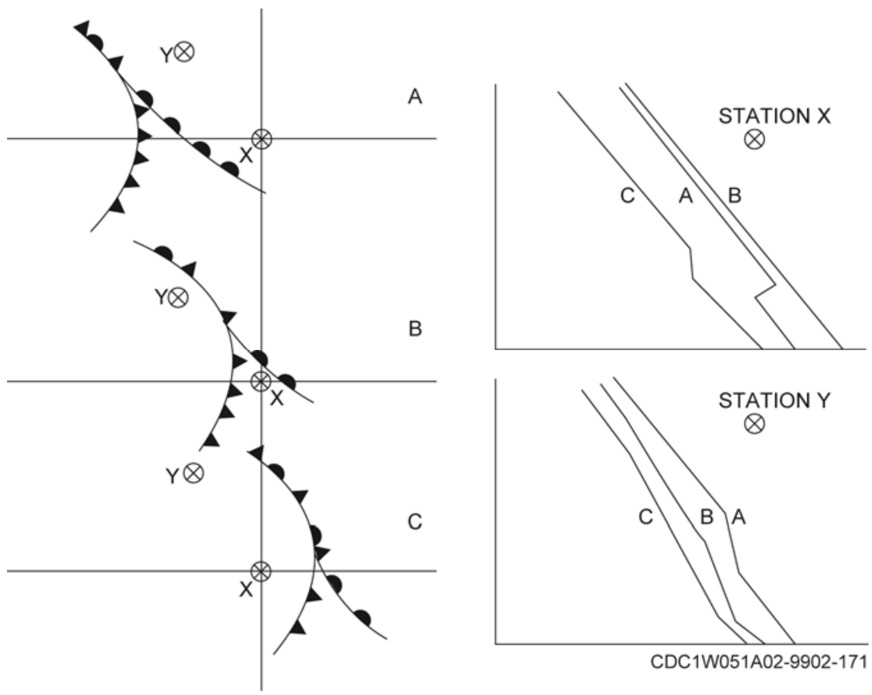

Figure 2–36. Illustration of typical lapse rate changes in a cold occlusion.

Warm occlusions

In a warm-type occlusion, the densest (usually the coldest) air mass is ahead of the warm front with a less dense (cool) air mass behind the cold front. This is illustrated on figure 2–37.

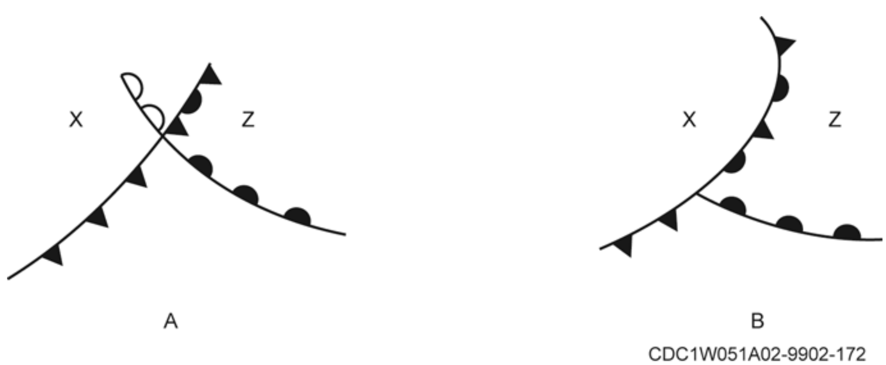

Figure 2–37. Warm-type occlusion.

When the three air masses come together, the densest cold air ahead of the warm front stays at the surface. This forces the advancing less dense, cool air mass aloft. Being the least dense, the warm air is wedged between the cool and cold air masses and is forced aloft, just as it was with the cold frontal occlusion. Since the cool air mass is overrunning the cold, a warm front stays at the surface, forcing the cold front aloft. As with the cold occlusion, the warm occlusion is named for the front remaining at the surface.

The cloud pattern with this system is much wider than with the cold-type occlusion. This is because of extensive overrunning of the cool air over the cold air. This overrunning situation is like that associated with a warm front. A line of thunderstorms with the warm occlusion is often embedded within the stratiform overcast layer and may precede the occlusion and surface warm front by 200 to 300 miles.

Characteristics of a cold frontal occlusion

In the initial stages of a cold occlusion, the weather and cloud sequence ahead of the occlusion resembles a warm front pattern. Near the surface position of the front, the cloud and weather patterns resemble a cold front. As the occlusion continues to lift the warm air higher, the warm front and prefrontal clouds disappear. Now the cold frontal pattern prevails, and most of the precipitation occurs just ahead of the occlusion. Clearing occurs rapidly following passage of the upper warm front or the surface occlusion, depending on the stage of development. Surface pressure drops rapidly ahead of the occlusion and rises rapidly following its passage.

Figure 2–36 shows the typical lapse rate changes in a cold front type occlusion. In the three curves, station X is first in the cool air overrun by the warm air (curve A), then in the warm sector (curve B), and last in the cold, subsiding air (curve C). Station Y is deep within the circulation of the low and the frontal discontinuities are weak due to the convergent lift of the air. Curve A shows station Y in the cool air overrun by the warm air. Curve B shows station Y in cold air overrun by cool air. Curve C indicates the station is in the cold air.

The occlusion usually disappears before it reaches the 700-mb level due to the loss of the warm front portion. As this occurs, the thickness gradient or mean isotherms associated with the surface front weakens. This is depicted in figure 2–37. In view A on figure 2–37, the upper warm front still exists and the gradient between X and Z is strong because of the warm air aloft. In view B on figure 2–37, the upper warm front has disappeared and the gradient has weakened between X and Z because the temperature contrast has decreased with the loss of the warm air. The vertical wind distribution resembles the individual frontal system’s wind distribution.

Characteristics of a warm frontal occlusion

Figure 2–38 illustrates the typical lapse rate changes with the passage of a warm-type occlusion. Station X is first in the cold air overrun by the warm air (curve A), then in the warm sector (curve B), and last in the cool air (curve C). Station Y is first in the cold air overrun by the warm air, next in the cold air overrun by cool air but also with the warm air above this (curve B), and then in the cool air (curve C).

The weather pattern of a warm front occlusion has the characteristics of both warm and cold fronts. Ahead of the occlusion, the sequence of clouds is similar to that of a warm front. The upper cold front may be accompanied by showers and thunderstorms if the warm air is moist and unstable. As the warm air is lifted higher, the weather activity diminishes. Normally, there’s clearing weather after passage of the upper cold front.

The surface pressure shows a steady fall ahead of the upper cold front, leveling off after passage, and a steady rise following passage of the surface occlusion. The temperature rises slightly following passage of the occlusion. Winds in the horizontal and vertical are similar to the individual fronts.

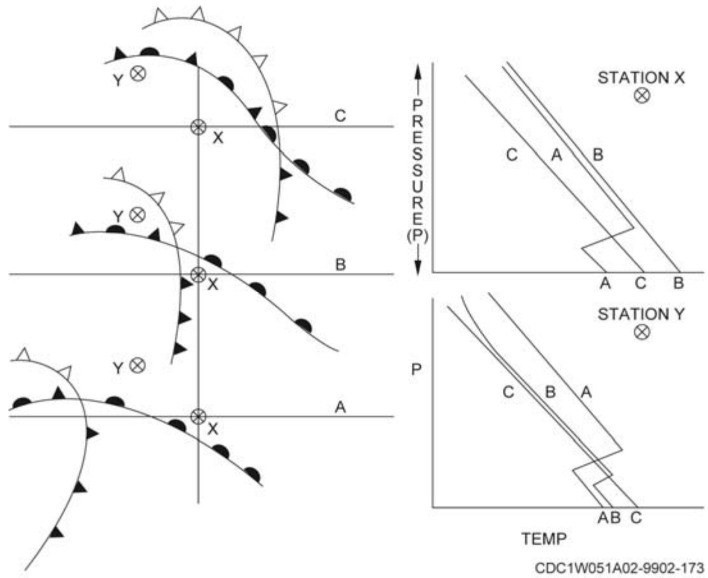

Figure 2–38. Illustration of typical lapse rate changes in a warm occlusion.

026. Formation, dissipation, and intensity of frontal systems

Earlier, we said the term front refers to a frontal zone, frontal surface, or surface front, depending on the user. We said a frontal zone is a transition zone between two air masses of different properties, a frontal surface is a surface of separation of the two different air masses, and a surface front is the line of intersection of either a frontal surface or frontal zone with the earth's surface. In this lesson, we'll cover four subject areas:

- Formation.
- Dissipation.
- Identifying frontogenesis and frontolysis.
- Frontal intensity.

Formation

The formation of a new front or the regeneration of an old front is called *frontogenesis*. This occurs only when these two conditions are met:

1. Two air masses of different densities exist next to one another.
2. A prevailing wind field brings the two air masses together.

Dissipation

The dissipation of a front is called frontolysis. This occurs when either the temperature difference between the two air masses disappears or the wind carries the particles of the air masses away from each other. Frontolytical processes are most effective in the lower layers of the atmosphere. Surface heating and turbulent mixing often quickly modify the air in a frontal zone, which eliminates the temperature contrast. Therefore, fronts are often more easily identified above the friction layer or with a thickness analysis.

Identifying frontogenesis and frontolysis

Frontogenesis and frontolysis occur under different conditions. In this lesson segment, we'll examine the ways to identify them. To do this we'll explore the following subjects:

- Low-level deformation zones.
- Convergence and divergence.
- Boundary-layer processes.
- Diabatic processes.

Low-level deformation zones

The distribution and concentration of isotherms within a deformation field consist of a *col* area of flat pressure between two opposing highs and two opposing lows.

Cols are favored areas for frontogenesis because the horizontal motions of the atmosphere in these areas contribute to sharp horizontal temperature gradients. Consider, for example, the circulation pattern shown in figure 2–39 to represent a col between opposing highs and lows. The flow along line B-B tends to move the isotherms toward the neutral point at the center of the col. This flow of maximum contraction along a common axis is known as the *axis of contraction*. This shrinking (or contraction) is complemented by a stretching (or dilatation) along line A-A, which is known as the *axis of dilatation*. The isotherms tend to become concentrated and parallel to line A-A.

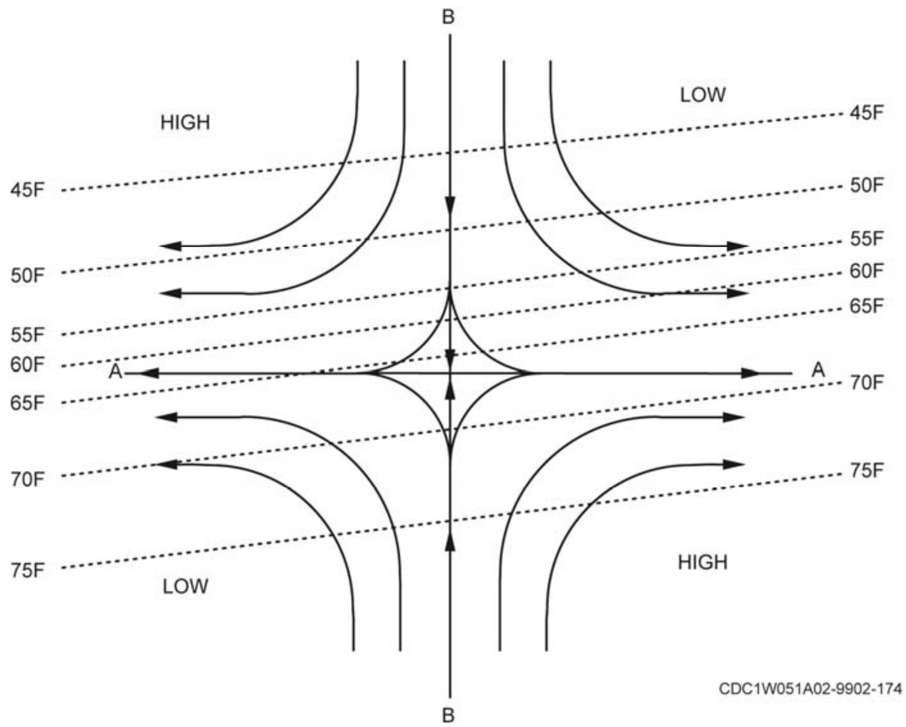

Figure 2–39. Deformation field indicating frontogenesis.

Petterssen showed that the angle between the axis of dilatation and the isotherm packing is important in determining whether frontogenesis will develop. If the angle between the axis of dilatation and the isotherm ribbon is less than 45°, like in figure 2–39, frontogenesis will likely occur. A warm or cold front can develop along the axis of dilatation if the conditions for frontogenesis are satisfied.

The axis of contraction is easily seen behind a moving cold front. You just simply look for the delineation between the cyclonically and anticyclonically curving isobars as seen in figure 2–40. What isn't apparent is the axis of contraction on the warm side of the front. Because the system is moving, the direction of the inbound parcels on the warm side of the front is masked. The axis of dilatation is parallel to the front and, therefore, the isotherms. This type of deformation zone is responsible for the constant reinforcement of the cold frontal boundary. Without it, the cold front would undergo rapid frontolysis due to diabatic effects (heated from below).

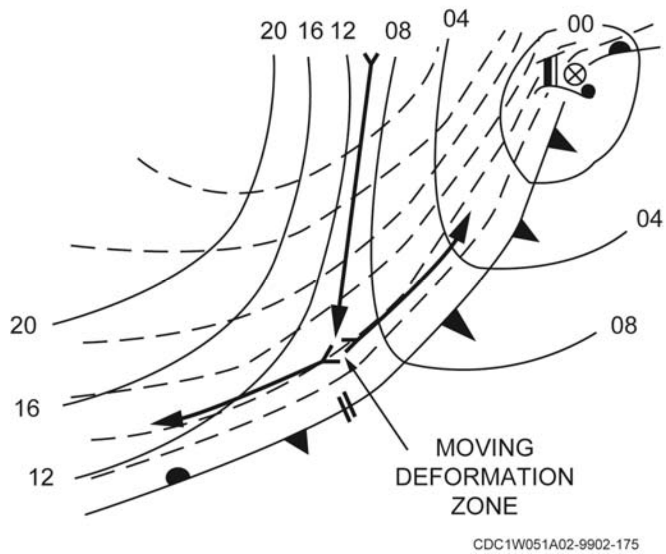

Figure 2–40. Moving deformation zone behind a cold front.

Another deformation zone pattern is the frontolytic pattern shown on figure 2–41. In this pattern, the axis of contraction (line B-B) and the isotherms are parallel and the axis of dilatation (line A-A) is perpendicular. This would result in the isothermal packing pattern being pulled apart by the axis of dilatation.

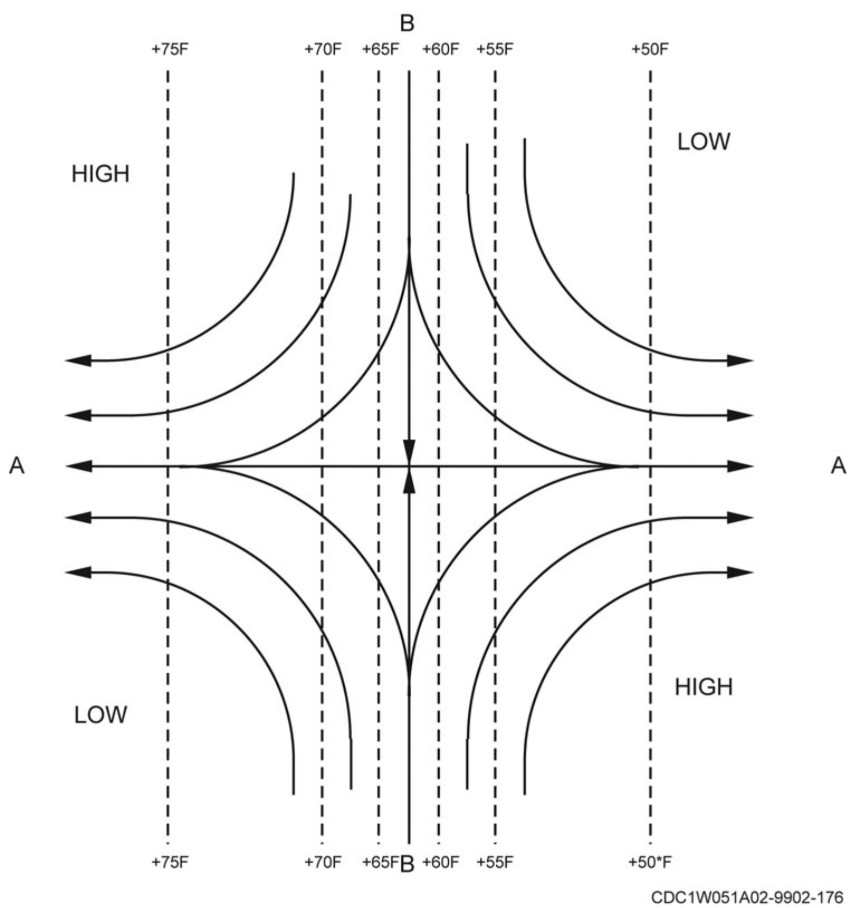

Figure 2–41. Ideal frontolytic pattern.

The above two examples are extreme. A vast majority of the deformation zones identified on products possess some degree of angle between the axes and the isotherms. The easiest way to remember this is to reference this angle difference to either the axis of dilatation or the axis of contraction. For ease of memorization (and this author's preference), we stick with the axis of dilatation. When the angle between the axis of dilatation and the isotherms are greater than 45°, frontolysis is favored. Conversely, if the angle is less than 45°, then frontogenesis is favored. Angles of 45° show doubtful frontogenesis, or even frontolysis. Figure 2–41 illustrates the isotherm and windflow pattern most favorable for frontolysis.

Convergence and divergence

The next factors we look at are convergence and divergence. Low-level convergence (divergence) always supports frontogenesis (frontolysis). However, it doesn't guarantee frontogenesis (frontolysis). Upper-level dynamics associated with the polar-front jet are critically important for frontogenesis and frontolysis. Divergence (convergence) aloft, as shown by PVA (NVA) on the vorticity product, is associated with low-level convergence (divergence) and supports frontogenesis (frontolysis).

Boundary-layer processes

Cyclonically (anticyclonically) curved flow combined with the friction found in the boundary layer leads to low-level convergence (divergence). Since surface fronts are found in regions of cyclonic turning of the wind, this effect helps maintain the front. A sharpening of the cyclonic turning strengthens the front. A front moving into a region where the low-level wind turns anticyclonically undergoes frontolysis.

Diabatic processes

The final phenomena affecting the frontal intensity are diabatic processes. A diabatic process is one that involves the exchange of energy (heat) between air parcels and their environment—the opposite of an adiabatic process. Recall from the resident course that in an adiabatic process, no heat or mass is exchanged between the parcel and its environment. The main diabatic effect we're concerned with is the transfer of energy between the earth's surface and the low levels of the atmosphere.

Frontogenesis is supported if some mechanism heats the warm air or cools the cold air. This is the primary mechanism by which air masses form; this effect is common in air mass source regions. Outside air-mass source regions, this situation is uncommon. One example of this is when the warm air ahead of a cold front is heated further by the warm waters off the East Coast.

Frontolysis is supported if some mechanism cools the warm air or heats the cold air. This is the typical case outside air-mass source regions. The warm air behind a surface warm front moves over a relatively cool land mass (previously occupied by cold air) and cools. The cold air behind a surface cold front moves over a relatively warm land mass (previously occupied by warm air) and warms. This effect is further enhanced because as cold air moves equatorward, it becomes shallow. The shrinking in the vertical (subsidence) causes adiabatic warming. A shallow layer of cool air is modified rapidly over a warm surface.

Any combination of these factors (low-level deformation zones, convergence/divergence, boundary-layer processes, and diabatic effects) may be present. If all of the factors are supporting frontogenesis (frontolysis), then a significant change in frontal intensity or the development (dissipation) of a front occurs. If some factors oppose one another, then the change in frontal intensity is small and difficult to estimate. Continuity is essential for tracking the intensity of fronts. Monitoring the lapse rate in the stable layer and the thickness/temperature contrast across the transition zone reveals important changes in the strength of the front. The strength of the front plays a large role in determining the conditions around the surface front. The stronger the front, the more likely significant weather is to occur along it.

Frontal intensity

Frontal intensity is a measure of the strength of the density contrast across the transition zone. This density contrast can be seen in the thickness gradient on a thickness product, temperature gradient of a front, and lapse rate across the front. The thickness gradient is the first property we investigate.

Thickness gradient

The thickness gradient is the difference between the thickness on the warm and cold sides of the front. It's also a measure of the density contrast across the front. A weak front has a weak thickness gradient (widely spaced thickness lines) in the transition zone. A strong front has a strong thickness gradient (closely spaced thickness lines) in the transition zone.

Temperature gradient

Examining the difference between the temperature on the warm and cold sides of the front (temperature gradient) is the second way to measure the density contrast across the front. This is illustrated on figure 2–42. The temperature gradient is evident at various levels (surface, 850mb, etc.). A weak front has a weak temperature gradient (widely spaced isotherms) in the transition zone. A strong front has a strong temperature gradient (closely spaced isotherms) in the transition zone.

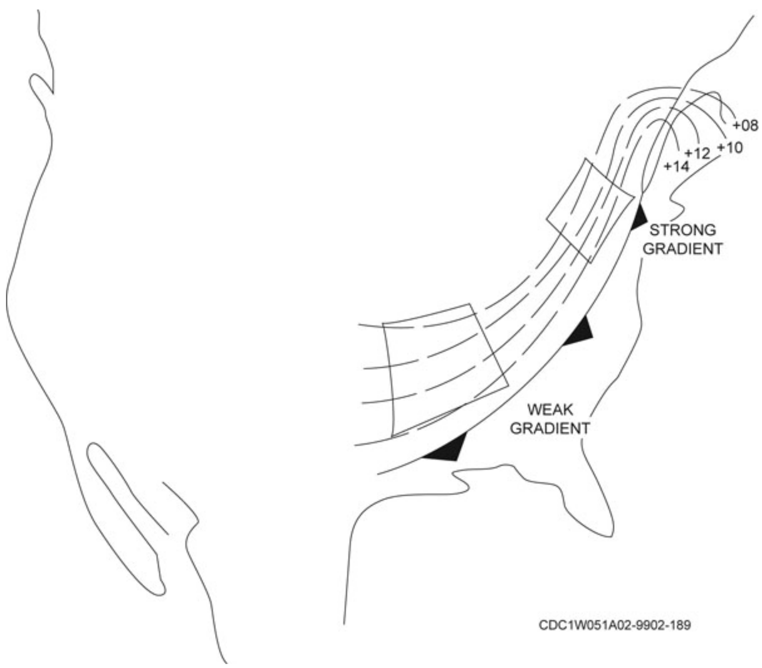

Figure 2–42. Temperature gradient indicating frontal intensity.

This is the parameter we most frequently use in finding the intensity of fronts. In practice, the temperature gradient for determining frontal intensity is the difference between the representative warm air immediately next to the front and the representative surface temperature 100 miles from the front on the cold air side. By convection, the transition zone is taken to be part of the cold air mass.

A suggested set of criteria based on horizontal temperature defines a weak front as having a gradient less than 10°F per 100 miles, a moderate front as when the gradient is 10 to 20°F per 100 miles, and a strong front as where the gradient is greater than 20°F per 100 miles.

Lapse rate

The third property used to evaluate the density contrast across a front is the lapse rate through the stable layer (frontal inversion) on a Skew-T. Remember that the lapse rate is positive if the temperature decreases with height. A weak front has a small positive lapse rate. The temperature decreases slowly with height. A strong front has an inversion (negative lapse rate). The temperature increases with height.

Not only are density differences across a front indicative of frontal intensity, you can also use turbulence and wind shear as an indicator of it.

Turbulence

A front may be intense with discontinuity across it, but may have no weather except strong winds and a change in temperature. A front, normally classified as *weak*, is considered moderate if turbulence and gustiness occur along it. The term *gustiness* includes thunderstorms and strong winds despite the amount of wind shear.

Wind shear

Wind shear may be either the vector difference between the surface-geostrophic wind components parallel to and immediately on either side of the front or the 1,000–500mb thermal wind shear. The greater the geostrophic wind shear, the more intense is the front. Convert thermal wind shear into frontal intensity by using the following relationships. If the thermal wind shear is:

1. Equal to or less than 25 knots, no front exists or frontolysis is in progress.
2. Greater than 25 knots but equal to or less than 50 knots, it’s a weak front or frontogenesis is in progress.
3. Greater than 50 knots but less than or equal to 75 knots, the front is of moderate intensity.
4. More than 75 knots, the front is strong.

Self-Test Questions

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

024. Frontal zone characteristics

1. Place a check mark by accurate statements concerning elements expected in the frontal zone.
   a. Lapse rate through the frontal zone shows a strong inversion and dew-point curve decreases rapidly through the inversion.
   b. Lapse rate through the frontal zone shows only slight decrease, but the dew-point curve increases sharply.
   c. Lapse rate through the frontal zone decreases sharply and dew-point curve increases sharply.
   d. A vertical sounding through the frontal zone may show a subsidence inversion above the frontal inversion.
   e. When precipitation falls through a subsidence inversion, the dew point may be raised enough so that the inversion appears as a frontal inversion.
   f. The winds veer with height through a cold front.
   g. The winds veer with height through a warm front.
   h. The winds back with height through a cold front.
   i. The winds back with height through a warm front.
   j. The surface winds back with passage of a cold front.
   k. The surface winds veer with passage of a warm front.
   l. The surface wind speeds are usually the greatest ahead of both warm and cold fronts.
   m. The surface winds are usually greatest behind both warm and cold fronts.

2. What conveyor belt is a set of streamlines that originates at low levels in the moist tropical air mass?

3. What conveyor belt originates in the low levels and is associated with subsidence well ahead of the low center?

4. Which conveyor belt originates at the upper levels?

025. Types of fronts and their effects

1. Classify the frontal type for each of the following characteristics:
   a. Slope of 1/75.
   b. Warm air replacing cold air.
   c. Cold front aloft; warm front on surface.
   d. Slope of 1/200.
   e. Warm front aloft; cold front on surface.
   f. Cold air replacing warm air.
   g. Cold air is parallel to the front.

2. Name the five factors that determine the type and intensity of frontal weather.

3. Why does weather associated with a frontal passage differ from one place to another along the same front?

4. For each of the following characteristics, classify the frontal system associated with it as active or inactive:
   a. Slope 1/40 to 1/80.
   b. Net windflow is up the frontal slope.
   c. Net windflow is down the frontal slope.

d. Katafront.

e. Anafront.

f. Line of thunderstorms 100 miles ahead of front.

g. Thunderstorms and rain in immediate vicinity of front only.

h. Sharp temperature gradient with front.

i. Dew point and winds are best indicators of frontal passage.

5. At what speeds do warm fronts generally move?

6. How does the width of the warm frontal band of weather differ from that of either the inactive or active cold fronts?

7. Describe the situation in which thunderstorms develop with a warm front.

8. Define the term *stationary front*.

9. Weather associated with a stationary front is similar to the weather found with what other type of front?

10. With the approach of a warm front, what change should occur to the pressure, wind, temperature, and dew point?

11. On the 850-mb product, what’s the relationship between the warm front and the isotherms?

12. Which regions of the United States have the most frontal occlusions?

13. In what season of the year are occluded fronts most common in the United States?

14. If the air behind the cold front is colder than the air ahead of the warm front, what type of occlusion will occur?

15. What type of occlusion occurs when the air behind the cold front is warmer than the air ahead of the warm front?

16. Where are the embedded thunderstorms with a cold frontal occlusion normally located?

17. Which type of occluded front normally has the wider cloud system?

18. What type of clouds and weather normally precede a cold occlusion?

19. With a cold occlusion passage, what will the pressure do?

20. As the occlusion process dissipates the warm front, what happens to the thickness gradient associated with the surface front?

21. If a pilot descends vertically through a cold-type occlusion to pass through both fronts, how will the wind switch?

22. In a well-developed warm occlusion, what's the relationship between the cold front portion and the warm front?

23. If showers and thunderstorms occur, where will they usually be located in relation to the front?

24. Select the characteristics usually occurring with warm-type occlusions:
    a. Increased cloudiness as the upper front approaches.

    b. Increased cloudiness with surface front passage.

    c. Greatest clearing of clouds following upper front passage.

d. Greatest clearing of clouds following surface front passage.

e. Steady rise of pressure following upper front.

f. Steady rise of pressure following surface front.

g. Surface temperature rises following surface front passage.

h. Surface temperature falls following surface front passage.

25. If a pilot descends vertically through a warm-type occlusion to pass through both fronts, how does the wind direction change?

026. Formation, dissipation, and intensity of frontal systems

1. Indicate whether each of the following statements refers to frontogenesis or frontolysis:
   a. Formation of a new front.
   b. Dissipation of a front.
   c. Intensification of a front.
   d. Requires two air masses of different densities.
   e. Prevailing wind brings the two air masses together.

2. Why are cols favored areas for frontogenesis?

3. What will a sharpening of the cyclonic turning do to a front?

4. How do diabatics support frontogenesis?

5. How do diabatics support frontolysis?

6. Name the three ways to measure frontal intensity by means of density contrast.

7. What's indicated by a negative lapse rate with a cold front in the vicinity?

8. What are two other ways to identify the strength of a front outside of density differences?

9. Classify the intensity of fronts having the following characteristics:
   a. Temperature gradient 15°F/100 miles and thermal wind shear 65 knots.

   b. Temperature gradient 8°F/100 miles and thermal wind shear 30 knots.

   c. Temperature gradient 25°F/100 miles and thermal wind shear 85 knots.

Answers to Self-Test Questions

018

1. Horizontal divergence refers to the spreading out of air. When this occurs, the air moves away from the center of the column. The original column of air contracts vertically and expands horizontally. In contrast, horizontal convergence refers to the packing or bringing together of air. The air converges horizontally toward the center of the column. The original column of air contracts horizontally and expands vertically.
2. The pressure at the surface directly relates to the mass of air in the vertical column above the surface. The surface pressure measures the net effect of the convergence and divergence.

019

1. Excess or net divergence aloft for lows. Excess or net convergence aloft for highs.
2. a. It deepens the upper-level low/trough.
   b. It builds the upper-level high/ridge.
3. Net divergence; decreases.
4. Net convergence; increases.

020

1. An area of closed counterclockwise circulation occurring on a frontal surface.
2. A stable wave has the amplitude decreasing or remaining the same with time and is usually filling or showing no change in intensity. An unstable wave has the amplitude increasing with time and is usually deepening.
3. a. Cyclogenesis.
   b. Cyclolysis.
   c. Cyclogenesis.
   d. Cyclolysis.
   e. Cyclolysis.
   f. Cyclogenesis.
   g. Cyclogenesis.
   h. Cyclolysis.

021

1. From the north/south temperature gradient produced by differential heating.
2. Large energy transfer (by means of thermal advection) from the temperature gradient to the wave.

3. By converting potential energy (transferred from the temperature gradient to the short wave) to kinetic energy. The short wave uses the potential energy to develop the low-level circulation (kinetic energy).
4. It's the primary mechanism responsible for the development of midlatitude synoptic-scale systems.
5. The thermal wave and contour wave are out-of-phase.
6. At and just downstream from a long-wave trough axis.
7. They cause stronger divergence and, therefore, support stronger cyclogenesis.
8. Difluent flow aloft.
9. Cyclogenesis occurs when and where an area of upper-level divergence (PVA) becomes superimposed over a low-level frontal zone across which the thermal advection is weak.
10. Divergence over the surface low.
11. By acting as a breaking mechanism.
12. Divergence.
13. Cold-air advection; 500-mb product.
14. Mature wave stage.
15. Mature wave stage.

022

1. A closed, clockwise wind circulation.
2. Anticyclones are greater in size and generally of less intensity than cyclones.
3. Those anticyclones associated with the semipermanent highs.
4. (1) Anticyclogenesis.
   (2) Anticyclolysis.
   (3) Anticyclolysis.
   (4) Anticyclogenesis, building.
   (5) Anticyclogenesis, building.

023

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

024

1. You should have checked these statements: c, e, g, h, k, and m.
2. Warm conveyor belt.
3. Cold conveyor belt.
4. Dry-air conveyor belt.

025

1. a. Cold front.
   b. Warm front.
   c. Warm occlusion.
   d. Warm front.
   e. Cold front occlusion.
   f. Cold front.
   g. Stationary.

2. a. The slope of the front.
   b. The water vapor content.
   c. Stability of the air masses.
   d. Speed of the front.
   e. The relative motion of air masses at the front.
3. Because air mass modifications and mixing create variations in characteristics within the air mass.
4. a. Inactive.
   b. Active.
   c. Inactive.
   d. Inactive.
   e. Active.
   f. Inactive.
   g. Active.
   h. Active.
   i. Inactive.
5. 10 to 15 knots.
6. Weather occurs up to a several hundred mile wide band ahead of the surface warm front; even the active cold frontal band of weather is much narrower.
7. If the overrunning air is unstable, embedded thunderstorms are likely with the warm front.
8. A front moving less than 5 knots steadily in any direction.
9. It basically resembles warm front weather, but is in a somewhat narrower band.
10. Pressure falls rapidly or unsteadily, wind velocity increases, temperature and dew point remain constant or rise slowly.
11. The isotherms will be packed ahead of the front and parallel to it.
12. The North—both the northwest and northeast portions.
13. During the winter months.
14. A cold frontal occlusion.
15. A warm frontal occlusion..
16. The embedded thunderstorms occur with the passage of the surface occluded front.
17. A warm frontal occlusion.
18. Warm frontal clouds and precipitation.
19. Rise rapidly following passage.
20. Thickness gradient decreases.
21. The winds will first back, and then veer.
22. The cold front will be aloft ahead of the warm front.
23. Ahead of and with the upper cold front portion.
24. Characteristics usually occurring are: a, c, f, and g.
25. First it will veer, and then back.

026

1. a. Frontogenesis.
   b. Frontolysis.
   c. Frontogenesis.
   d. Frontogenesis.
   e. Frontogenesis.
2. Because the horizontal motions of the atmosphere in these areas contribute to sharp horizontal temperature gradients.

3. Assists in maintaining or frontogenesizing the front.
4. By heating the warm air or cooling the cold air.
5. By cooling the warm air or heating the cold air.
6. Thickness gradient, temperature gradient, and lapse rate.
7. This indicates that the cold front is strong.
8. Turbulence and wind shear.
9. a. Moderate.
   b. Weak
   c. Strong.

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.

Do not return your answer sheet to Extension Course Program (A4L).

38. (018) Surface pressure changes are *largely controlled* by
    a. mass changes in the upper troposphere.
    b. the jetstream pattern in the upper troposphere.
    c. temperature advection in the middle troposphere.
    d. positive vorticity advection in the lower troposphere.

39. (018) Horizontal divergence within an air mass will
    a. cause the surface pressure to rise.
    b. increase the mass in the vertical column above the surface.
    c. increase the vertical extent of the column above the surface.
    d. vertically contract the original column of air and then expand it horizontally.

40. (018) As air converges at the surface and toward the center of a layer, there will be horizontal
    a. and vertical expansion.
    b. contraction and vertical expansion.
    c. expansion and vertical contraction.
    d. contraction and vertical contraction.

41. (018) An *increase* of mass in a column of air will cause the surface pressure to
    a. increase.
    b. decrease.
    c. remain steady.
    d. do nothing, pressure is not affected by mass.

42. (019) In the chimney effect, the *maximum* upward vertical wind motion will be located
    a. at the tropopause.
    b. at the surface of the earth.
    c. throughout the column of air.
    d. at the level of nondivergence (LND).

43. (019) The *primary* cause of surface pressure changes for a dynamic low is net
    a. divergence aloft.
    b. adiabatic cooling aloft.
    c. adiabatic warming aloft.
    d. divergence at the surface.

44. (019) The damper effect is comprised of upper-level
    a. divergence and surface low pressure.
    b. divergence and surface high pressure.
    c. convergence and surface low pressure.
    d. convergence and surface high pressure.

45. (020) A low-pressure system undergoing cyclogenesis is said to be
    a. dissipating or deepening.
    b. forming or deepening.
    c. dissipating or filling.
    d. forming or filling.

46. (020) When the central pressure of an anticyclone is rising, you can infer the anticyclone is
    a. filling.
    b. building.
    c. deepening.
    d. weakening.

47. (020) A high-pressure system undergoes anticyclolysis when the clockwise circulation area
    a. decreases or disappears.
    b. increases or disappears.
    c. decreases or forms.
    d. increases or forms.

48. (020) An *unstable* wave cyclone is one where the amplitude
    a. decreases with time and the wave fills.
    b. increases with time and the wave fills.
    c. decreases with time and the wave deepens.
    d. increases with time and the wave deepens.

49. (020) After warm air with an unstable wave is pushed aloft and cuts off from a cyclone, the cyclone will become
    a. baroclinic and begin to deepen.
    b. barotropic and begin to deepen.
    c. baroclinic and begin to fill.
    d. barotropic and begin to fill.

50. (020) Unstable waves are classified as
    a. barotropic highs.
    b. barotropic lows.
    c. baroclinic high.
    d. baroclinic lows.

51. (021) In the baroclinic instability process,
    a. kinetic energy is transferred to the major short wave by thermal advection.
    b. potential energy is transferred to the major short wave by thermal advection.
    c. kinetic energy is transferred to the major short wave by low-level circulations.
    d. potential energy is transferred to the major short wave by low-level circulations.

52. (021) When using Petterssen's rule, the factors needed for cyclogenesis are upper-level
    a. convergence and a frontal zone where thermal advection is weak.
    b. convergence and a frontal zone where thermal advection is strong.
    c. divergence and a frontal zone where thermal advection is weak.
    d. divergence and a frontal zone where thermal advection is strong.

53. (021) What causes the surface low to deepen during the self-development process?
    a. Divergence aloft.
    b. Convergence aloft.
    c. Cold-air advection.
    d. Warm-air advection.

54. (021) A low dissipates after proceeding through its entire life cycle which ranges from the
    a. top down as boundary layer convergence removes mass from the column of air.
    b. top down as boundary layer divergence removes mass from the column of air.
    c. bottom up as boundary layer convergence adds mass to the column of air.
    d. bottom up as boundary layer divergence adds mass to the column of air.

55. (021) How many stages are there in the life cycle of a low?
    a. Two.
    b. Three.
    c. Four.
    d. Five.

56. (021) In the life cycle of a low, the system will evolve into a cold barotropic low in the
    a. mature stage.
    b. dissipation stage.
    c. wave initiation stage.
    d. wave intensification stage.

57. (022) Anticyclogenesis *typically occurs* at, and just downstream from, long-wave
    a. ridges under difluent flow aloft.
    b. troughs under difluent flow aloft.
    c. ridges under confluent flow aloft.
    d. troughs under confluent flow aloft.

58. (022) What causes a surface high to build during the self-development process?
    a. Negative vorticity advection.
    b. Positive vorticity advection.
    c. Convergence aloft.
    d. Divergence aloft.

59. (022) Which is *primarily* responsible for low-level divergence acting as a braking mechanism for a high?
    a. Friction.
    b. Coriolis force.
    c. Centrifugal force.
    d. Pressure gradient force.

60. (023) A warm barotropic high is a type of pressure system that
    a. has great vertical extent and its axis tilts with height.
    b. has great vertical extent and is usually found over water areas.
    c. rarely exceeds 8,000 feet vertically and its axis tilts with height.
    d. rarely exceeds 8,000 feet vertically and is found over water areas.

61. (023) In regards to types of pressure systems, a heat low is a
    a. cold barotropic low that forms during the winter.
    b. cold barotropic low that forms during the summer.
    c. warm barotropic low that forms during the winter.
    d. warm barotropic low that forms during the summer.

62. (024) With the passage of a cold front in the Northern Hemisphere, the horizontal wind direction will
    a. become more variable.
    b. become more northeasterly.
    c. shift in a clockwise direction.
    d. shift in a counterclockwise direction.

63. (024) The cold conveyor belt originates in the low levels that are located
    a. west of a low center and flow westward.
    b. east of a low center and flow westward.
    c. west of a low center and flow eastward.
    d. east of a low center and flow eastward.

64. (025) The *average* slope of a cold front is
    a. 1/30 to 1/100.
    b. 1/65 and 1/250.
    c. 1/100 and 1/65.
    d. 1/250 and 1/50.

65. (025) A warm frontal occlusion occurs when the cool air
    a. ahead of the cold front lifts the cooler air ahead of the warm front.
    b. behind the cold front lifts the cooler air ahead of the warm front.
    c. behind the cold front overrides the colder air ahead of the warm front.
    d. ahead of the warm front overrides the colder air behind the cold front.

66. (025) In comparison to an active cold front, an *inactive* cold front is characterized by a relatively
    a. steeper slope, a wide weather pattern, and is called anafront.
    b. shallower slope, a wide weather pattern, and is called anafront.
    c. steeper slope, a narrow weather pattern, and is called katafront.
    d. shallower slope, a narrow weather pattern, and is called katafront.

67. (025) When an *inactive* front passes your station, the dew points will
    a. decrease slowly with the passage.
    b. decrease sharply with the passage.
    c. increase slowly with the passage.
    d. increase sharply with the passage.

68. (025) When only high and middle clouds are associated with a warm front, the
    a. underlying cold air is dry.
    b. underlying cold air is moist.
    c. overrunning warm air is dry.
    d. overrunning warm air is moist.

69. (025) In a cold occlusion, the coldest air is found
    a. behind the cold front.
    b. behind the warm front.
    c. ahead of the cold front.
    d. ahead of the warm front.

70. (025) In a cold occlusion, which type of front, if any, is found aloft?
    a. The occlusion.
    b. The warm front.
    c. The cold front.
    d. All fronts are at the surface.

71. (026) Frontogenesis requires two adjacent air masses with
    a. the same density and a wind flow separating the two air masses.
    b. different densities and a wind flow separating the two air masses.
    c. the same density and a wind flow to bring the air masses together.
    d. different densities and a wind flow to bring the air masses together.

72. (026) The frontolytic processes are *most* effective
    a. on the east coast of continents.
    b. on the west coast of continents.
    c. in the lower layers of the atmosphere.
    d. in the upper layers of the atmosphere.

Please read the unit menu for unit 3 and continue ➔

Unit 3. Sky Condition and Visibility

THE EVALUATION of clouds and visibility is probably the most significant aspect of a weather observation. As a weather forecaster apprentice, you must know 27 states of the sky and how each affects the weather. Sometimes, the sky condition and visibility determine whether an aircraft can land at a base or whether the pilot must seek an alternate airfield. In this unit, we explore the principles of weather observing. Our topics will include the states of the sky, sky condition, and visibility and runway visual range (RVR).

3–1. States of the Sky

At some time, you've probably looked up at the clouds in the sky and remarked. "It looks like rain." You're not alone. For centuries, clouds have been called the signposts of weather. In fact, clouds occurring in sequence describe a weather event much as chapters of a book reveal a story. For instance, the changes from cirrus (CI) to cirrostratus (CS) to altostratus (AS) clouds warn of an approaching warm front.

These cloud messages aren't difficult to understand, but they aren't obvious. Becoming a good cloud reader requires study and experience. To interpret clouds, we consider their formation and classification. Meteorologists categorize all clouds into the 10 basic types shown in this table.

1. Cumulus	6. Cumulonimbus
2. Stratocumulus	7. Stratus
3. Altostratus	8. Nimbostratus
4. Altocumulus	9. Cirrus
5. Cirrostratus	10. Cirrocumulus

Each of these basic types may be further classified into subtypes. The subtypes are recognized internationally as 27 states of the sky—arranged as low, middle, and high clouds. Each state of the sky possesses a distinguishing feature to separate it from the others. This feature may be the height, form, or appearance of the cloud. These distinguishing features provide the clues that signal approaching weather. Use either a copy of the *World Meteorological Organization International Cloud Atlas* or *United Kingdom Meteorological Office Cloud Types for Observers* along with this section to assist you in the recognition of these clouds. Every weather unit should have a copy of one or both of these publications.

In this section we explore these four major topical areas:

- Low cloud type identification.
- Middle cloud type identification.

- High cloud type identification.
- Orographic cloud forms.

027. Low cloud type identification

Low clouds are cloud layers whose base ranges from near the surface to 6,500 feet above ground level (AGL). Low clouds consist of these four basic cloud types:

- Cumulus (CU).
- Cumulonimbus (CB).
- Stratocumulus (SC).
- Stratus (ST).

Cumulus

In the year 1803, an English pharmacist named Luke Howard divided all clouds into three basic groups—cumulus, stratus, and cirrus. Cumulus, translated from Latin, means “heap.” Heap aptly describes this cloud in most of its stages. In the earliest stage of development, CU usually forms in, and suggests, good weather. Point A on figure 3–1 illustrates that CU has a clearly defined outline during the building stage and appears very white in color. The base of the CU (shown by point B on figure 3–1) becomes darker as the cloud builds in size, but generally remains horizontal. After the building stage has gone on for a while, or ended, the edges of CU become ragged as they’re fragmented by the wind. This is shown by point C on figure 3–1.

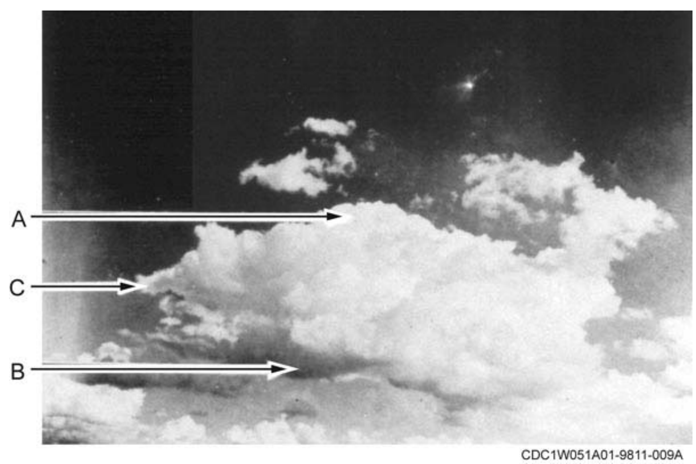

Figure 3–1. Cumulus.

Notice the bulging appearance in point A on figure 3–1. This is characteristic of building CU. Whatever its stage of development, CU always has the “cottony” appearance. Since these clouds form by convective action, the height of their base above earth’s surface relates directly to the amount of moisture at the surface—the higher the moisture content, the lower the cloud bases. Although there are many water droplets in CU, these droplets are very small in the cloud’s early stages.

As the cloud grows in size, large drops within the cloud increase in number. The large drops may be precipitated from the cloud or may continue to be carried within the cloud by updrafts.

Precipitation in the form of showers may occur with CU clouds of moderate development. Though this precipitation may be of moderate intensity, its duration is usually short lived. These clouds don’t produce the heavy rain and high winds as the larger well-developed CB clouds. Occasionally, the precipitation (rain showers) from CU clouds evaporates before it reaches the ground. This condition is called *virga* and is characterized by a dark area immediately below the nearly uniform base of the CU cloud.

This darkness, which is caused by precipitation, decreases in intensity as it descends beneath the cloud. It will do this until it disappears (complete evaporation). When virga consisting of snow or ice crystals occurs, the virga portion isn't as dark and appears whispier. This is caused by the greater influence wind has on falling snow and appears as a greater bending of the precipitation trails (virga). Either way, the precipitation doesn't reach the surface. Air Force Weather (AFW) classes two subtypes of CU for coding—L1 (cumulus) and L2 (cumulus).

L1 (cumulus)

CU clouds encoded as low cloud "1" have little vertical extent (cumulus humilus) or may appear flattened and ragged (cumulus fractus). Associated with good weather conditions (no precipitation), you should encode such a cloud as low cloud "1." When CU fractus clouds occur below CB or nimbostratus (NS) during precipitation, you code them as L7. Under these conditions, the CU clouds usually appear ragged and change shape rapidly. Thus, the difference in classification of L1 and L7 is precipitation. When the convective forces (which form CU) continue their action, L1 CU grows into L2 CU.

L2 (cumulus)

Low cloud "2" is a CU cloud of moderate or strong (towering) vertical development. Generally, other CU or SC clouds that have their bases at the same level accompany L2. When this cloud type develops a tower appearance, you should enter a remark in your observation stating the direction of the towering cumulus (TCU) from the station. The lack of massiveness distinguishes TCU from CB; that is, its strong vertical growth isn't matched by a horizontal spreading or bulging. Some important things to remember about this cloud type are:

- They don't have a cirriform top and are not capable of producing thunder.
- CU clouds of moderate or strong vertical extent may still produce precipitation in the form of showers.
- When a CU develops in height, strength, and massiveness, it enters the next basic cloud category—cumulonimbus.

Cumulonimbus

Energy forces within a CB are capable of producing the most intense storm known in weather—the tornado. However, as you observe CB clouds on the horizon, they appear strikingly beautiful. Their tall, rounded masses reach gracefully skyward, often penetrating above CI cloud formations. Overhead, they present a more menacing picture. When these clouds are overhead, other identifying guides such as thunder, lightning, and so forth are needed for CB identification, It isn't uncommon for these clouds to produce heavy rain, lightning, strong, gusty surface winds, hail, and, occasionally; tornadoes.

To classify CB clouds into the basic cloud forms, we distinguish them from CU clouds by the following characteristics:

- Massive appearance.
- Extensive vertical development.
- Fibrous or anvil top.
- Thunder and lightning.

Though the anvil feature of CB is an identifying feature, we sometimes observe only a fibrous appearance or a lack of sharp top outline. Formation of the anvil normally signals the transition to the dissipation stage. Figure 3–2 shows several interesting features. In point A, the anvil top is visible. Points B and C show the fibrous appearance of a CB top. At point B, the cloud is just beginning to lose its sharp outline. At point C, the fibrous appearance is evident. Point D points to a cell of convective activity that shows the typical sharp outlines of a building CU.

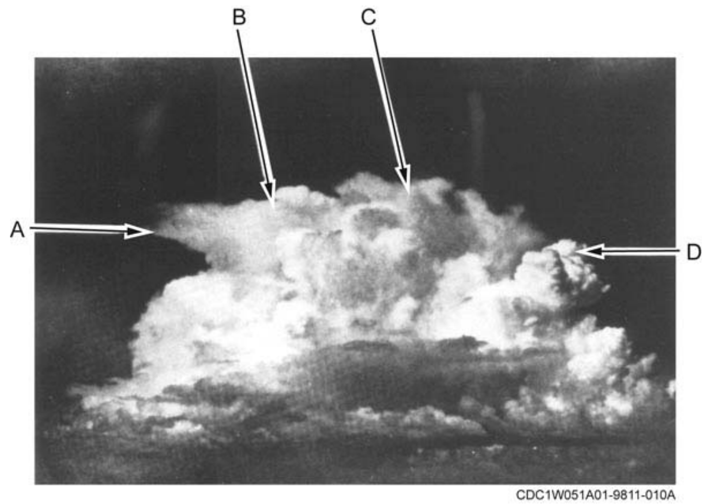

Figure 3–2. Cumulonimbus.

Often, you'll encounter a dissipating cell next to a building cell. Point A on figure 3–2 shows the anvil top frequently associated with the dissipating stage. During this stage, much of the cell's energy is directed downward. Thus, surface weather may be even more severe.

The question often asked is this: When does the CU (moderate development) become a CB? There are several points of CB identification. When viewed from a distance, the massiveness and the appearance of the cloud top we already mentioned offer positive means of CB identification. Overhead, other identifying guides are needed.

Thunder, lightning, or hail may be the sole indication of the CB presence. When you can't hear thunder overhead and are having trouble deciding between NS and CB, use the character of the precipitation (showery versus continuous) as a guide. A CB cloud that has begun dissipating is generally preceded by an outflow of cool air a few minutes before the storm cell reaches overhead. (This is normally when the strongest surface wind gusts occur.) Rapidly moving stratus fractus and cumulus fractus clouds usually accompany the dark lower portion of a CB. Usually, one of these signs can identify a CB from a CU.

A common occurrence with CB cloud varieties is mammatus development. This feature normally occurs at the base of the cloud in clearly defined bulges (pouches) but may appear at any level above the cloud base. This is illustrated on figure 3–3. In either case, these mammatus formations provide you with a good indication of the degree of instability present in the area. Though these cloud types may not produce tornadic activity, you can use them as indicators of potentially severe weather.

Figure 3–3. Mammatus.

Studies of tornado development reveal the base of the CB cloud is usually very dark and ragged before tornado activity occurs. The first sign of a funnel cloud often appears as a small appendage, which is often cone-shaped beneath the cloud. When you sight cone-shaped appendage with a CB, it will frequently appear and withdraw from several portions of the cloud. An appendage that continues to develop toward the ground is called a funnel cloud until it reaches the ground—it's then classified as a tornado.

The passage of a CB can cause a variety of changes in weather. Observing and disseminating these conditions presents a challenge. For coding purposes, two subtypes of CB exist—L3 (cumulonimbus) and L9 (cumulonimbus).

L3 (cumulonimbus)

Low cloud type "3" is CB in its earliest stage of development. A low "3" cloud differs from other CB clouds because the top lacks cirriform development (no anvil). CB clouds classified as L3 have tops that lack clear outlines and their shape is neither clearly fibrous (cirriform) nor in the form of an anvil. When you observe this type of cloud, add a remark to your observation to show the location (direction) of the CB cloud from the station and the direction toward which it's moving.

L9 (cumulonimbus)

Low cloud type "9" is distinguished from L3 by the presence of the cirriform anvil. If you find it difficult to decide whether the type is L3 or L9, keep in mind the occurrence of lightning, thunder, or hail is customarily associated with L9.

Stratocumulus

SC clouds form in several ways. They form when ST clouds near earth's surface lift, CU clouds flatten, or middle cloud layers lower. SC clouds are distinguished from CU by their flatter appearance. As SC clouds merge into one layer, they appear gray with dark areas. These dark areas are the thicker portions of the SC clouds. An SC cloud is sometimes mistaken for an altocumulus (AC) cloud. The best way to judge whether a cloud is SC or AC is by the size of the individual elements.

The *International Cloud Atlas* states that when the regularly arranged small elements of the cloud layer have an apparent width of more than 5º, the cloud is identified as SC. An easy method of determining this width is to hold three fingers at arm's length and see if the cloud element is larger than the three extended fingers. If it isn't, then perhaps the cloud is AC.

The rounded masses and rolls of L5 are unique features of the SC type cloud. The variety of SC shown as L4 frequently forms from the spreading out of CU in the later afternoon when the surface heating greatly diminishes.

Precipitation rarely occurs in association with SC clouds. When it does occur, it's usually weak and intermittent in character. Light snow showers are probably the most common forms of precipitation from SC. During cold weather, SC clouds frequently produce ice crystal virga. For coding purposes, three subtypes of SC exist:

- L4 and L5 (stratocumulus).
- L8 (cumulus and stratocumulus).

L4 and L5 (stratocumulus)

Encode low cloud as type "4" only when SC clouds form from the spreading out of CU or CB clouds. During this spreading process, CU clouds may still be present. When SC clouds form by other means, classify them as low cloud "5." L5 essentially includes all SC clouds not formed from the spreading out of CU clouds. If you can't determine that SC formed from CU, classify the cloud as L5.

L8 (cumulus and stratocumulus)

This state of sky is a combination of two other low cloud types—CU and SC. When CU and SC clouds have bases at different levels and the SC forms by other than the spreading out of CU, classify the cloud type as L8. Often, weather journeymen mistake a layer of SC as ST. Avoid that mistake by using the guidelines discussed in the next few paragraphs.

Stratus

You can discriminate between ST and a layer of SC by the uniform appearance of the ST cloud base. SC always has an unequal distribution of darkness. When they're dissipating, ST clouds may appear as large, irregular dark patches between lighter colored portions that are already thinning. The entire cloud takes on a mottled or blotched appearance.

An ST cloud usually forms very close to earth's surface and is called fog when it's in contact with earth. It may also form under other cloud layers such as AS and NS. ST is capable of producing only light continuous precipitation, such as drizzle, ice prisms, or snow grains.

ST clouds are frequently confused with NS and AS. To help clarify identification, study the comparison below.

Stratus	Nimbostratus
Produces only light precipitation, if any.	Always produces heavier precipitation
May reveal the Sun through its thinnest parts.	Never reveals the Sun.
Has a more uniform base than NS.	Has an uneven base.
Is generally gray.	Is usually darker in appearance than ST.

When the outline of the Sun is distinguishable through ST clouds, you can use it to distinguish between ST and AS. The Sun seen through ST has a sharp, well-defined outline. AS blurs the outline of the Sun as if you are viewing it through ground glass. When you evaluate ST cloud types, consider past observations of the clouds as a basis for proper cloud identification. Stratiform clouds don't develop as rapidly as cumuliform cloud types. They're usually associated with a stable condition in the atmosphere and, therefore, evolve slowly. For coding purposes, two subtypes of ST exist:

- L6 (stratus).
- L7 (stratus fractus or cumulus fractus).

L6 (stratus)

Low cloud "6" is an ST cloud in a more or less continuous sheet or layer, or in ragged shreds, or a combination of both, but it has no stratus fractus of bad weather. The primary difference between L6 and L7 is the presence of bad weather. This term refers to the conditions that exist a short time before, during, and after precipitation. Since all stratiform clouds appear grayish and continuous in form, be aware of the identifying features of each stratiform type.

L7 (stratus fractus or cumulus fractus)

Low cloud type "7" often occurs below layers of AS and NS. It's classified as L7 whenever the stratus fractus or cumulus fractus of bad weather are present. When these cloud types are present, but bad weather conditions don't exist, stratus fractus clouds are classed as L6 (ST) and cumulus fractus clouds as L1 (CU).

Ragged stratus fractus clouds never occur alone; instead, they're always associated with clouds of low and/or middle types. When you observe them below NS and similar precipitating clouds, they change shape rapidly and move fast. Stratus fractus and cumulus fractus are usually found beneath the base of precipitating CB clouds, but when this situation occurs, you encode only the CB cloud type (L3 or L9). The next cloud types we'll explore are the middle clouds.

028. Middle cloud type identification

Middle clouds are cloud layers whose bases range from 6,500 to 23,000 feet AGL. Middle clouds consist of these three basic cloud types:

- Altostratus.
- Nimbostratus.
- Altocumulus.

Altostratus

This middle height range cloud has features similar to the lower stratus. The primary difference between AS and ST is the composition of the clouds. An AS cloud consists primarily of ice crystals, snow crystals or flakes, and supercooled water droplets. The lower portion of low AS clouds may consist of ordinary water droplets and the upper portion contains a combination of ice crystals and supercooled water droplets. The composition explains the different features of each cloud.

AS clouds are generally uniform in appearance. They're grayish or bluish in color and appear fibrous. Four other basic characteristics are as follows:

- AS clouds are dense enough to prevent objects on the ground from casting shadows.
- The Sun appears as though you're viewing it through ground glass when an AS cloud is present.
- Halo phenomena never occurs with AS clouds.
- Precipitation is continuous.

Precipitation falling from an AS cloud frequently obscures the cloud base. When this occurs, clouds such as cumulus fractus and stratus fractus may form below the AS. Figure 3–4 illustrates this condition. During the hours of darkness, AS clouds are even more difficult to identify. At this time, watch for such things as a lowering of the ceiling and an increase in the intensity of precipitation. If this happens, you may have NS.

Figure 3–4. Fractus clouds.

AS clouds, like other middle clouds, are found at a height range from 6,500 to 23,000 feet in the temperate region. When they're at the higher levels of this middle cloud range, they're often erroneously identified as CS because of their lighter appearance. However, if they cast *no* shadows on the ground, they're AS. CS is never dense enough to prevent the casting of shadows when the Sun is high above the horizon. When the AS lowers, as during the approach of a warm front, it usually becomes thicker and completely obscures the Sun. For coding purposes, two subtypes of AS exist:

- M1 (altostratus).
- M2 (altostratus or nimbostratus).

M1 (altostratus)

Middle cloud type “1” is an AS cloud, the greater part of which is semitransparent. Usually, the Sun or Moon is dimly visible as though you are viewing it through ground glass. You can usually find this cloud type within the higher portion of the middle cloud range. This type of AS cloud usually forms from the gradual thickening and lowering of a CS layer. In a later discussion of the basic cloud form cirrostratus, you'll learn these clouds are never thick enough to prevent objects from casting shadows because of the Sun. Therefore, you can use this rule as a guide in determining whether you have AS clouds. More rarely, this type of AS cloud forms from the extensive spreading out of the middle or upper part of a CB cloud. When AS clouds continue to thicken, classify them as M2 clouds.

M2 (altostratus or nimbostratus)

An AS cloud classified as M2 is a darker gray or a darker, bluish gray than AS clouds encoded as M1. The greater part of this AS cloud (M2) is dense enough to hide the Sun or Moon. A NS cloud, which is also encoded as M2, is often caused by a further thickening of dense AS.

Nimbostratus

The word “nimbus” comes from the Latin term meaning violent rain or black rain cloud. NS clouds live up to this definition. An NS cloud produces continuous precipitation in the form of rain, snow, or ice pellets.

NS is a gray, often dark, cloud that appears diffuse when you observe it from the ground. The continuous precipitation that falls from this cloud causes this diffusion. NS is always thick enough to completely obscure the Sun and is almost exclusively found near frontal zones. It's common to find stratus fractus clouds below NS. These clouds are caused by the falling precipitation from the NS cloud but tend to dissipate completely when the precipitation becomes heavier.

Though NS is classified as a middle cloud, its base is most often found in the low cloud range. Examples of this are evident as warm fronts approach the station. Classify AS as NS when the cloud increases in density and heavier precipitation occurs. This cloud may continue to lower to within several hundred feet of the surface as the front approaches. Correctly identifying this cloud can alert you to the pattern of weather you can expect at the observation site.

NS clouds are distinguished from opaque AS clouds by their denser and darker appearance. The base of an NS cloud has a more diffuse and wet appearance than an AS cloud. However, classify both cloud forms as M2. NS clouds usually evolve from the thickening of AS clouds but may also evolve from CB clouds.

Altocumulus

An AC cloud is composed largely of water droplets but, at very low temperatures, it may have some ice crystal development. AC clouds often look very much like SC clouds. The primary differences between these two cloud types are the size of the elements and their height. One way to distinguish between AC clouds and other cloud forms is to figure out the size of the cloud elements. Extending three fingers at arm's length, the size of the elements should fall within the area covered by your fingers. If they don't cover at least one finger, they're probably cirrocumulus. This guide is useful only when the cloud elements in question are more than 30º above the horizon.

When an AC cloud doesn't have uniformly arranged elements, consider other identifying features of the clouds. AC clouds appear white, gray, or a combination of white and gray. They occur in any of the following seven forms:

- Rounded masses and rolls (such as SC).
- Banded.
- Semitransparent.
- Lenticular (unusual shaping by the wind).
- Castellated (tufts, turrets, etc.).

- Double layered.
- Dark and thick.

Of all basic cloud forms, AC has more varieties. AC clouds evolve from the lifting of lower clouds or, more rarely, from the thickening and lowering of cirrocumulus. As large CU clouds (TCU or CB) dissipate, the middle portion of the cloud frequently becomes AC. In this case, selection of the correct type of cloud has a definite meteorological significance to you, the weather craftsman.

The virga phenomenon is common with AC. When it occurs, the precipitation trails appear smaller than those associated with low clouds. A corona is often present with AC clouds when they're semitransparent. This phenomenon is especially useful in determining the type of cloud during hours of darkness. A corona appears as a small ring of light around the Moon and appears to blend with the Moon's light, whereas a halo presents a large distinct circle of light around the Moon.

Sometimes, a corona displays the rainbow colors faintly. The diffraction of light through water particles causes the corona. The diameter of the corona depends on the size of the water droplets in the cloud. Large water droplets produce a small corona and small water droplets produce a large corona. For coding purposes, seven subtypes of AC exist:

- M3 (altocumulus).
- M4 (altocumulus).
- M5 (altocumulus).
- M6 (altocumulus).
- M7 (altocumulus or altocumulus with altostratus).
- M8 (altocumulus).
- M9 (altocumulus).

M3 (altocumulus)

Middle cloud type "3" is an AC cloud, the greater part of which is semitransparent. The various elements of the cloud change slowly and are at the same level. This description of M3 clouds doesn't imply that some elements can't be opaque. Generally this cloud type has some degree of opaqueness, but it's predominantly semitransparent. The elements are somewhat small and undergo changes very slowly. This cloud type doesn't progressively invade the sky.

M4 (altocumulus)

Middle cloud type "4" is an AC cloud in patches; the greater part of it is semitransparent. The cloud elements occur at one or more levels and continually change in appearance. Often this cloud type (M4) appears as almond or fish shaped. These unusual lenticular-shaped cloud forms are mostly found near the mountainous regions but may occur at any location.

NOTE: An additional discussion of this lenticular cloud is presented in the lesson on orographically related clouds.

M5 (altocumulus)

Middle cloud type "5" AC, is arranged in semitransparent bands or in one or more fairly continuous layers that progressively invade the sky. In either case, the main characteristic of M5 clouds is that they generally thicken as a whole. Once the forward edge of the cloud reaches the part of the horizon opposite to that part where the clouds first appeared, the cloud is no longer classified as M5. This is also the case when the forward edge ceases advancing.

M6 (altocumulus)

AC clouds classified as M6 form from the spreading out of CU or CB clouds. As large CU clouds or CB clouds dissipate, their remains often consist of large, dark elements. They usually continue to dissipate and thin out to form into separate elements. The best guide to determine the presence of M6 is to view the actual transformation of CU to AC.

M7 (altocumulus or altocumulus with altostratus)

Middle cloud type "7" has AC clouds in two or more layers. They're usually opaque in places and don't progressively invade the sky. Middle cloud type "7" also has AC clouds with AS or NS clouds. This cloud type is a combination of other middle cloud types. For example, if AS (M2) and AC (M3) are present and together, encode the cloud type as "7." Generally the AC elements of this cloud type aren't continually changing.

M8 (altocumulus)

Middle cloud type "8" is an AC cloud with sproutings in the form of small towers or battlements. Figure 3–5 (M8 castellanus) illustrates the sproutings. Another form of middle cloud 8 is similar to very small CU clouds or tufts in the middle cloud range; it often appears ragged. When this cloud has the sproutings in the form of turrets, the cloud is called altocumulus castellanus. A remark with the observation emphasizes this significant cloud.

CDC1W051A01-9811-013A

Figure 3–5. Altocumulus.

M9 (altocumulus)

Middle cloud type "9" is an AC cloud form of a chaotic sky and occurs at several levels. As seen from the ground, this cloud type appears heavy and stagnant. Meteorologically speaking, AC clouds of a chaotic sky are found near low-pressure areas that contain some storm activity. AC clouds are frequently forced to higher levels in the atmosphere. When this occurs, they're called cirriform clouds.

029. High cloud type identification

High clouds are clouds whose bases range from 16,500 to 45,000 feet AGL. High clouds consist of these three basic cloud types:

- Cirrus.
- Cirrostratus.
- Cirrocumulus.

Cirrus

CI clouds generally form between 16,500 and 45,000 feet in the temperate zone. They appear as very white clouds, usually in patches or filaments. The form of CI cloud most readily identified is the hook-shaped CI. Figure 3–6 illustrates CI clouds. As you can see, this type of cloud has very fine strands that the wind shapes into the form of a hook.

CDC1W051A01-9811-014A

Figure 3–6. Cirrus.

CI clouds of a denser variety (H3) frequently evolve from the dissipation of other basic cloud forms such as CB. The cirriform remains of a CB cloud may spread out to a great extent and completely lose their former identity (anvil shape). CI clouds also form from middle cloud layers that are forced aloft. CI and CS clouds are often combined in one layer. When an extensive CS layer approaches the station from the distant horizon, the leading edge is usually CI clouds. As the layer continues to approach, the cloud layer becomes more uniform and usually thickens. This situation is common in advance of a warm front.

The halo phenomena shown in figure 3–7 can occur with CI clouds, but this is somewhat rare. When a halo is present with CI, it’s usually only a partial halo because of the characteristics of CI (strands, filaments, etc.). When the halo is a complete circle, suspect the presence of cirrostratus. For coding purposes, four subtypes of CI exist:

- H1 (cirrus).
- H2 (cirrus).
- H3 (cirrus).
- H4 (cirrus).

CDC1W051A01-9811-015

Figure 3–7. Halo.

H1 (cirrus)

High cloud "1" is a CI cloud in the form of filaments, strands, or hooks that don't progressively invade the sky. This cloud type is often present with other CI clouds. In this case, classify the cloud type as H1 only when the total amount of hooks, filaments, or strands is greater than the combined total of the other CI clouds present. Whatever the situation, remember that H1 doesn't progressively invade the sky.

H2 (cirrus)

High cloud "2" is a dense CI cloud that's in patches or entangled sheaves that usually don't increase in size. These patches sometimes seem to be the remains of the upper part of a CB. An H2 cloud can also be CI with sproutings as small turrets or battlements or CI that has the appearance of cumuliform tufts. This dense CI cloud doesn't originate from CB clouds, although the patches are sometimes opaque and have borders of entangled filaments. This can give the erroneous impression that the cloud patches are the remains of cumuliform clouds. When an H2 cloud is present with other CI clouds, the H2 characteristics predominate. This will cause you to encode the clouds as such. H2 and H3 clouds are often mistaken for each other. When you're certain the cloud evolved from a CB cloud, classify the cloud as H3.

H3 (cirrus)

High cloud type "3" is a dense cloud that's often in the form of an anvil and is the remains of the upper parts of a CB cloud. The best guide to classify this cloud type is to observe the upper part of a CB cloud as it transforms into dense CI. If you have sufficient evidence that the dense CI cloud evolved from cumuliform clouds, you may classify dense CI clouds as H3, even though you don't actually see the transformation. This evidence may come from pilot sightings of CB clouds near your area or the unmistakable features associated with the dissipation of cumuliform clouds (M6 for example).

H4 (cirrus)

High cloud type "4" is a CI cloud in the form of hooks and/or filaments that progressively invades the sky and becomes more dense. This cloud type is very similar to H1 except an H4 cloud progressively invades the sky and becomes denser. These clouds appear to fuse near the horizon where they first appear, but no CS clouds are present. When CS conditions are present, examine the clouds closely to decide whether to classify the type as H5.

Cirrostratus

A CS cloud is a whitish veil that's very similar in appearance to CI clouds. The primary difference is the great horizontal extent of CS and its more veil-like appearance. CS clouds usually produce a halo when the cloud composition is thin enough. CS often appears as AS on the distant horizon. In this case, consider the speed of movement of the cloud (a CS cloud appears to move more slowly) and the slower changes in form and appearance that are characteristic of CS. CS clouds on the horizon are sometimes confused with haze. You can distinguish the haze by its dirty yellow-to-brown color.

A CS cloud is never thick enough to prevent objects on the ground from casting shadows when the Sun is higher than 30º above the horizon. For example, a CS layer may be so thin that only the presence of a halo reveals its presence. For coding purposes, four subtypes of CS exist:

- H5 (cirrus and cirrostratus or cirrostratus alone).
- H6 (cirrus and cirrostratus or cirrostratus alone).
- H7 (cirrostratus).
- H8 (cirrostratus).

H5 (cirrus and cirrostratus or cirrostratus alone)

High cloud type “5” is CI and CS clouds or CS clouds only. (The CI clouds are often in bands that converge toward one point or two opposite points of the horizon.) In either case, they progressively invade the sky and generally grow denser, but the continuous veil doesn’t reach 45º above the horizon. Usually, the leading edge of this cloud type is in the form of CI filaments or hooks and, occasionally, resembles the skeleton of a fish. When this cloud type progresses to 45º above the horizon, classify it as H6.

H6 (cirrus and cirrostratus or cirrostratus alone)

High cloud type “6” has the same appearance and features of H5 but extends to more than 45º above the horizon, without the sky being actually covered. Similar to H5, it progressively invades the sky and becomes denser. When the cloud layer covers the entire sky, classify it as H7.

H7 (cirrostratus)

High cloud type “7” is a veil of CS clouds that covers the celestial dome. This cloud is uniform in structure and shows few distinctive details. On occasion, the continuous veil of H7 is so thin (transparent) that the only indication of its presence is a halo phenomenon. When lower clouds obscure parts of an overcast CS layer, you may still classify it as H7 if you have evidence that the layer covers the sky. If the CS layer doesn’t cover the sky, classify the cloud type as H8.

H8 (cirrostratus)

High cloud type “8” is CS that’s no longer progressively invading the sky and doesn’t completely cover the celestial dome. When H8 is present with other cirriform cloud types, it must be predominant to be classified as H8. Though the definition of this cloud type specifically states that the CS clouds aren’t progressively invading the sky, this refers to the continuous veil form of the CS formation. When CS is in patches (not CI), H5, H6, and H7 aren’t appropriate classifications. Even if patches of CS clouds are increasing, classify them as H8, even though CI and cirrocumulus (CC) clouds may also be present, but not predominant.

Cirrocumulus

CC clouds (H9) are very much like the regularly arranged elements of high AC clouds. The difference is their size and composition. To be CC clouds, the element must have an apparent width of less than 1º. Measure this by extending your little finger at arm’s length. If the element you’re evaluating isn’t larger than your finger, the cloud type is probably CC. Again, this guide is only reliable when the cloud element is higher than 30º above the horizon.

CC clouds consist primarily of ice crystals, but they can also consist of minute supercooled water droplets that are usually replaced rapidly by ice crystals. CC clouds are observed with a slight corona phenomenon that adds to the beauty of the cloud. When this cloud is present, the sky is often called a *mackerel sky* because of the cloud layer’s resemblance to the scales of a fish. Some other terms used to identify this cloud are pebbles on a beach, honeycomb, and netlike.

Some forms of CC clouds are similar to altocumulus castellanus clouds. They appear as small tufts or turrets; yet, they must be less than 1º in width to be classified as CC. Some elements appear so small they’re difficult to discern with the naked eye.

High cloud “9” is CC clouds by themselves or accompanied by CI and/or CS clouds, but the CC clouds must be predominant. Be sure to remember that the elements of CC must have an apparent width of less than 1º.

030. Types of orographic cloud forms

Certain types of clouds form because of air moving over rough terrain. These clouds suggest the presence of a mountain-wave condition in the atmosphere. Because severe turbulence is associated with mountain-wave conditions, these clouds are significant to flight operations. Therefore it's very important that you properly identify these clouds.

The most common orographic clouds belong to the same class as AC, SC, and CU clouds. Listed below are the orographically produced clouds associated with mountain waves:

- Lenticular—SC, AC or CC.
- Rotor (roll)—CU.
- Cap (foehn wall, collar)—SC.

Lenticular

The lenticular cloud is a SC (L5), AC (M4) or CC (H9) cloud that's almond or fish shaped. You'll observe these clouds in patches at one or more levels. The elements are continually changing in appearance but generally remain stationary in spite of the high wind speed. They constantly form on their windward side and dissipate on their downwind side of the leeward side of the mountain. Since the cloud patches are of limited horizontal extent and their elements are continually changing, these clouds are usually semitransparent (opaque). The patches, as a whole, may have the form of large lenses and aren't progressively invading the sky. CC and especially SC lenticulars are rare.

Rotor

Rotor clouds are cumuliform in appearance and are found on the leeward side of a mountain range. Rotor clouds, similar to lenticular clouds, are stationary and are constantly forming on their windward side and dissipating on the leeward side. Because of their vertical development and cumuliform appearance, we usually encode them as low cloud type "2."

Cap

The cap cloud is SC in appearance and is usually classified as low cloud type "5." This cloud hugs and at times obscures the top of the mountain. Most of the cloud is on the windward side with part of the cloud flowing down the lee-side of the mountain, thus producing the appearance of a waterfall.

Self-Test Questions

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

027. Low cloud type identification

1. Match the low cloud classification in column B with the specific description and cloud characteristics in column A. Column B items may be used only once.

Column A	*Column B*
____ (1) A cloud of moderate or strong vertical development and doesn't have a cirriform top.	a. L1.
____ (2) Consists of clouds formed by means other than the spreading out of CU and has clouds with bases at different levels.	b. L2.
____ (3) In its earliest stage of development, this cloud type usually forms in, and suggests, good weather.	c. L3.
____ (4) Has a summit that lacks clear outline and isn't clearly fibrous nor in the shape of an anvil.	d. L4.
____ (5) Forms from the spreading out of CU or CB clouds.	e. L5.
____ (6) Forms by means other than the spreading out of CU .	f. L6.
____ (7) Is in a more or less continuous sheet or layer, or in ragged shreds, or is a combination of both and bad weather isn't present.	g. L7.
____ (8) Has a cirriform anvil.	h. L8.
____ (9) Often occurs below layers of AS and NS during bad weather.	i. L9.

028. Middle cloud type identification

1. Match the middle cloud classification in column B with the specific description and cloud characteristics in column A. Items in column B may be used only once.

Column A	*Column B*
____ (1) The greater part of this cloud type is semitransparent and the Sun or Moon is dimly visible as though you're viewing it through ground glass.	a. M1.
____ (2) Forms from the spreading out of CU or CB clouds.	b. M2.
____ (3) Is sufficiently dense to hide the Sun or Moon.	c. M3.
____ (4) This cloud type, the greater part of which is semitransparent, doesn't progressively invade the sky.	d. M4.
____ (5) Has two or more layers, is opaque in places, and doesn't progressively invade the sky.	e. M5.
____ (6) Is called a chaotic sky and occurs at several levels.	f. M6.
____ (7) Forms in patches and continually changes in appearance.	g. M7.
____ (8) Is arranged in bands or in one or more fairly continuous layers that progressively invade the sky.	h. M8.
____ (9) Forms sproutings in the form of small towers or battlements.	i. M9.

029. High cloud type identification

1. Match the high cloud classification in column B with the specific description and cloud characteristics in column A. Items in column B may be used only once.

Column A	*Column B*
____ (1) Often called a "mackerel sky."	a. H1.
____ (2) A dense cloud and is often in the shape of an anvil.	b. H2.
____ (3) Uniform in structure, covers the celestial dome, and may have a halo as the only indication of its presence.	c. H3.
____ (4) Is in filaments, strands, or hooks that don't progressively invade the sky.	d. H4.
____ (5) Progressively invades the sky and generally grows more dense, but the continuous veil doesn't reach 45º.	e. H5.
____ (6) Can have sproutings in the form of small turrets or battlements.	f. H6.
____ (7) Is in the form of hooks and/or filaments that progressively invade the sky and become more dense.	g. H7.
____ (8) Progressively invades the sky and generally grows more dense. The continuous veil extends to more than 45º.	h. H8.
____ (9) This cloud type isn't progressively invading the sky and doesn't completely cover the celestial dome.	i. H9.

030. Types of orographic cloud forms

1. Why is it so important that weather personnel correctly identify orographically produced clouds in the observation?

2. Supply the name and subtype number for each of the following orographic clouds:

   (a) This middle-level orographic cloud type is observed in patches at one or more levels; the elements are continually changing in appearance but generally remain stationary in spite of high wind speeds.

   (b) This orographic cloud type hugs the top of a mountain and gives the appearance of a waterfall.

   (c) This orographic cloud type is found on the leeward side of a mountain range and has vertical development.

3–2. Sky Condition

Determining the sky condition is largely subjective and requires, above all, practical experience. There's one important reason for a careful evaluation of the sky—almost all changes in surface weather are preceded or accompanied by clouds. For example, frontal passages give advance warning of their presence by a series of changes in clouds and sky condition. The weather craftsman interprets the significance of these changes from your observation. Your training in surface observations will prepare you to recognize changes in sky condition and apply them to the forecast process.

Looking at a series of observations, you can see sky transitions by the changes in the observed layer. A change in the amount of a layer from $^6/_8$ to $^5/_8$ may appear unimportant from one observation to the next. However, when you regard this minor change within a trend and in relation to all the other sky data, an approaching weather situation may be foretold. In this section we look at two subject areas:

- How to determine sky cover.
- How to determine ceiling types.

031. How to determine sky cover

Understanding sky condition begins with knowledge of sky cover. When determining sky cover, consider each layer starting with the lowest to the highest, in ascending order. At each layer, the amount of coverage is equal to the amount of sky cover at-and-below that layer (summation principle). After the amount of coverage is assessed, then the height of the layer is determined.

In this lesson we'll explore these three major areas:

- Characteristics of cloud layers.
- Amount of sky covered by each layer.
- Heights of layers.

Characteristics of cloud layers

As it applies to clouds, a *layer* is defined as "clouds or obscuring phenomena which have bases at the same approximate level." A layer may appear as continuous cover, such as stratus, or it may appear as detached elements, such as fair-weather CU. Also, both continuous and detached elements may combine to form a layer. The essential requirement is that the bases be at the same approximate level. The upper portions of a CB cloud are often spread horizontally by wind and form dense CI clouds. Regard these horizontal extensions of the CB clouds as separate layers only if the bases of these extensions appear horizontal and at a different level from the parent cloud. Otherwise, regard the entire cloud system as a single layer at a height corresponding to that of the base of the parent CB cloud.

A layer can be a combination of cloud types or obscuring phenomena at the same level. You should consider any combination of clouds or other phenomena aloft having a base at the same height as a single layer. However, obscuring phenomena, such as haze, are often present in the atmosphere but have no apparent base. In this situation, don't consider them a layer. After you've divided the state of the sky into layers of clouds, obscuring phenomena, or both, determine the amount of each layer.

Amount of sky covered by each layer

Although you observe the amount of sky covered by each layer in terms of eighths of sky covered, you should use contractions to describe the sky cover. Use these sky cover contractions for each layer of clouds or obscuring phenomena. Each contraction represents the portion of the sky covered at that layer and below. Figure 3–8 illustrates this "at and below" principle of assigning sky cover contractions. The difference between layer and sky cover is also shown. This table provides the reportable contraction for the summation amount of cloud cover.

Reportable Contraction	Meaning	Summation Amount of Layer
VV	Vertical Visibility	8/8
CLR	Clear	0
FEW*	Few	1/8–2/8
SCT	Scattered	3/8–4/8
BKN	Broken	5/8–7/8
OVC	Overcast	8/8
NOTE: * Any layer amount less than 1/8 is reported as FEW.		

In figure 3–8, classify the first layer (2/8 stratus) as FEW, per the table above. This 2/8 amount also represents the total sky cover at this level and below.

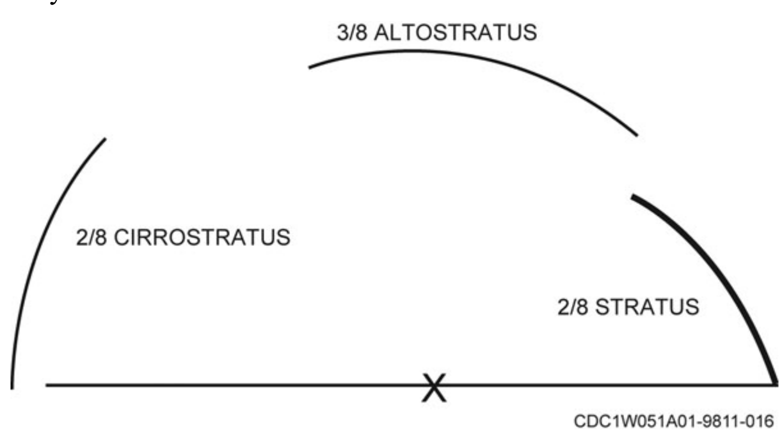

Figure 3-8. Exaggerated layer and sky cover.

The next layer tells a different story. Although the AS covers only 3/8 sky as a layer, the total sky cover-up to this point is 5/8 because of the combined amounts of the first two layers. Because of the idea described as “at and below” sky cover, there’s 5/8 of the sky covered at and below this level. Thus, the classification (in the table) for the AS layer is broken (BKN). Also, you’d classify the highest layer 2/8 (cirrostratus) as BKN, because the combined total equals 7/8 sky cover.

You can understand how meaningless it would be to enter three separate scattered (SCT 3/8 to 4/8 sky coverage) contractions to report these individual layers. To a pilot flying above the highest layer and looking for ground navigational aids, your report of SCT sky cover could hide 7/8 of the ground from the pilot’s view. By reporting a BKN sky cover, you’ve more accurately described the sky condition to the pilot.

Figure 3–9 shows a typical sky condition. What sky cover symbols/contractions should you enter in your observation for this example? You’re correct if you accomplished the following:

- Selected BKN for the first layer (fog).
- Selected BKN for the second layer (cumulus fractus).
- Selected overcast (OVC) for the highest layer.
- Added a remark about the surface-based obscuring phenomena.

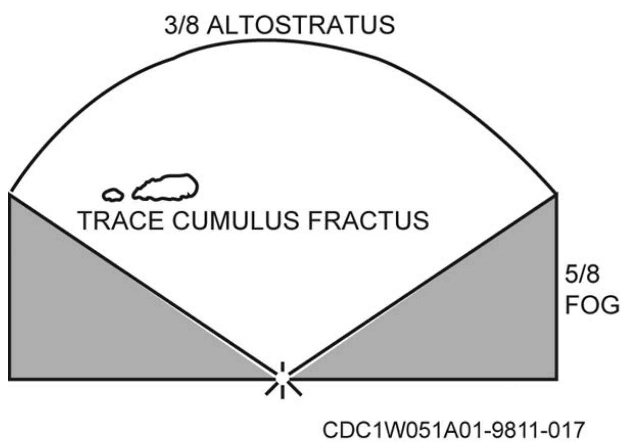

Figure 3–9. Surface-based sky cover.

The earlier table illustrates two principles. First, the 5/8 fog. Although it is surface based, the fog hides part of the sky just as if it were a cloud aloft. Second, treat the trace of CU fractus as a layer. Though this layer covers less than 1/8 sky, it’s a layer by definition and also meets the criteria for BKN sky cover. This is true because the total at and below that level (including the 5/8 fog) hides enough sky to require the BKN contraction.

When sky cover layers are advancing or receding on the horizon, use the left-hand column of the following table as a guide to determine the number of eighths of the sky covered by a layer. When a layer of sky cover surrounds the station, use the right-hand column of the table as a guide to determine the number of eighths of sky coverage. The table takes much of the guesswork out of estimating sky coverage at difficult angles of observation.

Angle of Advancing or Receding Layer Edge	Eighths of Sky Cover	Angular Elevation of Layer Surrounding Station
Trace to 50°	1	> 0 to 10°
51 to 68°	2	11 to 17°
69 to 82°	3	18 to 24°

Angle of Advancing or Receding Layer Edge	Eighths of Sky Cover	Angular Elevation of Layer Surrounding Station
83 to 98°	4	25 to 32°
99 to 112°	5	33 to 41°
113 to 129°	6	42 to 53°
130 to less than ≤ 179°	7	54 to 89°
180°	8	90°

During an observation of sky cover you should be alert for layers that occur directly beneath another layer. Here, you can't add the amounts of both layers to arrive at total sky cover because they hide the same section of the sky (for example, when $^2/_8$ of CU is below $^3/_8$ of altocumulus). Together these two layers hide only $^3/_8$ of the sky. These few examples help to illustrate the layer versus sky cover principle.

Height of layers

Heights of layers are reported according to established reportable values. The table below shows the reportable values that you can report. For example, during the evaluation of sky cover, suppose the following opaque layers were detected:

- Surface-based fog.
- SC at 4,780 ft.
- A trace of AC at 9,230 ft.
- AS at 16,500 ft.

Range of Height Values (feet)	Reportable Increment (feet)
≤5,000	To nearest 100
> 5,000 but ≤10,000	To nearest 500
> 10,000	To nearest 1,000

If you use the table correctly, the height entries for each layer should be: FEW000 SCT048 SCT090 BKN160. Notice that the last layer (altostratus) is exactly halfway between two reportable values. In this case, select the lower height.

For all sky coverage—whether FEW, SCT, BKN, or OVC, clouds or obscuring phenomena—use the height obtained from the most reliable method. Several methods are available for obtaining heights. You must consider not only the reliability of the height data but also the distance from the observation point, the height of the layer, and the time of observation. However, the same rules for obtaining heights apply for all layers, despite amount.

032. How to determine ceiling types

A ceiling is the lowest broken or overcast layer aloft or the vertical visibility in a surface-based layer of obscuring phenomena. In many cases, ceiling layers are the controlling factor for aircraft departures and landings. Low ceilings demand the most accurate measurement possible. Sometimes, a difference of 100 feet in the ceiling layer determines if pilots can safely land or if they must seek an alternate airfield. Therefore, you must be able to correctly judge the height of that sky cover. In this lesson, we'll cover the preferred and alternative methods of determining ceiling types.

Preferred method

Layer heights are determined in feet above the surface to the nearest reportable value. The preferred methods for evaluation of layer height or vertical visibility into a surface-based indefinite ceiling are to use one of the following methods:

- A ceilometer operating within its operational range.
- Known heights of unobscured portions of abrupt, isolated objects (buildings, towers, etc.) within $1^1/_2$ nautical miles of the runway.

Alternative methods

There are other methods that you can use to obtain the height of any layer. We'll discuss these methods next.

Aircraft

You can use layer heights reported by a pilot and converted from height above mean sea level (MSL) to height AGL when, in your judgment, the ceiling heights are representative of conditions over the airfield.

Balloon

You can determine a layer height by using balloons whenever necessary. For example, if the layer is at or below the minimum height for visual flight rules (VFR) operations or the ceiling height is 2,000 feet or less and the presence of a stratus-type cloud layer makes estimation difficult, you may use a balloon to determine the layer height.

A balloon height is based on the known ascension rate of a pilot or ceiling balloon. Ascension rates are fixed by the amount of lift given to the balloons. Proper balloon inflation (neither over nor underinflation) controls the lift. When you use a balloon to determine ceiling heights, use the following procedures:

1. Choose the appropriate color of balloon—red for thin clouds and blue or black balloons under all other conditions.
2. Watch the balloons continuously, determining with a stop watch (or any watch having a second hand) the length of time that elapses between the release of the balloon and its entry into the base of the cloud layer. Consider the point of entry as midway between the time the balloon begins to fade and the time of complete disappearance.
3. Then determine the height above the surface from the prepared table in AFMAN 15–111, *Surface Weather Observations*.

The accuracy of the height obtained by a balloon decreases due to the conditions such as:

- The balloon is released while precipitation is occurring and when a light is attached (at night).
- Ascension of the balloon is slowed by strong winds combined with poor horizontal visibilitys.
- Entry into an unrepresentative portion of the cloud base or through a break in the layer distorts the accuracy of the actual height attained.

Convective cloud heights

This method isn't suitable for stations in mountainous or hilly terrain. Use it only when the clouds present form by active surface heating and the height of cloud bases is at 5,000 feet or less. Also, use the figure with caution when the surface temperature is below freezing. Recent dew-point and free-air temperature readings must be available. This table may be used to find the difference between the free-air temperature and the dew-point temperatures. Reportable height values between those in the table may be obtained by means of interpolation using the difference to the nearest tenth of a degree.

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

Known heights of objects

You can use known heights of unobscured portions of natural landmarks or objects more than 1 1/2 nautical miles from any runway of an airfield to determine a layer height. Most weather stations have visibility charts that provide the heights of any hills, mountains, TV towers, and so forth, on or around your base. If, for example, there's a hill 3 miles from your base with a known height of 600 feet and the cloud base you're trying to evaluate is touching the top of the hill, you can estimate the layer height as 600 feet.

Observational experience

You can estimate a cloud height by observational experience provided the sky isn't completely hidden by surface-based obscuring phenomena and if other guides are lacking or, in your judgment, are considered unreliable. You can also consider the persistence of heights previously measured. Whenever possible, check such estimations against a reliable method of measurement. This comparison tells whether you usually estimate high or low under different sky cover conditions.

Ceilometers

A ceilometer is the most accurate method for measuring cloud height. Ceilometers are used to measure cloud bases. An automated sky condition is derived instrumentally by detecting the frequency and height of clouds passing over the ceilometer's sensor over a period of 30 minutes. An algorithm processes the data from the sensor into data on layers, amounts, and heights of clouds. The sensor derived sky condition is considered functionally equivalent to a manually derived sky condition. Automated Meteorlogical Observing Systems (AMOS) reports sky condition from 100 feet up to a maximum of 25,000 feet. AMOSs do not report sky condition above the range of the sensor.

Cloud elements

Use the apparent size of cloud elements, rolls, or features visible in the cloud layer to estimate layer heights. Large rolls usually indicate the cloud layer is somewhat low while small rolls or elements usually indicate the layer is somewhat high. The three-finger method of distinguishing SC from AC is an example of using cloud element size to determine cloud height.

City lights

You can estimate cloud height by using the reflection of city lights or other lights at night. During darkness, lights may reflect off the base of a layer. Use this observation to help verify the etage and height of clouds. Reflections are usually observed for low clouds, but over larger, brighter cities, middle etage clouds sometimes reflect light. Your unit may have rules on cloud light reflections already documented.

Self-Test Questions

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

031. How to determine sky cover

1. What's a layer?

2. Using the Reportable Contraction, Meaning, and Summation Amount of Layer table determine the amount of sky coverage of a layer of CS that's progressively invading the sky and whose leading edge is 89º.

3. If a layer was exactly between two reportable height values, which value would you report?

032. How to determine ceiling types

1. What's a ceiling?

2. What are the two preferred methods used to obtain a layer height?

3. List at least five alternative methods used to obtain a layer height.

3–3. Visibility and Runway Visual Range

Visibility, as well as ceiling height, aids in decisions involving air traffic control. For this reason, your observations of visibility must be timely, accurate, and representative. In this section, we explore the characteristics of visibility and runway visual range. Over the past decade, determining visibility has evolved from a manual estimation to a more accurate automated value determines by an AMOS. However, it's important to understand the manual process of determining visibility as well as the algorithms used by the AMOS to determine visibility.

033. Characteristics of visibility

In this lesson we'll cover two major subject areas—visibility markers and visibility determinations.

Visibility markers

You should use suitable objects in determining visibility. In order for your visibility observations to be representative, you should select visibility markers that meet certain criteria. Here are three marker types:

- Daytime markers.
- Nighttime markers.
- Visibility charts and aids.

Daytime markers

The most suitable and usable daytime markers are prominent dark-colored objects that can be observed in silhouette against a light-colored background, preferably the horizon sky. If you use an object against a terrestrial background, it should stand well in front of the background (a distance at least half that of the object from the point of observation).

Nighttime markers

The most desirable night-visibility markers are unfocused lights of moderate intensity. You may use the red or green lights of airway beacons and TV or radio tower obstruction lights. Because of their intensity, don't use focused lights of airway beacons; however, their brilliance may serve as an aid in estimating whether the visibility is greater or less than the distance to the light source.

Visibility charts and aids

A weather flight should have prepared, posted, and maintained current charts, lists, or other positive means of identifying objects selected as visibility markers. These should depict or list objects and include a description, distance, and bearing from the point of observation. Also, they should include a means of identifying objects that have been selected as night-visibility markers. The height of an object above the ground should be indicated (if known), especially for those objects which may be useful in determining sky cover height data.

Visibility determinations

Once you've determined the markers you'll use, you must be able to make decisions on the different visibilitys. These include the following:

- Prevailing visibility.
- Variable visibility.
- Sector visibility.
- Differing location visibility.

Prevailing visibility

Prevailing visibility is the greatest distance at which you can see and identify known objects throughout half or more of the horizon circle surrounding the station. The visibility doesn't have to be continuous throughout 180 consecutive degrees. To aid in determining the prevailing visibility, observing stations maintain a visibility chart or a list that identifies objects suitable for visual sightings.

Unfortunately, objects that meet the requirements to be good visibility markers aren't always present in every direction. When this happens, the station should use all available objects. However, identify markers of questionable value by indicating under what conditions they're most and/or least reliable.

If your station is such that your view of portions of the horizon are obstructed by trees, buildings, and so forth, use control tower values of prevailing visibility as a guide in determining your prevailing visibility. For example, the presence of a surface-based obstruction to vision that's uniformly distributed to heights above the level of the control tower is sufficient reason for evaluating your prevailing visibility as that of the control tower. If your station falls in this category, reevaluate your prevailing visibility, when practicable, upon notification of a differing control tower value or of a reportable change at control tower level.

Report prevailing visibility in statute miles in US locations (including Hawaii, Alaska, and Guam) and in meters overseas. If the visibility is halfway between two reportable values, enter the lower value. When you estimate the prevailing visibility to be more than the distance to the farthest visibility marker, estimate that visibility to the nearest reportable value. The prevailing visibility entry represents the prevailing ground-level visibility taken at your height from as many established observation points as necessary to view the entire horizon.

Use of an AMOS such as the FMQ–19 can also be used to determine visibility. AMOS's use algorithms to convert a sensor value and calculate an average visibility. Most weather stations are automated so be aware the prevailing visibility is only being detected from one point and not the 360 degree horizon. For this reason AMOS's cannot detect sector visibility.

Variable visibility

When you take a visual sighting on markers less than 3 miles away and the markers seem to appear and disappear alternately, thus indicating increasing and decreasing visibility, don't panic; instead, decide whether the visibility varies by one or more reportable values. At the same time, note all your observed values, average them, and if your average is less than a reportable value of 3 miles and varies by at least $^1/_2$ mile, report the visibility as variable. This simply means that you used the average value as the prevailing visibility. This average value, of course, represents the prevailing visibility which is common throughout half or more of the horizon circle. AMOS's are able to determine variable visibility through the use of an algorithm.

Sector visibility

Sector visibility is actually a part of determining prevailing visibility. Sector visibility points out a part of the horizon in which the visibility is uniform. When observing visibility, divide the horizon circle into a few or as many parts as needed to separate the different sector visibilitys. Figure 3–10 illustrates a horizon divided into four unequal sectors. Each sector presents uniform visibility within itself. For example, from north through east you can see 7 miles, east through south, 8 miles, and so forth. These two sectors together give the prevailing visibility of 7 miles. The number of sectors you need depends upon the uniformity of the visibility around the entire horizon.

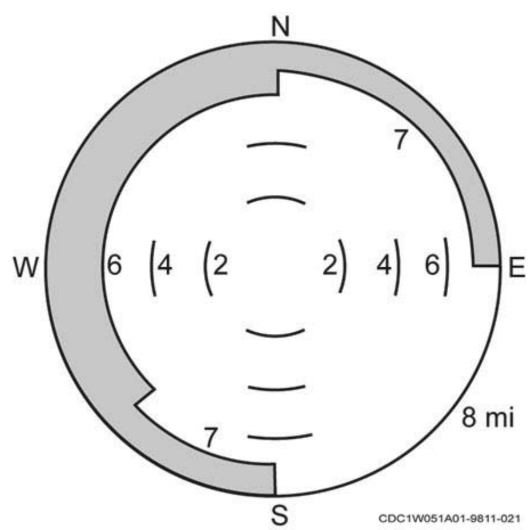

Figure 3–10. Sector visibility.

Sector visibility sometimes requires an explanation or remark. Refer to figure 3–11 and determine the prevailing visibility. The greatest distance seen throughout the western half of the circle is 4 miles, so this is the prevailing visibility. Imagine the pilots' surprise if they approach the field from the east. The entire sector from northeast to southeast has a visibility of $^1/_2$ mile.

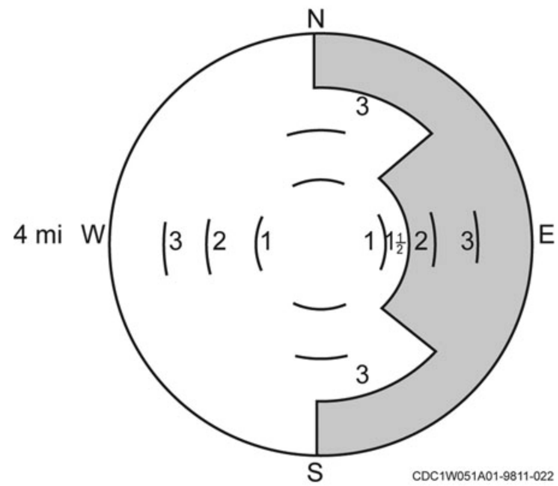

Figure 3–11. Sector visibility.

As defined in the AFMAN 15–111, sector visibility may be reported in remarks of weather observations when it differs from the prevailing visibility by one or more reportable values and either the prevailing or the sector visibility is less than 3 miles.

There's another point to consider concerning the significance of sector visibility. Suppose the sector visibility differs from prevailing visibility but is more than 3 miles (fig. 3–12). Both criteria for reporting sector visibility aren't met, but the difference is operationally significant. Operational significance is a legitimate reason for entering a remark on a sector visibility. Otherwise, with a prevailing visibility of 7 or more miles, no hint of an obstruction to vision is contained in the observation. Suppose there's a north-south active runway in figure 3–12. The 4 miles visibility to the north will go unnoticed unless you make an operationally significant remark to alert the pilot of the visibility in the sector. Therefore, when a sector visibility of 3 miles or more differs from the prevailing visibility and you consider the difference operationally significant, enter a sector visibility remark.

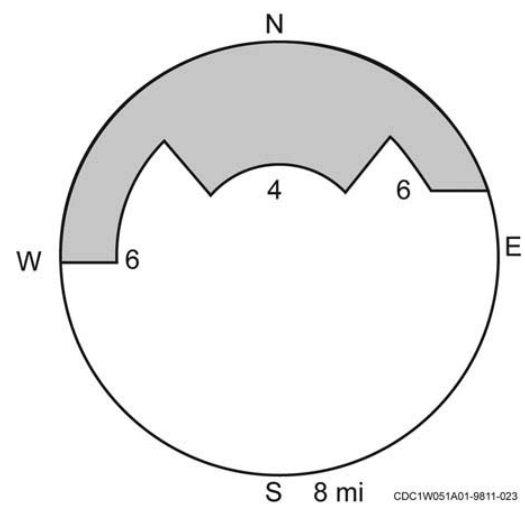

Figure 3–12. Sector visibility.

Use eight compass points to identify sector direction: N, NE, E, SE, S, SW, W, and NW. You can also use intermediate directions (NNE) if necessary, although you can describe most directions with the eight compass points. When more than one sector needs reporting, list the sectors in a clockwise direction, starting with the northernmost sector. As mentioned previously in this section, AMOS's cannot determine sector visibility. Sector visibility can only be determined while manually observing or when augmenting an AMOS.

Differing location visibility

To report differing location visibility, the prevailing visibility of either location (the official observing site or other reporting location) must be less than 4 miles with one location reporting a different prevailing visibility. Normally, the control tower reports visibilitys.

034. Characteristics of Runway Visual Range

Runway visual range (RVR) is an instrumentally derived value that represents the horizontal distance that a pilot can see down the runway. The maximum distance in the direction of takeoff or landing at which the runway, or specified lights or markers delineating it, can be seen from a position above a specified point on its center line at a height corresponding to the average eye level of pilots at touch-down. Stations disseminate RVR locally and longline using the values in the DOD flight information publications (FLIP) for RVR minima.

The AMOS's visibility sensor is also utilized to determine RVR. Additionally, the ambient light sensor provides accurate and reliable ambient light data for RVR calculations. The ambient light sensor faces north at Northern Hemisphere sites and south at Southern Hemisphere sites so it never faces the sun. RVR is reported in the body of the observation whenever the prevailing visibility is ≤ 1 statute mile (1600 meters) and/or the RVR for the designated instrument runway is ≤ 6000 feet (≤ 1500 meters when in OCONUS).

Reading RVR from observations

On a meteorological aviation report (METAR) observation, the RVR is reported after the surface visibility. A typical RVR reading looks like this: R27/1200FT. The "R" indicates runway heading, the two digits indicate the runway number. Sometimes the two numbers are followed by a letter such as "L," "C," or "R" to denote left, center, or right runway. This is only used for airfields which have multiple runways which run parallel to one another. In the example we gave, "27" indicates runway 27. The solidus simply separates the runway number from the visual range. The number after the "/" is the RVR either in feet as in our example or meters. The "FT" specifies the RVR is in feet. So, in our example the RVR for runway 27 is 1200 feet. Fairly easy isn't it?

When reading an observation from a station where RVR is reported in meters, the RVR is annotated by four numerals without the "FT" appended. An example is R27/1200. The runway is 27 and the RVR is 1200 meters. So, remember if after the four digits there's no letters, then the RVR is being reported in meters. If the letters FT are annotated after the four digits the RVR is being reported in feet.

In some cases RVR is required but won't be able to be reported. An example is when you're having equipment problems. When RVR can't be obtained because of equipment outages, report RVR not available (RVRNO) in the longline remarks.

Keep in mind the maximum and minimum RVR values that can be reported in feet are 6000 and 1000 respectively. The maximum and minimum RVR values that can be reported in meters are 1830 and 300 meters respectively. RVR sensing equipment can indicate that the RVR is greater than the maximum value and less than the minimum value. Report these situations using the letter "P" or "M" before the actual RVR value. Two examples of this would be R09/P6000FT or R09/M0300. The "P" simply stands for "plus" or "greater than" and the "M" for "minus" or "less than."

Like visibility, RVR can vary during the evaluation period for the observation. Varying RVR is reported similar to varying visibility by listing the lower value a "V" then the higher value. For example R18/1000V2000FT would indicate the RVR on runway 18 is varying from 1000 feet to 2000 feet.

As you can see, RVR is very important to pilots and gives them a more accurate indication of what visibility is doing on and near the runway.

Self-Test Questions

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

033. Characteristics of visibility

1. From the following list of visibility markers, indicate those best suited for night markers with an “N” and those best suited for day markers with a “D.” Add appropriate qualifying remarks after “N” or “D” as applicable. (**NOTE**: Some markers may only be good under certain conditions. Indicate under what conditions they might be poor. Also, indicate those makers that are only good as guides to estimation of visibility.)

   (a) A red light marker on a large building.

   (b) A dark brown house.

   (c) A line of trees.

   (d) An airfield runway beacon light on an Air Force base.

   (e) A white farmhouse and barn.

   (f) Smokestack of a manufacturing plant.

   (g) Red light on top of a TV tower.

   (h) A church steeple.

2. What’s the meaning of the term *prevailing visibility*?

3. When reporting sector visibility, how many compass points normally are used to identify sector direction? What are they?

4. What are the two requirements which must be met to report sector visibility?

5. When reporting sector visibility and more than one sector needs to be reported, how are the sectors listed?

6. Can you report the sector visibility if the sector visibility differs from prevailing visibility (but is more than 3 miles)? Explain your answer.

034. Characteristics of runway visual range.

1. What's the RVR for the following report: R09/1600?

2. Can you tell the exact RVR from the following report—R15/P6000FT? Explain.

3. What RVR is reported if the RVR is required to be reported but the equipment is inoperative?

Answers to Self-Test Questions

027

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

028

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

029

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

030

1. Because they're important to flight operations due to the association of mountain-wave turbulence with these types of clouds.
2. (a) Lenticular, M4.
   (b) Foehn wall, L5.
   (c) Rotor, L2.

031

1. Clouds or obscuring phenomena that have bases at the same approximate level. The layer may appear as continuous cover or as detached elements. Also, both continuous and detached elements may combine to form a layer.
2. Four-eighths of sky cover.
3. The lower of the two values.

032

1. The lowest broken or overcast layer aloft or the vertical visibility in a surface-based obscuring phenomena.
2. A ceilometer operating within its operational range and the known heights of unobscured portion of abrupt, isolated objects within $1^1/_2$ nautical miles of the runway.
3. Your answer should include any five of the following methods:
   (1) Heights reported by a pilot.
   (2) Based on the ascension rate of a ceiling balloon.
   (3) Convective cloud height table.
   (4) Based on the height of natural landmarks more than $1^1/_2$ miles from the airfield.
   (5) Your own observational experience.
   (6) Ceilometer indications that equal or exceed 10 times the baseline.
   (7) Based on the size of the cloud elements.
   (8) Reflection of city lights off of cloud bases.

033

1. (a) N (a good one).
   (b) D (a good one).
   (c) D (a good one).
   (d) N (but only to estimate visibility).
   (e) D (but is poor because of its color).
   (f) D (a good one).
   (g) N (a good one).
   (h) D (a good one).

2. The greatest distance known objects can be seen and identified throughout half or more of the horizon circle surrounding the station.
3. Eight. N, NE, E, SE, S, SW, W, NW.
4. (1) When sector visibility differs from prevailing visibility.
   (2) When sector visibility is less than 3 miles.
5. In a clockwise direction beginning with the northernmost sector.
6. Yes. If you consider it operationally significant.

034

1. On runway 09, the RVR is 1600 meters.
2. No, it's greater than 6000 feet but how much greater isn't known. The maximum reportable value has been exceeded.
3. RVRNO.

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.

73. (027) The difference in cloud classification between L1 and L7 is
    a. precipitation.
    b. vertical development.
    c. the size of the rain drops.
    d. size of the cloud elements.

74. (027) Which low cloud type and classification is identified by the presence of a cirriform anvil?
    a. Cumulus-L1.
    b. Cumulonimbus-L9.
    c. Cumulonimbus-L3.
    d. Towering cumulus-L2.

75. (027) The *best* way to distinguish stratocumulus from altocumulus clouds is to use
    a. movement.
    b. appearance.
    c. the color of the elements.
    d. the size of the elements.

76. (028) A corona is often present at night with
    a. stratus clouds.
    b. altostratus clouds.
    c. altocumulus clouds.
    d. stratocumulus clouds.

77. (029) Cirrus clouds in the form of an anvil are classified as
    a. H4.
    b. H3.
    c. H2.
    d. H1.

78. (029) Which high cloud can occasionally be so transparent that the only indication of its presence is a halo phenomenon?
    a. H1.
    b. H4.
    c. H7.
    d. H8.

79. (029) Which high cloud classification is also referred to as a mackerel sky?
    a. Cirrus-H4.
    b. Cirrostratus-H6.
    c. Cirrostratus-H7.
    d. Cirrocumulus-H9.

80. (030) Which orographic cloud resembles an almond or a fish?
    a. Rotor.
    b. Lenticular.
    c. Foehn wall.
    d. Castellanus.

81. (031) “Clouds or obscuring phenomena that have bases at the same approximate level” is the definition of
    a. sky condition.
    b. sky cover.
    c. a ceiling.
    d. a layer.

82. (032) What color balloon would you use to determine the ceiling heights of thin clouds?
    a. Red.
    b. Blue.
    c. Black.
    d. Yellow.

83. (032) You are preparing to use a convective cloud height table to determine the heights of clouds. To do this properly, you *must first* determine the
    a. wet-bulb temperature and free-air temperature.
    b. dew-point temperature and free-air temperature.
    c. wet-bulb temperature and dew-point temperature.
    d. dew-point temperature and dry-bulb temperature.

84. (033) How do you report prevailing visibility at US stations and overseas stations?
    a. Meters for both.
    b. Statute miles for both.
    c. Statute miles for US stations and meters for overseas stations.
    d. Meters for US stations and statute miles for overseas stations.

85. (033) To properly report the visibility for more than one sector, you would list the sectors in a
    a. clockwise direction starting with the northernmost sector.
    b. clockwise direction starting with the southernmost sector.
    c. counterclockwise direction starting with the northernmost sector.
    d. counterclockwise direction starting with the southernmost sector.

86. (034) For a runway visual range (RVR) report of R22/1000V1600FT, what is the visual range that a pilot can expect to see down the runway?
    a. 1,000 feet.
    b. 1,600 feet.
    c. 1,300 feet.
    d. 1,000 to 1,600 feet varying.

Please read the unit menu for unit 4 and continue ➔

Unit 4. Other Weather Observation Elements

TO THIS POINT, you've seen how pilots and weather craftsmen can use the sky condition and visibility entries to make decisions. Weather and obstructions to vision aren't only a continuation of this knowledge but are also directly related to sky condition and visibility. The latter, visibility, is determined by the type of weather or obstruction to vision present. By knowing the type of phenomenon restricting visibility, you can make a better prediction of the future visibility. Pilots use this information to determine what impact, if any, the weather phenomenon will have on their aircraft. For instance, correctly reporting freezing precipitation is of utmost concern to the pilots because their aircraft could develop aerodynamic problems from icing. In this unit we cover precipitation, weather, and obstructions to vision.

4–1. Precipitation

In this section, we examine the techniques used to identify, classify, and measure different forms of precipitation. To do this, we'll cover these three major subject areas:

- Forms of precipitation.
- Classification of the forms of precipitation.
- How to measure precipitation.

035. Forms of precipitation

Precipitation falls in many forms. Generally the various types are classified into these three main forms:

- Liquid.
- Freezing.
- Frozen.

Liquid precipitation

Rain and drizzle are the only two forms of liquid precipitation. Rain is distinguished from drizzle by the size of the water droplet (particle) and the spacing between droplets. Rain droplets have a diameter usually greater than 0.02 inch (0.5 mm), whereas drizzle has a droplet size less than 0.02 inch (0.5 mm). The drizzle droplets are very close together and appear to float with the air currents. Drizzle, frequently accompanied by low visibility and fog, usually falls from low stratus clouds.

Freezing precipitation

Freezing precipitation is liquid precipitation that falls and freezes upon impact with objects on the ground or in flight. Usually, freezing precipitation is caused by supercooled water particles, but it may occur when the surface is cold enough to freeze water particles that are near freezing. When water particles do freeze upon impact with the ground or objects in flight or on the ground, classify the precipitation as either freezing rain or freezing drizzle.

Frozen precipitation

For observation purposes, classify frozen precipitation into the following forms:

- Ice pellets.
- Hail.
- Snow.
- Snow pellets.
- Snow grains.
- Ice crystals.

Ice pellets

Ice pellets are either transparent or translucent particles of ice that are round or irregular in shape (rarely conical) and have a diameter of 0.2 inch (5 mm) or less. Ice pellets form by two different processes. If continuous precipitation (such as rain or melted snowflakes) freezes, the result is a transparent ice pellet (formerly called *sleet*). Snow pellets that become encased in a thin layer of ice are classified as *ice pellets*. This occurs when a snow pellet begins to melt and refreeze, or it may occur when snow pellets contact water droplets while they're falling. Then, the water freezes, producing a thin layer of ice around the snow pellet. This type falls as showers. Ice pellets usually rebound when they strike hard ground and make a sound on impact.

Hail

Hail is distinguished from other frozen precipitation by its irregular shape and generally large size. Hail falls almost exclusively from strong convective clouds (cumulonimbus) that are usually accompanied by thunder. Some hailstones consist of alternately opaque and clear layers of ice, which are formed by the strong up and down drafts within the cloud. On occasion, hailstones freeze together and fall in irregular lumps. When hail falls at the station, you must determine the size of the largest available hailstone. Hail seldom occurs when surface temperatures are near or below freezing.

Snow

Snow is precipitation in the form of ice crystals, most of which appear branched as six-pointed stars. At temperatures higher than about 23° F, the crystals are generally clustered to form snowflakes.

Snow pellets

Snow pellets are white, opaque grains of ice. The grains are round or sometimes conical. Diameters range from about 0.08 inch (2 mm) to 0.2 inch (5 mm). Snow pellets are brittle and easily crushed. When they strike hard surfaces, they bounce and often break apart. When conditions are right, snow pellets serve as the nuclei for hail development. Snow pellets form exclusively in convective clouds which produce showery precipitation.

Snow grains

Snow grains are very small, white, opaque grains of ice, similar in structure to snow. The primary difference is the smallness of each element and the fairly flat or elongated shape of the snow grains in comparison to snow. They don't burst or shatter when they strike hard surfaces. Snow grains usually fall in small quantities, mostly from stratus clouds, and never as showers. Snow grains are the solid equivalent of drizzle.

Ice crystals

Ice crystals are unbranched and occur in the form of needles, columns, or plates. They're often so tiny that they appear to be suspended in the air. Ice crystals, also termed "ice prisms," may fall from a cloud or from clear air. The crystals are visible mainly when they glitter in the sunshine or other bright light (diamond dust). They may produce a luminous pillar or other optical phenomena. Ice crystals (frequently seen in polar regions) occur only at very low temperatures in stable air masses.

036. Classification of precipitation forms

Knowing the type of precipitation associated with low ceilings and visibilitys is made even more meaningful when we add the character and intensity of precipitation. Pilots are very interested in knowing the presence of rain showers versus rain. Rain showers warn them to expect greater fluctuations in visibility and usually more turbulence in showery precipitation as they land and take off from an air base. The decision of whether the precipitation is showery or continuous, light or moderate is determined by the observer's judgment based on experience and established guidelines. In this lesson, we cover two important subject areas:

- Precipitation character.
- Intensity of precipitation.

Precipitation character

Precipitation character is based upon established criteria such as the type of cloud originating the precipitation. The character of precipitation is divided into these three categories:

- Continuous.
- Intermittent.
- Showery.

Continuous

Continuous precipitation increases or decreases gradually in intensity, if at all. Precipitation of this character is usually associated with stratiform cloud types such as:

- Altostratus.
- Nimbostratus.
- Stratus.

Intermittent

Intermittent precipitation also increases or decreases gradually in intensity. To be classified as intermittent, the precipitation must stop and start at least once within the hour preceding the observation. This category of precipitation is used with precipitation types not classified as showery.

Showery

Showery precipitation changes intensity rapidly, or the shower begins or ends abruptly. These clouds produce showery precipitation:

- Swelling cumulus.
- Cumulonimbus.

Intensity of precipitation

Intensity of precipitation is an indication of the amount of precipitation falling at the time of observation. Each intensity is defined with respect to the type of precipitation occurring.

Determination of precipitation intensity is a rather subjective process. There are published standards to aid you in the process. Tables 1, 2, 3, and 4, are reproductions of tables found in AFMAN 15–111, *Surface Weather Observations*. These tables make your job of precipitation intensity determination much easier.

Table 1. Estimating intensity of rain or ice pellets based on a rate-of-fall basis.

Intensity	Criteria
Light	Up to 0.10 inch per hour; maximum 0.01 inch in 6 minutes.
Moderate	0.11 inch through 0.30 inch per hour; more than 0.01 inch through 0.03 inch in 6 minutes.
Heavy	More than 0.30 inch per hour; more than 0.03 inch in 6 minutes.

Table 2. Estimating the intensity of rain or freezing rain.	
Intensity	**Criteria**
Light	From scattered drops that, regardless of duration, don't completely wet an exposed surface up to a condition where the individual drops are easily seen.
Moderate	Individual drops aren't clearly identifiable; spray is observable just above pavements and other hard surfaces.
Heavy	Rain seemingly falls in sheets; individual drops aren't identifiable; heavy spray to a height of several inches is observed over hard surfaces.

Table 3. Intensity of snow or drizzle based on visibility (occurring alone with no other obscurations).	
Intensity	**Criteria**
Light	Visibility > $^1/_2$ mile (800 meters).
Moderate	Visibility > $^1/_4$ mile (400 meters) but ≤$^1/_2$ mile (800 meters).
Heavy	Visibility ≤$^1/_4$ mile (400 meters).

Table 4. Estimating intensity of Ice pellets.	
Intensity	**Criteria**
Light	Scattered pellets that do not completely cover an exposed surface regardless of duration; visibility is not affected.
Moderate	Slow accumulation on ground; visibility reduced by ice pellets to less than 7 statute miles (9999 meters).
Heavy	Rapid accumulation on ground; visibility reduced by ice pellets to less than 3 statute miles (4800 meters).

Frequently, the total precipitation (water equivalent) for the day isn't supported by the intensities you report during the day. For example, suppose you carry moderate continuous rain interspersed with short periods of light rain for a 6-hour period. After the 6-hour period, your measurement was only 1 inch. Since an intensity of moderate for an entire 6-hour period should yield between 0.66 and 1.80 inches (based on table 1), the intensity of moderate you carried was correct. Remember, a check of precipitation amounts for a 6-hour period and for the day is a good indication of whether you're entering the correct intensities.

037. Precipitation measurement

Precipitation measurement is accomplished at automated and manual stations, which are described in this lesson.

Automated stations

Measurement of precipitation is accomplished automatically through the use of an automated meteorological observing system (AMOS). The AMOS's precipitation gauge allows repeatable liquid precipitation measurements. Most likely your station will have an FMQ–19 AMOS. The FMQ–19 rain gauge has a .01" capacity before tipping and reporting precipitation amount. The FMQ–19's precipitation gauge is mounted on a pedestal. The FMQ–19 uses a freezing rain sensor to detect freezing precipitation. This sensor operates in temperatures as cold as –30 degrees Celsius.

Manual stations

Precipitation measurements at manual observing locations are normally made by means of a standard 8-inch rain and snow gauge, the ML–17 (rain gauge). The ML–217 (a 4-inch plastic gauge) and automatic precipitation measuring devices also are used.

Liquid precipitation

To measure liquid precipitation, determine the amount of precipitation by measuring the collection in the rain gauge using a measuring stick to the nearest .01-inch. A trace is reported when the measureable amount is less than .005 of an inch.

The ML–17 measuring tube has a 2 inch precipitation capacity. When more than 2 inches of precipitation has occurred without emptying the rain gauge, the excess precipitation will spill into the rain gauge's overflow can. If this occurs carefully remove and empty the 2 inches of precipitation from the measuring tube. Pour the remaining liquid from the overflow container into the measuring tube. Record two inches of precipitation for every completely filled measuring tube. If the measuring tube is not completely filled up measure the precipitation remaining and add it to the previous measurements. For example, if your station received 5 inches of rain the measuring tube would be full. The additional three inches would have spilled into the overflow can. After removing the initial two inches of precipitation from the measuring tube you would have poured the excess precipitation from the overflow can filling it once again. After pouring out the second full measuring tube of water, the remaining precipitation would measure one inch, for a total of 5 inches.

Frozen precipitation

When frozen precipitation is expected, collect it in the overflow unit of the rain gauge and not in the measuring tube. Since you are interested in determining water equivalents of frozen precipitation, the frozen precipitation must be permitted to accumulate in the collection device unobstructed. Eventually, when it's time to measure, a measured amount of water will be added to the frozen accumulation to melt it and then the entire amount will be added to the measuring tube.

After the frozen precipitation is melted, the process of determining the water equivalent is very similar to measuring just straight liquid precipitation. First, measure the entire amount of liquid with the measuring stick to the nearest .01 of an inch. Second, determine the water equivalent by subtracting the amount of water that you added to the sample.

NOTE: This process may be more complicated in windy conditions.

During snowfall accompanied by strong or gusty winds, the amount collected in the overflow container may not represent actual snowfall. If the rain gauge collection is not representative, disregard the catch and (if possible) obtain the water equivalent by means of core sampling per AFMAN 15–111.

If this procedure isn't possible, estimate the water equivalent. To estimate water equivalent of frozen forms of precipitation, first obtain a measurement of the snowfall. Next, convert the actual depth to its water equivalent based on a 10:1 ratio (or other ratio if representative for the station).

For example, if 1.6 inches of snow has fallen, the water equivalent is approximately .16 inch (1.6 /10=0.16 by using the 10:1 ratio).

Self-Test Questions

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

035. Forms of precipitation

1. Name the two forms of liquid precipitation.

2. Define "freezing precipitation."

3. Match the precipitation form in column B with the description in column A. Items in column B may be used once, more than once, or not at all.

Column A	*Column B*
____ (1) Transparent or translucent particles of ice that is round or irregular in shape. They usually rebound when striking hard ground and make a sound on impact.	a. Drizzle.
____ (2) Very small, white, opaque grains of ice, which usually fall in small quantities, mostly from stratus clouds, and never as showers.	b. Ice pellets.
____ (3) Ice crystals that appear mostly branched in the form of six-pointed stars.	c. Snow pellets.
____ (4) Falls from strong convective clouds and occasionally freeze together, falling in irregular lumps.	d. Hail.
____ (5) Unbranched and in the form of needles, columns, or plates.	e. Freezing rain.
____ (6) Form exclusively in convective clouds and, under the right conditions, serve as the nuclei for hail development.	f. Snow grains.
	g. Ice crystals.
	h. Snow.

036. Classification of precipitation forms

1. What are the three categories of precipitation?

2. What category of precipitation is usually associated with stratiform cloud types?

3. What category of precipitation stops and starts at least once within the hour preceding the observation.

4. What category of precipitation do swelling cumulus and cumulonimbus clouds produce?

5. What precipitation category changes intensity rapidly or begins and ends abruptly?

6. Match the precipitation classification in column B with the description in column A. Items in column B may be used once or not at all.

Column A	*Column B*
____ (1) Individual raindrops aren't identifiable.	a. Heavy snow.
____ (2) Snow is falling. There are no obstructions to vision present. The prevailing visibility is ¼ mile.	b. Moderate snow.
____ (3) Ice pellets are falling and there are no obstructions to vision. The prevailing visibility is 5 miles.	c. Heavy rain.
____ (4) Rain is falling. Scattered drops don't completely wet an exposed surface.	d. Light rain.
____ (5) Drizzle is falling and there are no obstructions to vision present. The prevailing visibility is 1 mile.	e. Light drizzle.
	f. Moderate drizzle.
	g. Moderate ice pellets.
	h. Light ice pellets.

037. Precipitation measurement

1. What is the capacity of the FMQ–19's rain gauge before tipping?

2. Describe the process of measuring frozen precipitation.

4–2. Weather

Though the term "weather" is often used in a broad sense, in observing, it refers specifically to atmospheric phenomena which are reported in the body of a weather observation. In this section, we cover characteristics of storm-related phenomena such as tornadoes, funnel clouds, waterspouts, and thunderstorms.

038. Characteristics of tornadic and thunderstorm activities

Tornadoes, funnel clouds, and waterspouts are weather phenomena that occur in areas where intense thunderstorm activity is possible. These phenomena stem from the same cloud that produces thunderstorms—the cumulonimbus. This cloud spawns the frontal and air-mass thunderstorms usually characterized by thunder, lightning, strong wind gusts, heavy rain showers, and sometimes hail. Under certain conditions, the potentially destructive energy produced within a cumulonimbus mass is released as a whirling vortex beneath the cloud.

When the whirling vortex doesn't reach the ground, it's called a *funnel cloud*; when the vortex, with its low pressure and tremendous winds, touches the ground, it's called a *tornado*. If the vortex descends to the surface over water, it's called a *waterspout*.

The distinguishing feature of a tornado, funnel cloud, or waterspout is a funnel-shaped appendage that hangs from the base of the cloud. Sometimes thunderstorms are in progress at the time the funnel descends and precipitation prevents easy detection of the funnel cloud or tornado. Depending on the distance from the point of observation, funnel cloud or tornado identification ranges from the obvious to the doubtful. For example, the ragged appearance of cumulus fractus clouds that are frequently in the area during thunderstorm activity may suggest a funnel-shaped appendage. Since these cloud elements usually change their appearance rapidly, close observation for a short time usually resolves the question of whether a funnel actually exists. Your judgment, based on all available information, is the key ingredient to proper identification.

Significant remarks for storm phenomena provide added information. Storm phenomena present a constant threat to the public and to flying operations. Your remarks on tornadic activity alert pilots to the location of the tornadic activity, the direction in which it's moving, and any other necessary information.

Thunderstorm activity

Thunderstorm activity, though not as serious as tornadoes, presents many hazards to flight operations. For observing purposes, a thunderstorm is present and occurring at the station when:

- Thunder is first heard.
- Hail is falling or lightning is observed in the immediate vicinity of the airfield and the local noise level prevents you from hearing the thunder.
- Lightning detection equipment indicates lightning strikes w/in 5nm of the airfield.

You should encode the location of each storm center (with respect to the station) to include the distance, if known, and the direction toward which the storm is moving (or moved), if known. You should continue to report the thunderstorm in this manner until 15 minutes after the last occurrence of thunder, hail, or lightning. Subsequently, the thunderstorm will be reported as ended. It's important to understand how to report thunderstorm activity in regards to proximity to your station. When an AMOS is used, thunderstorms will be reported at the station if the sensor detects cloud-to-ground strikes within 5 nautical miles of the point of observation and in the vicinity of the station if the cloud-to-ground strikes occur between 5 and 10 nautical miles of the observing point. Cloud-to-ground lightning strikes detected beyond 10 nautical miles are reported as distant. These distances are also applied when taking a manual observation.

Lightning

Lightning, though not considered as weather, is associated with thunderstorms. Lightning can occur in the cloud, from the cloud to air, from the cloud to the ground, and from the cloud to another cloud.

Hail

Whenever thunderstorms and lightning are present, hail is very possible. Large hail can cause extensive damage to aircraft structures. When you observe hail at your station, be sure to include a remark on your observation.

Self-Test Questions

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

038. Characteristics of tornadic and thunderstorm activities

1. What's the term for a whirling vortex which doesn't reach the ground?

2. What's the term for a whirling vortex which descends to the surface over water?

3. What's the term for a whirling vortex which touches the ground?

4. What distinguishing feature do questions 1, 2, and 3 have in common?

5. What criteria must be met for tornadic activity to be classified as occurring "at the station?"

6. When can a thunderstorm be included in an observation even if thunder isn't heard?

7. When does a thunderstorm end?

4–3. Obstructions to Vision

The term *obstructions to vision* includes all other types of atmospheric phenomena not considered "weather." Since visibility is affected by obstruction to vision phenomena, the weather craftsman studies reports of obstructions at the local station as well as reports from surrounding stations. This data and this knowledge of the meteorological factors that influence changes to obstructions to vision are extremely important aids in flight operations and scheduling. In this section we'll look at the techniques used to identify various obstructions to vision.

039. How to identify various obstructions to vision

All obstructions to vision are classified as either *hydrometeor* or *lithometeor*. They aren't encoded on the observation unless they restrict visibility to 9,000 meters or less. Remember that when obstructions to vision cover $^1/_8$ or more of the sky, they're considered as sky cover.

Hydrometeors

Hydrometeors are atmospheric phenomena that consist of liquid or frozen water particles. When these particles are falling, they're called *precipitation*. When they're suspended in the atmosphere, they're called *obstructions to vision*. For observing purposes, five hydrometeors are considered obstructions to vision:

- Mist.
- Fog.
- Blowing snow.
- Freezing fog.
- Blowing spray.

Mist

Mist is a visible aggregate of minute water particles suspended in the atmosphere that reduces surface visibility. Mist is similar to fog; however, report it as mist when the visibility is less than 7 statute miles (9999 meters) but greater than or equal to 5/8 statute miles (1000 meters).

Fog

Fog is a suspension of small water droplets in the air that reduce horizontal visibility at earth's surface. Fog is distinguished from other obstructions to vision by its dampness and gray appearance. Usually fog doesn't form or exist when the difference between the temperature and dew point is greater than 4°F (2°C); however, report fog whenever you observe it. When temperatures are below freezing, the difference may exceed 4°F. Heavy fog sometimes produces rime or glaze ice on cold, exposed objects. Fog is similar to mist; however, report it as fog when the visibility is less than 5/8 statute miles (1000 meters).

Blowing snow

Blowing snow exists when the wind blows snow to moderate or great heights. Blowing snow is closely related to drifting snow. The main difference is that blowing snow restricts visibility (9,000 meters or less) and the sky may become obscured when the particles are raised to great heights. Drifting snow doesn't reduce visibility to 9,000 meters or less at eye level.

Freezing fog

Freezing fog is a rare form of fog because it usually forms at temperatures below freezing. Freezing fog, also called *ice fog*, can produce rime or glaze on cold objects. It has elements very similar to ice crystals except that freezing fog particles are suspended in the atmosphere. Ice fog produces optical effects similar to those produced by ice crystals, such as halo phenomena, luminous vertical columns, or sparkling effect.

Blowing spray

Blowing spray is reported only at sea stations near large bodies of water. To be reported, the spray, which is water droplets that are blown from the water by the wind, must restrict the visibility at eye level (6 ft on shore, 33 ft at sea) to 9,000 meters or less. Unless you're assigned to a station near a large body of water, you'll never report this obstruction to vision in your observation.

Lithometeors

All obstructions to vision that don't have a water composition (hydrometeor) and aren't classified as "weather" are called *lithometeors*. They're classified into the following five separate types:

1. Dust.
2. Sand.
3. Haze.
4. Smoke.
5. Volcanic ash.

Dust

Dust is finely divided earthly matter uniformly distributed in the atmosphere. It can be distinguished from other lithometeors by the tan or gray tinge it gives to distant objects. When dust is present, the sun's disk is pale and colorless or has a yellow tinge. Dust can be broken down further into these four types:

- Blowing dust.
- Drifting dust.
- Duststorm.
- Well-developed dust whirls.

Blowing dust

Blowing dust is dust the wind picks up from the surface and blows about in clouds or sheets. To be classified as blowing dust, the dust must be raised to the height of 6 feet or more and restrict horizontal visibility to 9,000 meters or less.

Low Drifting dust

Drifting dust is the same as blowing dust except it's raised by the wind to less than 6 feet above the ground.

Duststorm

A duststorm is an unusual, but frequently severe weather condition characterized by strong winds and dust-filled air over an extensive area. You should report duststorms if the prevailing visibility is reduced to less than $^5/_8$ miles, but not less than $^5/_{16}$ miles. Report a severe (heavy) duststorm if the visibility is reduced to less than $^5/_{16}$ miles.

Well-developed dust whirls

Dust whirls are a collection of particles of dust, sometimes accompanied by small litter, that are raised from the ground in the form of a whirling column of varying height with a small diameter and an approximately vertical axis. They're commonly called *dust devils*.

Sand

Particles of sand raised to a sufficient height that reduces visibility.

Sandstorms

Sandstorms consist of particles of sand ranging in diameter from 0.008 inches to 1 millimeter and carried aloft by a strong wind. The sand particles are mostly confined to the lowest 10 feet and rarely rise more than 50 feet above the ground. You should report sandstorms if the prevailing visibility is reduced to less than $^5/_8$ miles, but not less than $^5/_{16}$ miles. Report a severe (heavy) sandstorm if the visibility is reduced to less than $^5/_{16}$ miles.

Blowing sand

Blowing sand is sand the wind picks up from the surface and blows about in clouds or sheets. To be classified as blowing sand, the sand must be raised to the height of 6 feet or more and restrict horizontal visibility to 9,000 meters or less. Two types are possible:

1. Drifting sand.
2. Well-developed sand whirls.

Low Drifting sand

Drifting sand is the same as blowing sand except it's raised by the wind to less than 6 feet.

Well-developed sand whirls

Well-developed sand whirls are a collection of particles of sand, sometimes accompanied by small litter, which are raised from the ground in the form of a whirling column of varying height with a small diameter and an approximately vertical axis. These columns are commonly called *dust devils*.

Haze

Haze is a suspension in the air of extremely small, dry particles, such as salt, dust, or pollen. They're invisible to the naked eye and sufficiently numerous to give the air a milky appearance. In spite of the fineness of haze particles, haze restricts visibility.

Over the landscape, haze resembles a uniform veil that subdues natural colors (such as green trees) along the horizon. When viewed against a dark background such as a mountain, haze produces a bluish tinge. It causes a dirty yellow or orange tinge against a bright background such as the sun, clouds on the horizon, or snowcapped mountain peaks. When the sun is well above the horizon, its light sometimes has a peculiar silvery tinge because of haze. These color effects distinguish haze from thin fog, even when the thickness of haze approaches that of thin fog.

Smoke

Smoke is a very common obstruction to vision near large cities and industrial areas. It has fine particles that are produced by combustion. When we view the disk of the sun through smoke at sunrise or sunset, it appears very red. When the sun is above the horizon, it may have an orange tinge. Evenly distributed smoke from distant sources generally has a light grayish or bluish appearance. A transition to haze may occur when smoke particles have traveled great distances (for example, 25 to 100 miles or more) and when the larger particles have settled out and the remaining particles have become widely scattered through the atmosphere.

Volcanic ash

Volcanic ash is fine particles of rock powder that have erupted from a volcano and remain suspended in the atmosphere for long periods of time. Volcanic ash is always reported when observed regardless of the visibility.

Self-Test Questions

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

039. How to identify various obstructions to vision

1. List the five hydrometeors.

2. List the five lithometeors.

3. Classify the specific hydrometeor or lithometeor described in each of the following statements:

   (a) Small water droplets suspended in the air, visibility ¼ statute mile.

   (b) Reported only at sea stations near large bodies of water.

   (c) Finely divided earthly matter uniformly distributed in the atmosphere.

   (d) Loose sand blown by the wind and restricting visibility to 4,800 meters.

   (e) A uniform veil that subdues natural colors.

   (f) The disk of the sun appears very red at sunrise and the visibility is 8,000 meters due to a suspension of particles in the air.

4–4. Pressure, Temperature, Dew Point, and Wind

The prime consideration for any major product analysis that you'll make is the distribution of atmospheric pressure. You're also concerned with the temperature and dew point differences between air masses and the direction and speed of the wind. The material we present in this section will cover these four areas: pressure, temperature, dew point, and wind.

040. Types of pressure measurements

When dealing with pressure measurements, you must first determine a basic pressure value that's used for the computation of the other two operational pressure values (station pressure).Second, weather craftsmen need a pressure value they can use in the analysis of weather systems and the prognosis of these weather systems which removes the bias of variable terrain (sea-level pressure). Third, pilots must have a pressure value that provides them with a reliable indication of their in-flight altitude above sea level (altimeter setting). In this lesson, we focus on three pressure concepts:

1. Station pressure.
2. Sea-level pressure (SLP).
3. Altimeter setting (ALSTG).

In the weather career field, these are the three different types of pressure measurements used on observations.

Station pressure

Station pressure is the actual atmospheric pressure at your station in inches of mercury. This concept is illustrated on figure 4–1. This value, although not transmitted locally or longline, is the *basis* for determining other pressure values.

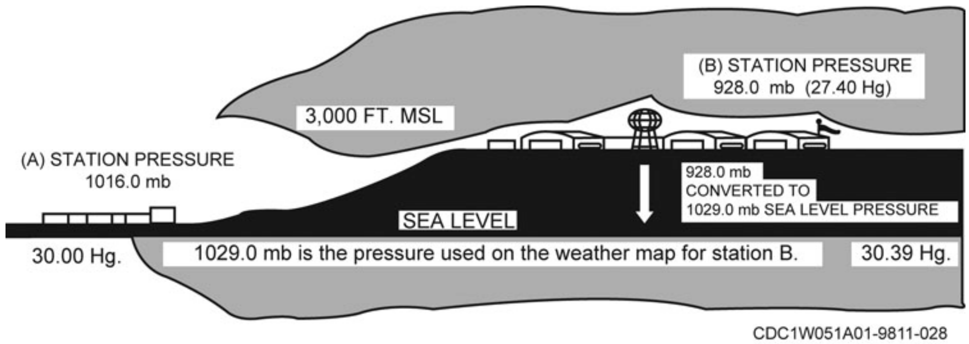

Figure 4–1. Station pressure.

Atmospheric pressure readings are obtained at base weather stations from the AMOS or tactical field equipment. Of the two, the AMOS is considered the primary pressure measuring instrument.

NOTE: Atmospheric pressure is "the pressure exerted on a specified unit area by a column of air extending vertically from the reference surface to the top of the atmosphere."

Sea-level pressure

SLP is the pressure at mean sea level. It's directly measured at sea level. If not at sea level, SLP is calculated. In either case, it is the *reference level* for *all* pressure values.

You should calculate the SLP from the following:

- Station pressure.
- The 12-hour mean temperature.
- Station elevation.

Plot the SLP on surface maps so that there's a standard pressure level. This standard pressure level is needed so you can compare the pressure from several stations for one given time.

Altimeter setting

The ALSTG is a *calculated* sea-level pressure in inches of mercury (Hg). The pilot uses it to adjust the altimeter of the aircraft. If the altimeter is set to the current setting, it indicates the field elevation when the aircraft is parked on the runway.

The altimeter in an aircraft is an aneroid barometer, which is calibrated to indicate altitude instead of pressure. For example, it indicates 10,000 feet when the pressure is 20.58 inches (this will occur regardless of whether or not the altitude is actually 10,000 feet). When the altimeter is properly adjusted for the current setting, the indicated altitude corresponds to the equivalent pressure in the actual atmosphere. Similarly, a pilot flying between a high-pressure (warm) area and a low-pressure (cold) area with a constant setting cranked into the altimeter finds that the true altitude varies both above and below the indicated altitude. Figure 4–2 shows this. So, a pilot changes the setting of the altimeter according to the changes supplied by air traffic controllers along the flight path. In this way, the pilot can use the altimeter to maintain a reasonably true altitude. This helps to eliminate the hazard of flying into other aircraft or mountains because of altimeter error.

NOTE: It's important to remember that ALSTG is based on field elevation.

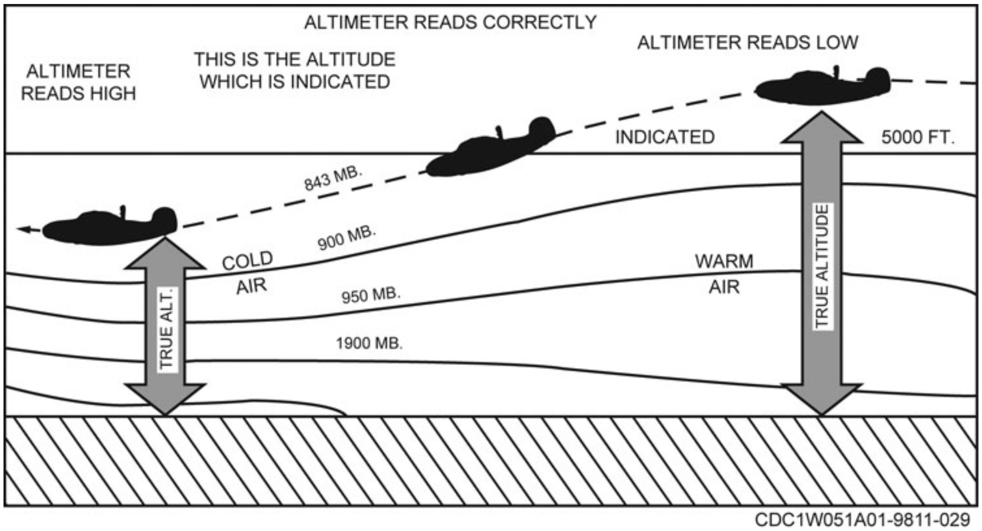

Figure 4–2. Altimeter setting.

041. Characteristics of temperature, dew point, and relative humidity

Temperature is one of the most common and easily understood elements in a weather observation. Besides its usefulness as a tool for analyzing frontal positions, the pilot uses temperature data, along with pressure-altitude, to compute the runway distance needed to reach takeoff speed. In this lesson, we'll explore these three topic areas:

- Temperature.
- Dew point.
- Relative humidity.

Temperature

Temperature is the measure of the average kinetic energy of the molecules of the air. It's commonly measured according to the Fahrenheit (surface) and Celsius (upper air) scales. The temperature of the air is sometimes called *ambient temperature*. This means air is freely moving about, unaffected by controlled heating or cooling sources.

Dew point

The dew point indicates the temperature to which a given parcel of air must be cooled, with constant water vapor content and pressure, to reach saturation. The dew point is important because it's the temperature beyond which further cooling produces visible condensation. To calculate the dew point, you use the following data:

- Dry-bulb temperature.
- Wet-bulb temperature.
- Wet-bulb depression.

Dry-bulb temperature

This is the ambient temperature registered by an ordinary dry-bulb thermometer. It's identical with the temperature of the air and may be used in that sense.

Wet-bulb temperature

The wet-bulb temperature of the air is the temperature as measured by a wet-bulb thermometer. A wet-bulb thermometer is an ordinary dry-bulb thermometer whose bulb is enclosed in a wetted sac or wick. The wet-bulb is the temperature an air parcel would have if cooled adiabatically to saturation at constant pressure by evaporation of water into it. It differs from the dry-bulb temperature by an amount dependent on the moisture content of the air and, therefore, is generally the same as or lower than the dry-bulb temperature.

Wet-bulb depression

Wet-bulb depression refers to the difference between the wet-bulb and dry-bulb temperatures. You use the dry-bulb temperature and the wet-bulb depression to calculate the dew point. When you've determined the value of the dew point, it's possible to calculate the relative humidity.

Relative humidity

Relative humidity is the ratio (expressed as a percentage) of the actual vapor pressure of the air to the saturation vapor pressure.

As a weather apprentice, it is important that you understand the terms we've discussed. In addition, you must have a thorough knowledge of the indicators and how they correlate with each other. For example, the closer the temperature and dew point are together, the greater the amount of moisture in the atmosphere. This greater amount of moisture can lead to a decrease in visibility.

042. Characteristics of wind

Air has characteristics, such as temperature, moisture, and movement. Air movement (wind) sometimes produces unusual conditions in certain areas. Because they're unusual, these conditions have been given names. Here are some examples:

- "Chinook" on the leeward side of the Rockies.
- "Nor'easter" in New England.
- "Santa Ana" in California.

Some winds affect large areas, whereas others occur on a local scale. Whatever their extent, your wind observations should include theses three factors:

- Direction.
- Wind speed.
- Wind character.

Wind direction

The direction from which the wind is blowing gives its name to the wind. A west wind is one coming from the west. Two geographic points—the true North Pole and the magnetic north pole—are references for any direction. The observing equipment is oriented to magnetic north; however, AFW observation entries require true north orientation. This is an important fact because when you take an observation, you must convert from magnetic north to true north. However, when you brief a pilot—always brief any takeoff winds as magnetic. This is also the case if the pilot is performing a local training flight or doing touch-and-go landings. In this case, list the wind for landing as magnetic.

We'll now cover the two conversion methods you'll use:

1. Converting true direction from magnetic direction.
2. Converting magnetic direction from direction.

Converting true direction from magnetic direction

The line along which the true and magnetic directions are the same is called the agonic line or 0° magnetic variation. The variations on one side of the 0° line are termed easterly variations and variations on the other side of the line are westerly variations. To convert magnetic wind direction to true direction, refer to figure 4–3 and perform these two actions:

- ADD EASTERLY variation to magnetic direction.
- SUBTRACT WESTERLY variations from magnetic direction.

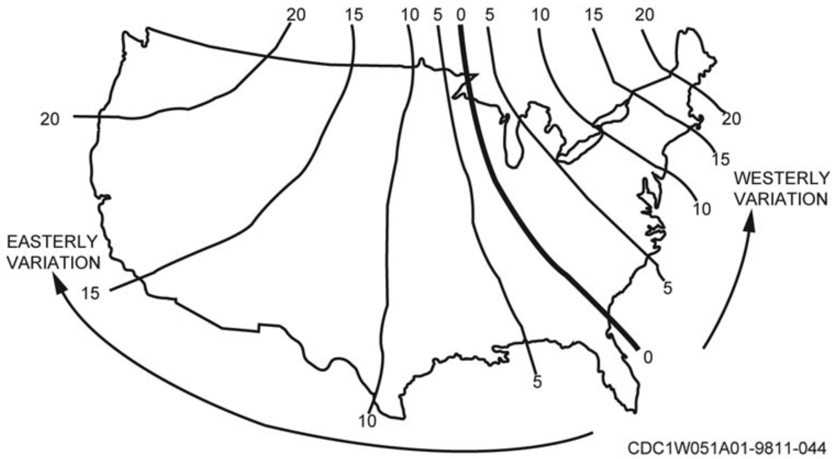

Figure 4–3. Magnetic variation/corrections.

Converting magnetic direction from true direction

If you find it necessary to convert true direction to magnetic direction, then reverse the above procedure.

Earth's magnetic field is continually shifting and local variation changes by several minutes of arc each year at most localities. You should monitor revised geomagnetic products regularly for any changes in local variations.

The direction is determined over a 2-minute period and read to the nearest 10°.

Rarely does the wind blow from a single direction; it usually swings across a wide range of directions. You simply choose the direction that occurs most frequently during the observation period.

Wind speed

The speed is determined over a 2-minute period and is read to the nearest knot. The AMOS is the primary instrument used by weather observers to determine the wind speed. When the AMOS isn't available, you have to obtain wind measurements from other devices, such as control tower readouts using a 2-minute average.

Estimating wind direction and speed

On occasion, it may be necessary to estimate the wind direction and/or the speed. When this situation arises, use free-moving objects to determine the wind direction. If a wind cone or tree is available, you should have little trouble in estimating the direction. Estimating the wind speed can pose more of a problem.

When instruments aren't available, wind speed is estimated using the Beaufort wind scale shown in the table below.

Beaufort Number	Description	Knots	Specifications for Estimating Windspeed
0	Calm	Less than 1	Calm; smoke rises vertically.
1	Light air	1–3	Direction of wind shown by smoke drift but not by wind vanes.
2	Light breeze	5–6	Wind felt on face; leaves rustle; vanes moved by wind.
3	Gentle breeze	7–10	Leaves and small twigs in constant motion; wind extends light flag.
4	Moderate breeze	11–16	Raises dust and loose paper; small branches are moved.

Beaufort Number	Description	Knots	Specifications for Estimating Windspeed
5	Fresh breeze	17–21	Small trees in leaf begin to sway; crested wavelets form on inland waters.
6	Strong breeze	22–27	Large branches in motion; whistling heard in telegraph wires; umbrellas used with difficulty.
7	Near gale	28–33	Whole trees in motion; inconvenience felt when walking against the wind.
8	Gale	35–40	Breaks twigs off trees; generally impedes progress.
9	Strong gale	41–47	Slight structural damage occurs.
10	Storm	48–55	Trees uprooted; considerable structural damage occurs.
11	Violent storm	56–63	Accompanied by widespread damage.
12	Hurricane	64 +	

Wind character

A simple report of direction and speed doesn't always represent the observed wind. The terms "gusts," "squall," and "wind shift" help describe the characteristics of the wind not revealed by direction and speed.

Gusts

A gust is a sudden intermittent increase in windspeed with at least a 10-knot variation between peaks and lulls within the past 10 minutes. Gustiness complicates aircraft touchdown and takeoff maneuvers in a way similar to, but much more serious than, the handling affects you feel while driving a car during gustiness. Further importance is added to gusts by their association with frontal passage and thunderstorms.

Squalls

A squall is distinguished from a gust by a sudden increase of windspeed of at least 16 knots and a sustained average of 22 knots or more maintained for at least 1 minute before the speed diminishes. Squalls usually indicate that turbulence is present near earth's surface and, like gusty winds, poses a problem to flight operations. Although they're a wind event, report squalls as a present weather event.

Wind shift

Wind shift is a change in wind direction of 45°or more that takes place in less than 15 minutes. You should report wind shifts when the speed throughout the shift is 10 knots or more.

Since a wind shift occurs over a period of time, you can't report it until the complete change in direction has actually taken place. The solution is to report the wind shift in the first observation following the complete occurrence of the phenomenon. If the wind shift is the result of a frontal passage, the acronym "FROPA" will be included in the body of the weather observation.

Variable wind direction

Variable wind direction is a condition in which the wind direction is fluctuating by 60° or more during the observation. This condition isn't considered as a reportable remark when the windspeeds are 6 knots or less.

Self-Test Questions

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

040. Types of pressure measurements

1. What pressure concept is the *basis* for determining other pressure values?

2. Which one of the three main pressure *concepts* is the reference for all other pressure values?

3. Which pressure concept is a *calculated* sea-level pressure in inches of mercury?

041. Characteristics of temperature, dew point, and relative humidity

1. What does temperature measure?

2. What indicates the temperature to which a given parcel of air must be cooled with constant water vapor content and pressure to reach saturation?

3. What’s the wet bulb depression?

4. Relative humidity is the ratio expressed as a percentage of what factors?

042. Characteristics of wind

1. How do you convert magnetic north wind directions to true north?

2. Over what period of time is wind direction determined? How is it read?

3. Over what period of time is windspeed determined? How is it read?

4. When wind instruments aren’t available, what’s one way to estimate the windspeed?

5. What’s a gust?

6. Define a squall.

7. What’s a wind shift?

8. What constitutes a variable wind direction?

Answers to Self-Test Questions

035

1. Rain and drizzle.
2. Liquid precipitation that falls and freezes upon impact with objects on the ground or in flight.
3. (1) b.
   (2) f.
   (3) h.
   (4) d.
   (5) g.
   (6) c.

036

1. (1) Continuous.
   (2) Intermittent.
   (3) Showery.
2. Continuous.
3. Intermittent.
4. Showery.
5. Showery.
6. (1) c.
   (2) a.
   (3) g.
   (4) d.
   (5) e.

037

1. .01 of an inch.
2. Remove the collector-funnel unit and measuring tube. Pour a measured amount of water into the overflow can to melt the solid precipitation, pour the melted liquid into the measuring tube, measure the total amount of liquid, and subtract the amount of warm water used.

038

1. Funnel cloud.
2. Waterspout.
3. Tornado.
4. A funnel-shaped appendage hanging from the base of the cloud.
5. The phenomenon must be visible from the observation site.

6. When hail is falling or lightning is observed in the immediate vicinity of your station and the local noise level prevents you from hearing the thunder.
7. 15 minutes after the last occurrence of thunder, hail, or lightning.

039

1. (1) Fog.
   (2) Blowing snow.
   (3) Freezing fog.
   (4) Blowing spray.
   (5) Mist.
2. (1) Dust.
   (2) Sand.
   (3) Haze.
   (4) Smoke.
   (5) Volcanic ash.
3. (a) Fog.
   (b) Blowing spray.
   (c) Dust.
   (d) Blowing sand.
   (e) Haze.
   (f) Smoke.

040

1. Station pressure.
2. Sea-level pressure.
3. Altimeter setting.

041

1. Kinetic energy.
2. Dew point.
3. Difference between wet and dry bulb temperatures.
4. Actual vapor pressure of the air to the saturation vapor pressure.

042

1. Add easterly variations to the magnetic direction and subtract westerly variations from the magnetic direction.
2. A 2-minute period (10 minutes overseas). It's read to the nearest 10°.
3. A 2-minute period (10 minutes overseas) and is read to the nearest knot.
4. The Beaufort wind scale.
5. A sudden intermittent increase in windspeed with at least a 10-knot variation between peaks and lulls within the past 10 minutes.
6. A sudden increase of windspeed of at least 16 knots with a sustained average of 22 knots or more maintained for at least 1 minute before the speed diminishes.
7. A change in wind direction of 45° or more that takes place in less than 15 minutes.
8. A condition in which the wind direction is fluctuating by 60° or more during the observation.

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 Extension Course Program (A4L).

87. (035) What type of precipitation might you observe with clear skies?
    a. Snow.
    b. Ice crystals.
    c. Ice pellets.
    d. Snow grains.

88. (036) You would classify precipitation as intermittent if it
    a. stopped and started at least once within the preceding hour.
    b. stopped and started at least twice within the preceding hour.
    c. stopped and started at least once within the preceding 15 minutes.
    d. occurred any time in the preceding hour but not at the time of observation.

89. (037) When frozen precipitation is expected, you would
    a. collect it in the in the overflow unit of the rain gauge.
    b. collect it in the measuring tube of the rain gauge.
    c. use a 24-inch measuring stick to measure the frozen precipitation.
    d. use a 36-inch measuring stick to measure the frozen precipitation.

90. (038) The distinguishing feature of *any* tornadic activity is
    a. intense thunderstorm activity.
    b. wind gusts over 35 knots in speed.
    c. cumulonimbus mammatus clouds.
    d. the funnel-shaped appendage that hangs from the base of the cloud.

91. (038) A thunderstorm is present and occurring at your station. In addition, the local noise level is preventing you from hearing the thunderstorm. For observation purposes you would say
    a. hail is falling.
    b. heavy rain is falling.
    c. there is lightning within 10 nautical miles.
    d. a cumulonimbus cloud is present overhead.

92. (038) A thunderstorm *officially* ends
    a. ten minutes after the last occurrence of thunder, hail, or lightning.
    b. immediately after the last occurrence of thunder, hail, or lightning.
    c. fifteen minutes after the last occurrence of thunder, hail, or lightning.
    d. twenty minutes after the last occurrence of thunder, hail, or lightning.

93. (039) For observing purposes, five hydrometers are considered to be obstructions to vision. They include mist, fog, blowing snow,
    a. blowing spray, and blowing sand.
    b. freezing fog, and blowing sand.
    c. freezing fog, and blowing dust.
    d. freezing fog, and blowing spray.

94. (039) Blowing spray is reported *only* at sea stations near large bodies of water and when visibility at eye level is restricted to
    a. 9,000 meters or less.
    b. 7,000 meters or less.
    c. 6,000 meters or less.
    d. 3,200 meters or less.

95. (040) What pressure value is the *basis* for determining all other pressure values?
    a. Station pressure.
    b. Altimeter setting.
    c. Sea-level pressure.
    d. Corrected aneroid reading.

96. (040) What is the *reference level* for *all* pressure values?
    a. Station pressure.
    b. Altimeter setting.
    c. Sea-level pressure.
    d. Corrected aneroid reading.

97. (041) The temperature to which a given parcel of air must be cooled, with constant water vapor content and pressure, to reach saturation is called
    a. dew point.
    b. relative humidity.
    c. the dry-bulb temperature.
    d. the wet-bulb temperature.

98. (041) The temperature an air parcel would have if it were cooled adiabatically to saturation at constant pressure by evaporation of water into it is called
    a. dew point.
    b. relative humidity.
    c. dry bulb temperature.
    d. wet bulb temperature.

99. (042) Wind observing equipment is oriented to
    a. true north.
    b. true south.
    c. magnetic north.
    d. magnetic south.

100. (042) A change in wind direction of 45° (or more) that takes place in less than 15 minutes is called a
    a. gust.
    b. squall.
    c. wind shift.
    d. variable wind.

Glossary

Terms

Air Force Weather—All Air Force activities which function together in a system to produce worldwide weather service for the Air Force, Army, unified commands, national programs, and other military and government agencies. It includes base and post weather stations; staff functions; centralized weather, climatology production facilities; and communication systems serving AFW.

baroclinicity—When contours and isotherms are out-of-phase, advection is occurring and the atmosphere is described as baroclinic. With baroclinicity, the axis of the weather system tilts with height.

Chinook—Term applied to a foehn wind (downsloping) in North America. The wind is characterized as hot and dry.

frontal lift—Mechanical lifting of the air caused by a frontal boundary.

geostrophic windflow—The wind that would result if there were a balance between the coriolis force and the pressure gradient force over a region.

hydrostatic balance—The state of a fluid whose surfaces of constant pressure and constant mass (or density) coincide and are horizontal throughout.

inertia—A property of matter by which it remains at rest or in uniform motion.

inversion—In meteorology, the departure from the normal decrease of temperature with altitude. Instead of the temperature normally decreasing with height, it increases.

migratory—To move to another region.

nocturnal—Pertaining to, or occurring in, the night.

Nor'easter—Cyclonic storm occurring off the East Coast of North America. It is notorious for producing heavy rain, snow, and tremendous waves.

opaque—A term used in meteorology that denotes the quality of impervious to radiant energy (blocks the sun).

orographic—Of or relating to the mountains.

Santa Ana—Hot, dry wind generally from the east that funnel through the Santa Ana river valley in California.

transmissometer—A meteorological instrument that measures the ability of the air to transmit light.

hydrometeor—Any form of atmospheric water vapor, including those blown by the wind off the earth's surface. Liquid or solid water formation that is suspended in the air includes clouds, fog, ice fog, and mist.

lithometeor—Atmospheric phenomena which affect the state of the atmosphere. They constitute dry particles that hang suspended in the atmosphere, such as dust, smoke, sand, and haze.

transparency—A term used in meteorology that denotes the quality of being pervious to radiant energy (sunlight passes through).

Abbreviations and Acronyms

µm	micrometers
A	Arctic
AC	Altocumulus
ACC	Altocumulus castellanus
ACSL	Altocumulus standing lenticular (also SC, CC)
AFMAN	Air Force Manual
AFW	Air Force Weather
AGL	above ground level
ALSTG	altimeter setting
AMOS	automated meteorological observing system
AS	altostratus
BKN	broken
BLC	boundary layer convergence
C	Centigrade
CAA	cold-air advection
CB	cumulonimbus
CC	cirrocumulus
CDC	career development course
CeF	centrifugal force
CFC	cholorofluorocarbons
CGF	contour gradient force
CI	cirrus
CO_2	carbon dioxide
CoF	Coriolis force
CONUS	continental United States
CONV	convergence
cP	continental polar
CS	cirrostratus
cT	continental tropical
CU	cumulus
DIV	divergence
E	equatorial
F	Fahrenheit
FLIP	flight information publication
Fr	frictional force
ft	feet
Hg	inches of mercury
k	cooler
Km	kilometers

kt	knot(s)
LFC	level of free convection
λ	Greek symbol lambda
LND	level of nondivergence
m	meters
mb	millibar
METAR	meteorological aviation report
µm	micrometers
mP	maritime polar
MSL	mean sea level
mT	maritime tropical
NETWCZ	near-equatorial trade wind convergence zone
nm	nautical miles
NNE	intermediate directions
NS	nimbostratus
NVA	negative vorticity advection
OJT	on-the-job training
OVC	overcast
OWS	Operational Weather Squadron
PFJ	polar-front jet stream
PGF	pressure gradient force
PVA	positive vorticity advection
QTP	qualification training package
RH	relative humidity
RVR	runway visual range
RVRNO	RVR not available
SC	stratocumulus
SKT	scattered
SLP	sea-level pressure
sm	statute miles
ST	stratus
STJ	subtropical jet stream
TCU	towering cumulus
Td	dew point
V	direction of flow
$\mathbf{V_a}$	anticyclonic gradient circulation
$\mathbf{V_c}$	cyclonic gradient circulation
$\mathbf{V_{cyc}}$	cyclostrophic wind
VFR	visual flight rules
$\mathbf{V_g}$	gradient windflow
$\mathbf{V_{sub}}$	subgradient wind

V_{super}	supergradient wind
w	warmer
WAA	warm-air advection

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