CHAPTER 3

ATMOSPHERIC CIRCULATION

To understand large-scale motions of the atmosphere, it is essential that the Aerographer's Mate study the general circulation of the atmosphere. The sun's radiation is the energy that sets the atmosphere in motion, both horizontally and vertically. The rising and expanding of the air when it is warmed, or the descending and contracting of the air when it is cooled causes the vertical motion. The horizontal motion is caused by differences of atmospheric pressure; air moves from areas of high pressure toward areas of low pressure. Differences of temperature, the cause of the pressure differences, are due to the unequal absorption of the Sun's radiation by Earth's surface. The differences in the type of surface; the differential heating; the unequal distribution of land and water; the relative position of oceans to land, forests to mountains, lakes to surrounding land, and the like, cause different types of circulation of the air. Due to the relative position of Earth with respect to the Sun, much more radiation is absorbed near the equator than at other areas, with the least radiation being absorbed at or near the poles. Consequently, the principal factor affecting the atmosphere is incoming solar radiation, and its distribution depends on the latitude and the season.

GENERAL CIRCULATION

LEARNING OBJECTIVE: Recognize how temperature, pressure, winds, and the 3-cell theory affect the general circulation of Earth's atmosphere.

The general circulation theory attempts to explain the global circulation of the atmosphere with some minor exceptions. Since Earth heats unequally, the heat is carried away from the hot area to a cooler one as a result of the operation of physical laws. This global movement of air, which restores a balance of heat on Earth, is the general circulation.

WORLD TEMPERATURE GRADIENT

Temperature gradient is the rate of change of temperature with distance in any given direction at any point. World temperature gradient refers to the change in temperature that exists in the atmosphere from the equator to the poles. The change in temperature or temperature differential, which causes atmospheric circulation can be compared to the temperature differences produced in a pan of water placed over a gas burner. As the water is heated, it expands and its density is lowered. This reduction in density causes the warmer, less dense water to rise to the top of the pan. As it rises, it cools and is forced to the edges of the pan. Here it cools further and then sinks to the bottom, eventually working its way back to the center of the pan where it started. This process sets up a simple circulation pattern due to successive heating and cooling.

Ideally, the air within the troposphere may be compared to the water in the pan. The most direct rays of the Sun hit Earth near the equator and cause a net gain of heat. The air at the equator heats, rises, and flows in the upper atmosphere toward both poles. Upon reaching the poles, it cools sufficiently and sinks back toward Earth, where it tends to flow along the surface of Earth back to the equator. (See fig. 3-1).

Simple circulation of the atmosphere would occur as described above if it were not for the following factors:

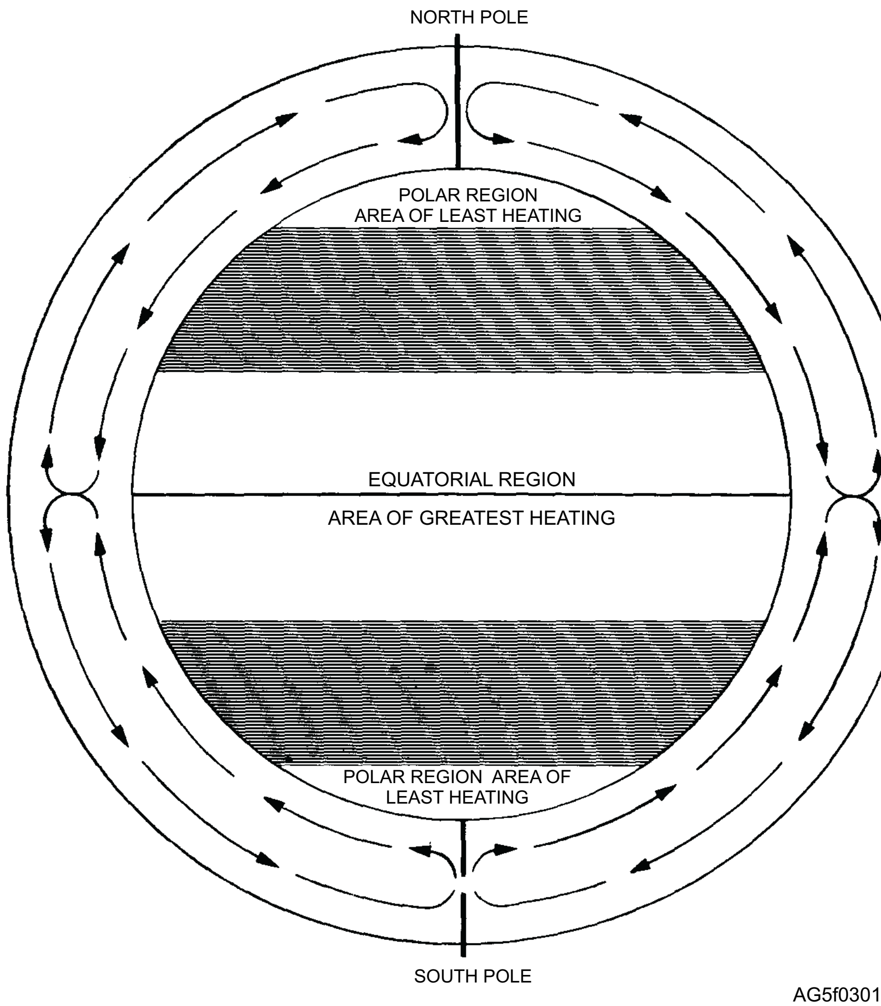

Figure 3-1.—Simple circulation.

1. Earth rotates, resulting in an apparent force known as the Coriolis force (a deflecting force). This rotation results in a constant change to the area being heated.

2. Earth is covered by irregular land and water surfaces that heat at different rates.

Regions under the direct rays of the Sun absorb more heat per unit time than those areas receiving oblique rays. The heat produced by the slanting rays of the Sun during early morning may be compared with the heat that is produced by the slanting rays of the Sun during winter. The heat produced by the more direct rays at midday can be compared with the heat resulting from the more direct rays of summer. The length of day, like the angle of the Sun's rays, influences the temperature. The length of day varies with the latitude and the season. Near the equator there are about 12 hours of daylight with the Sun's rays striking the surface more directly. Consequently, equatorial regions normally do not have pronounced seasonal temperature variations.

During the summer in the Northern Hemisphere, all areas north of the equator have more than 12 hours of daylight. During the winter the situation is reversed; latitudes north of the equator have less than 12 hours of daylight. Large seasonal variation in the length of the day and the seasonal difference in the angle at which the Sun's rays reach Earth's surface cause seasonal temperature differences in middle and high latitudes. The weak temperature gradient in the subtropical areas and the steeper gradient poleward can be seen in figures 3-2A and 3-2B. Note also how much steeper the gradient is poleward in the winter season of each hemisphere as compared to the summer season.

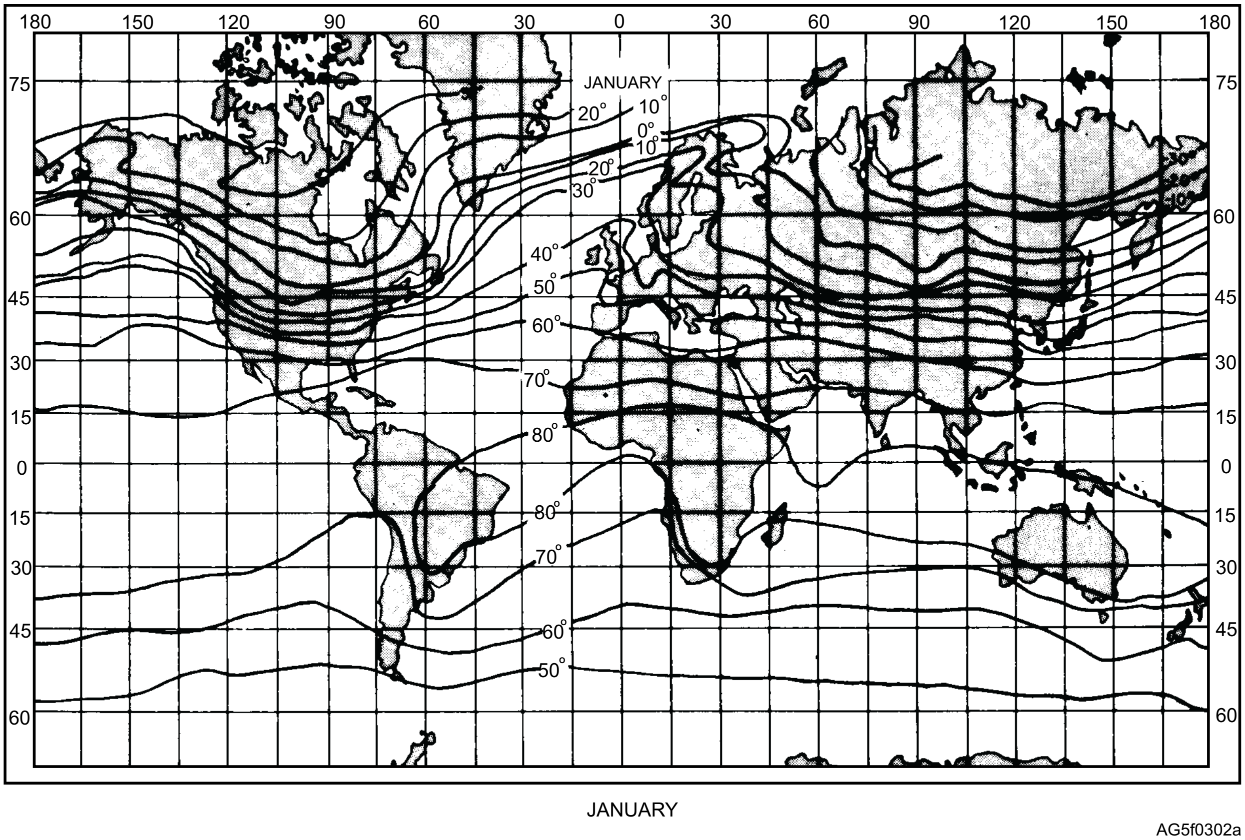

Figure 3-2A.—Mean world temperature for January.

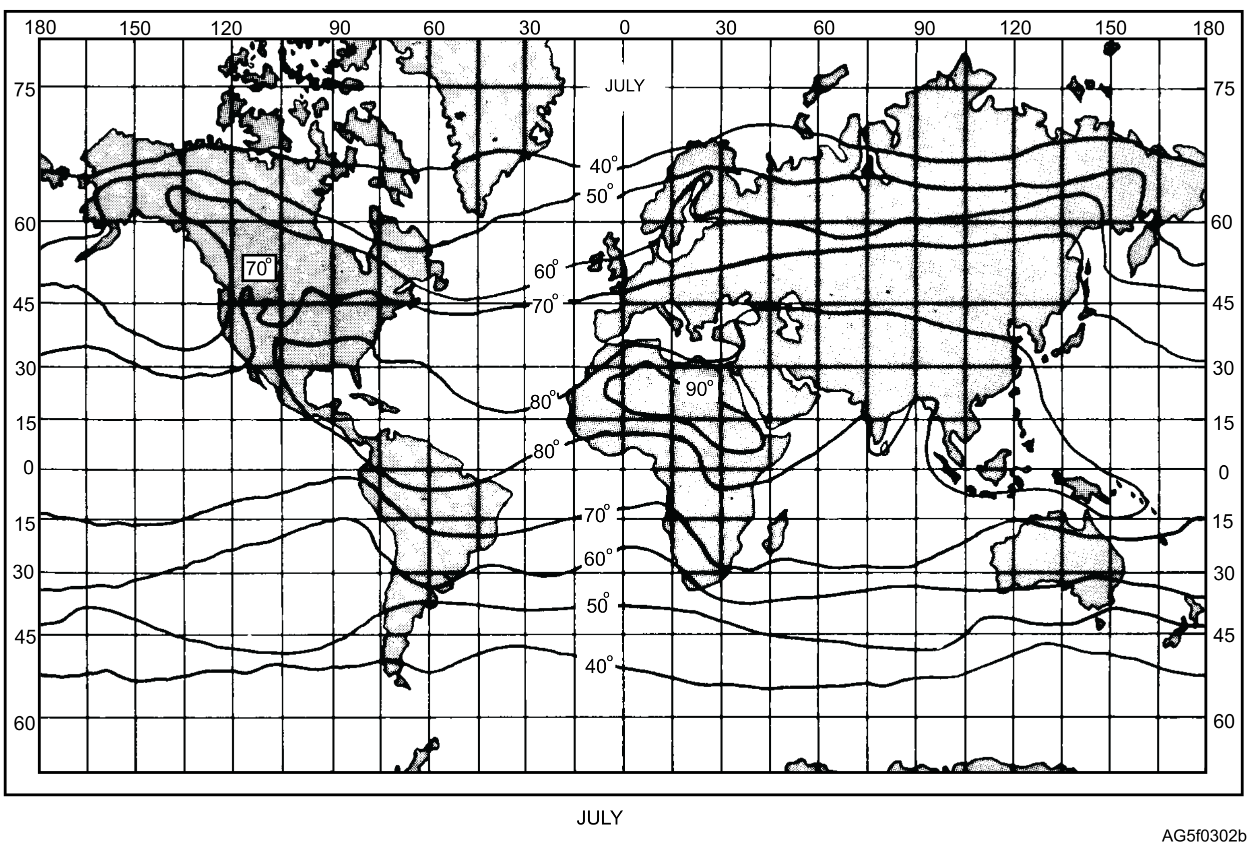

Figure 3-2B.—Mean world temperature for July.

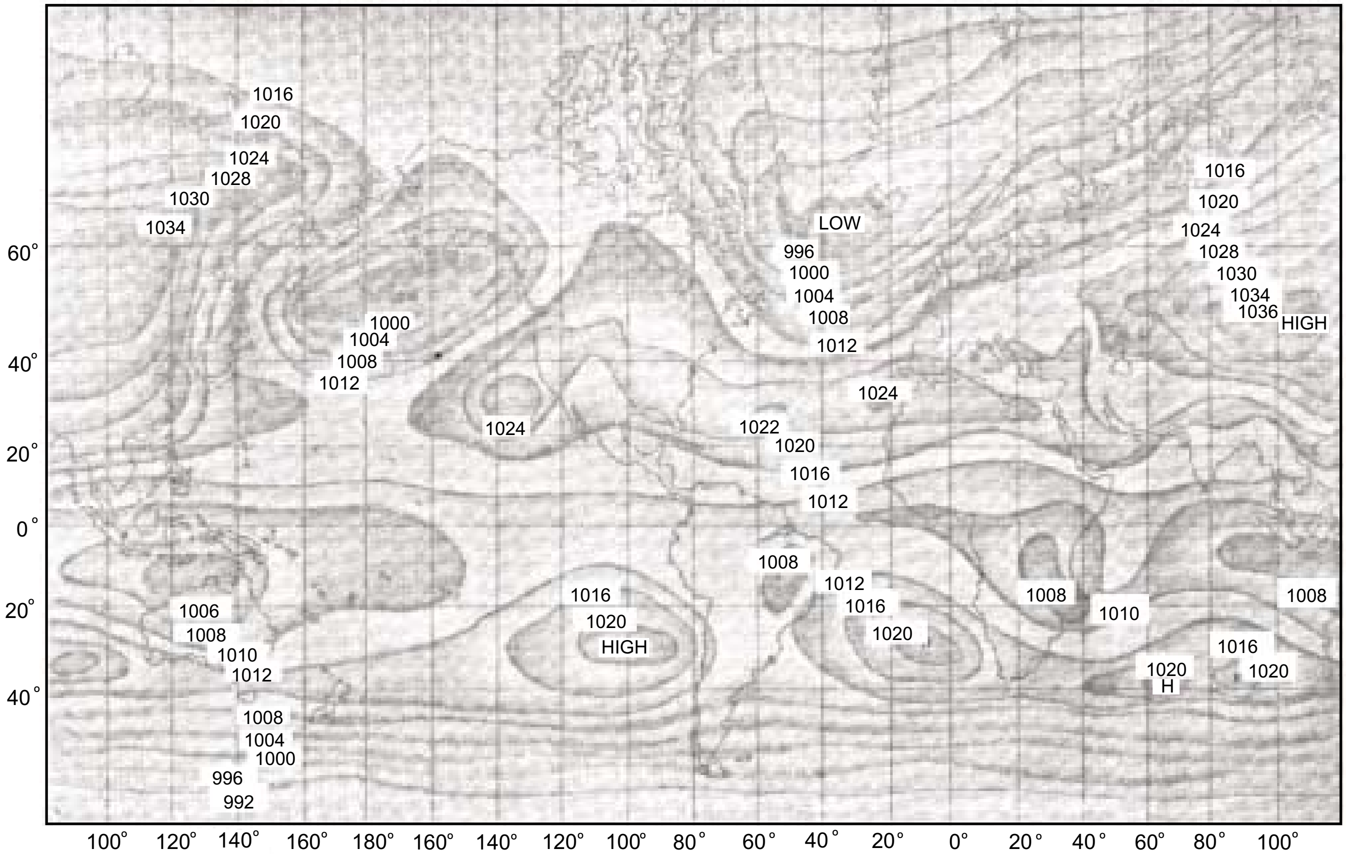

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Figure 3-3A.—Mean world pressure for January.

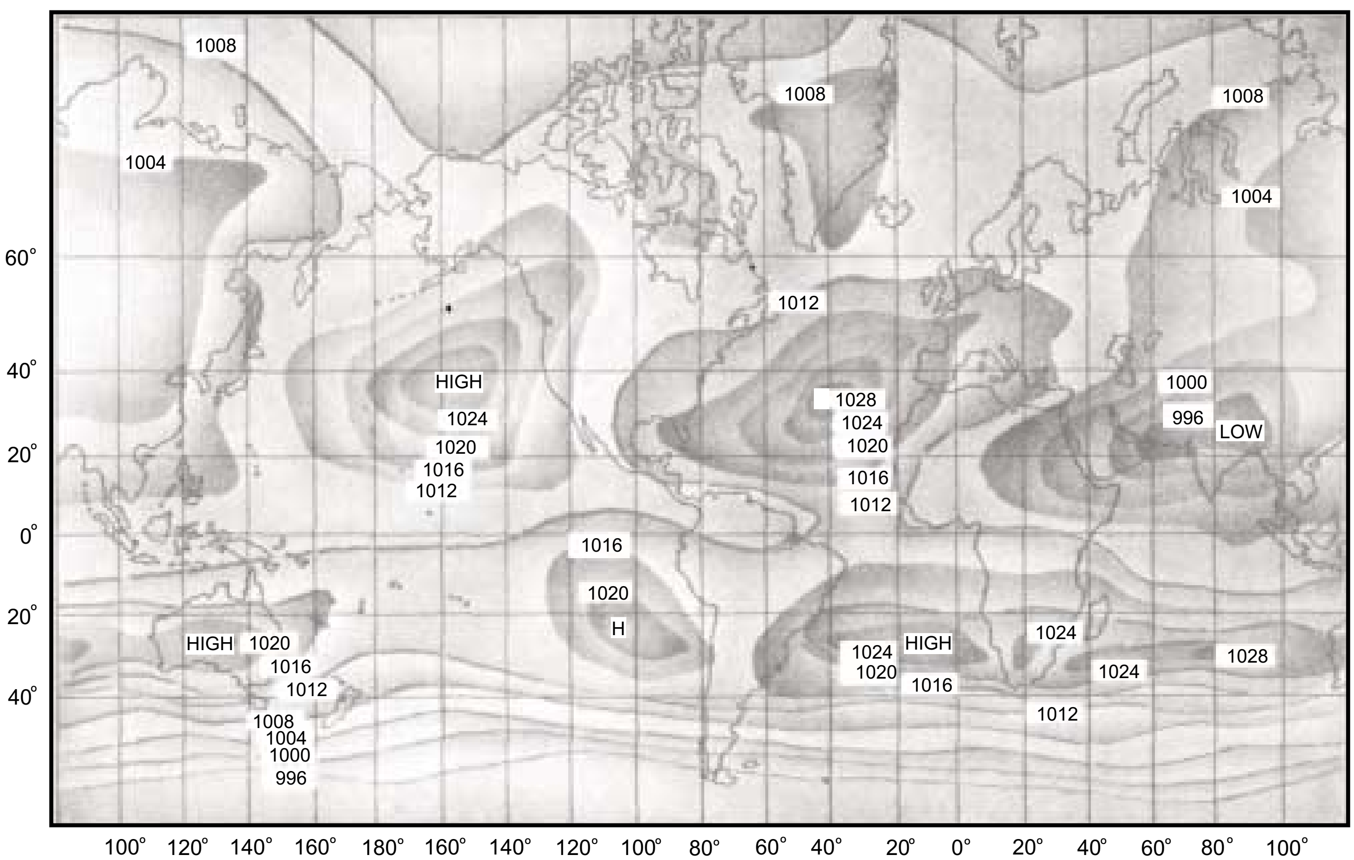

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Figure 3-3B.—Mean world pressure for July.

PRESSURE OVER THE GLOBE

The unequal heating of Earth's surface due to its tilt, rotation, and differential insolation, results in the wide distribution of pressure over Earth's surface. Study figures 3-3A and 3-3B. Note that a low-pressure area lies along the intertropical convergence zone (ITCZ) in the equatorial region. This is due to the higher temperatures maintained throughout the year in this region. At the poles, permanent high-pressure areas remain near the surface because of the low temperatures in this area throughout the entire year. Mainly the "piling up" of air in these regions causes the subtropical high-pressure areas at 30°N and S latitudes. Relatively high or low pressures also dominate other areas during certain seasons of the year.

ELEMENTS OF CIRCULATION

Temperature differences cause pressure differences, which in turn cause air movements. The following sections show how air movements work and how they evolve into the various circulations—primary, secondary, and tertiary.

To explain the observed wind circulation over Earth, three basic steps are used. The first step is to assume Earth does not rotate and is of uniform surface; that is, all land or all water. The second step is to rotate Earth, but still assume a uniform surface. The third step is to rotate Earth and assume a non-uniform surface. For now, we deal with the first two steps, a non-rotating Earth of uniform surface and a rotating Earth of uniform surface.

Static Earth

The circulation on a non-rotating Earth is referred to as the thermal circulation because it is caused by the difference in heating. The air over the equator is heated and rises (low pressure); while over the poles the air is cooled and sinks (high pressure). This simple circulation was shown in figure 3-1.

Rotating Earth

In thermal circulation, the assumption was made that the Earth did not rotate, but of course this is not true. The rotation of Earth causes a force that affects thermal circulation. This rotation results in the deflection to the *right* of movement in the Northern Hemisphere, and to the *left* of the movement in the Southern Hemisphere. This force is called the *Coriolis* force. The Coriolis force is not a true force. It is an apparent force resulting from the west-to-east rotation of Earth. The effects, however, are real.

Arctic rivers cut faster into their right banks than their left ones. On railroads carrying only one-way traffic, the right hand rails wear out faster than the left-hand rails. Artillery projectiles must be aimed to the left of target because they deflect to the right. Pendulum clocks run faster in high latitudes than in lower latitudes. All these phenomena are the result of the Coriolis force, which is only an apparent force. The most important phenomena are that this force also deflects winds to the right in the Northern Hemisphere. Therefore, it is important to understand how this force is produced.

As Earth rotates, points on the surface are moving eastward (from west to east) past a fixed point in space at a given speed. Points on the equator are moving at approximately 1,000 miles per hour, points on the poles are not moving at all, but are merely pivoting, the points somewhere between are moving at speeds between 1,000 and zero miles per hour depending upon their relative position. Refer to view A in figure 3-4.

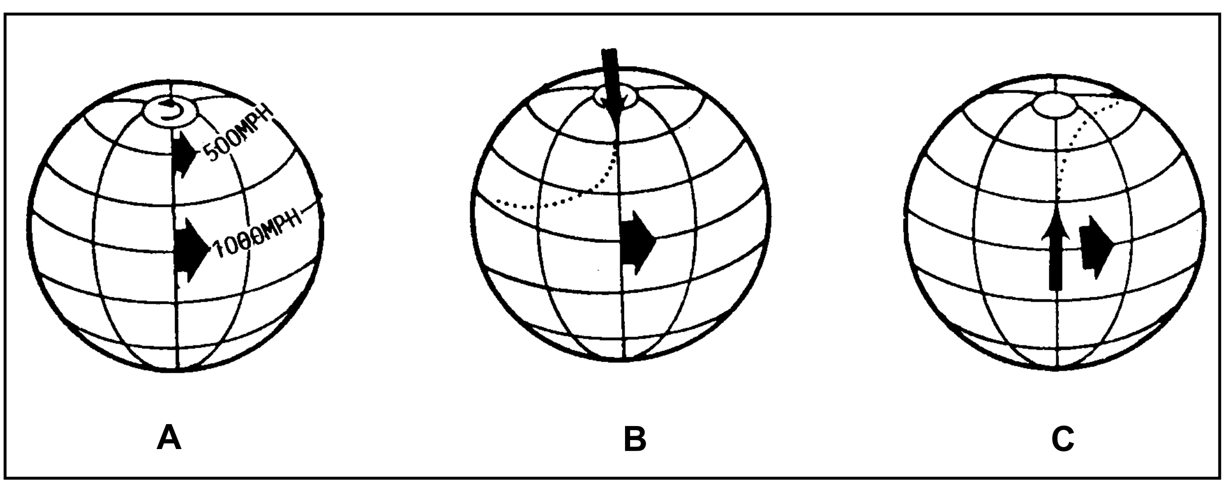

Figure 3-4.—Coriolis force.

Assume that a missile located at the North Pole is launched at a target on the equator. The missile does not have any eastward lateral velocity, but the target has an eastward velocity of 1,000 miles per hour. The result is that the missile appears to be deflected to the right as the target moves away from its initial position. Refer to view B in figure 3-4.

A similar condition assumes that a missile located on the equator is launched at a target at the North Pole. The missile has an eastward lateral velocity of 1,000 miles per hour, while the target on the pole has no lateral velocity at all. Once again the missile appears to be deflected to the right as a result of its initial eastward lateral velocity. Refer to view C in figure 3-4.

Due to Earth's rotation and the Coriolis effect, the simple circulation now becomes more complex as shown in figure 3-5. The complex on resulting from the interplay of the Coriolis effect with the flow of air is known as the theory. (See fig. 3-6.)

3-CELL THEORY

According to the 3-cell theory, Earth is divided into six circulation belts—three in the Northern Hemisphere and three in the Southern Hemisphere. The dividing lines are the equator, latitude, and 60°N and S latitude. The general circulation of the Northern Hemisphere is similar to those of the Southern Hemisphere. (Refer to fig. 3-6 during the following discussion.)

First, note the tropical cell of the Northern Hemisphere that lies between the equator and 30°N latitude. Convection at the equator causes the air to heat

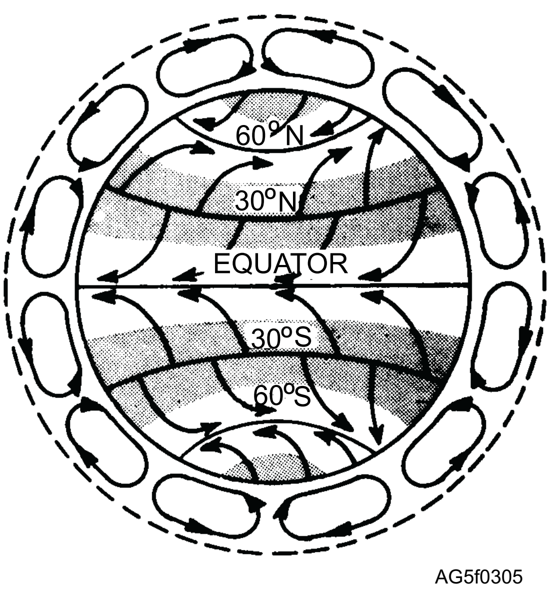

Figure 3-5.—Coriolis effect on windflow.

and rise, due to convection. When it reaches the upper portions of the troposphere, it tends to flow toward the North Pole. By the time the air has reached 30°N latitude, the Coriolis effect has deflected it so much that it is moving eastward instead of northward. This results in a piling up of air (convergence) near 30°N latitude and a descending current of air (subsidence) toward the surface which forms a belt of high pressure. When the descending air reaches the surface where it flows outward (divergence), part of it flows poleward to become part of the mid-latitude cell; the other part flows toward the equator, where it is deflected by the Coriolis effect and forms the northeast trades.

The mid-latitude cell is located between 30° and 60°N latitude. The air, which comprises this cell, circulates poleward at the surface and equatorward aloft with rising currents at 60° (polar front) and descending currents at 300 (high-pressure belt). However, in general, winds both at the surface and aloft blow from the west. The Coriolis effect easily explains this for the surface wind on the poleward-moving surface air. The west wind aloft is not as easily explained. Most authorities agree that this wind is frictionally driven by the west winds in the two adjacent cells.

The polar cell lies between 60°N latitude and the North Pole. The circulation in this cell begins with a flow of air at a high altitude toward the pole. This flow cools and descends at the North Pole and forms a high-pressure area in the Polar Regions. After reaching the surface of Earth, this air usually flows equatorward and is deflected by the Coriolis effect so that it moves from the northeast. This air converges with the poleward flow from the mid-latitude cell and is deflected upward with a portion circulating poleward again and the remainder equatorward. The outflow of air aloft between the polar and mid-latitude cells causes a semi-permanent low-pressure area at approximately 60°N latitude. To complete the picture of the world's general atmospheric circulation, we must associate this prevailing wind and pressure belts with some basic characteristics.

WORLD WINDS

In the vicinity of the equator is a belt of light and variable winds known as the doldrums. On the poleward side of the doldrums are the *trade winds*; the predominant wind system of the tropics. These easterly winds are the most consistent on Earth, especially over the oceans. Near 30°N and 30°S latitudes lie the sub-tropical high-pressure belts. Winds are light and

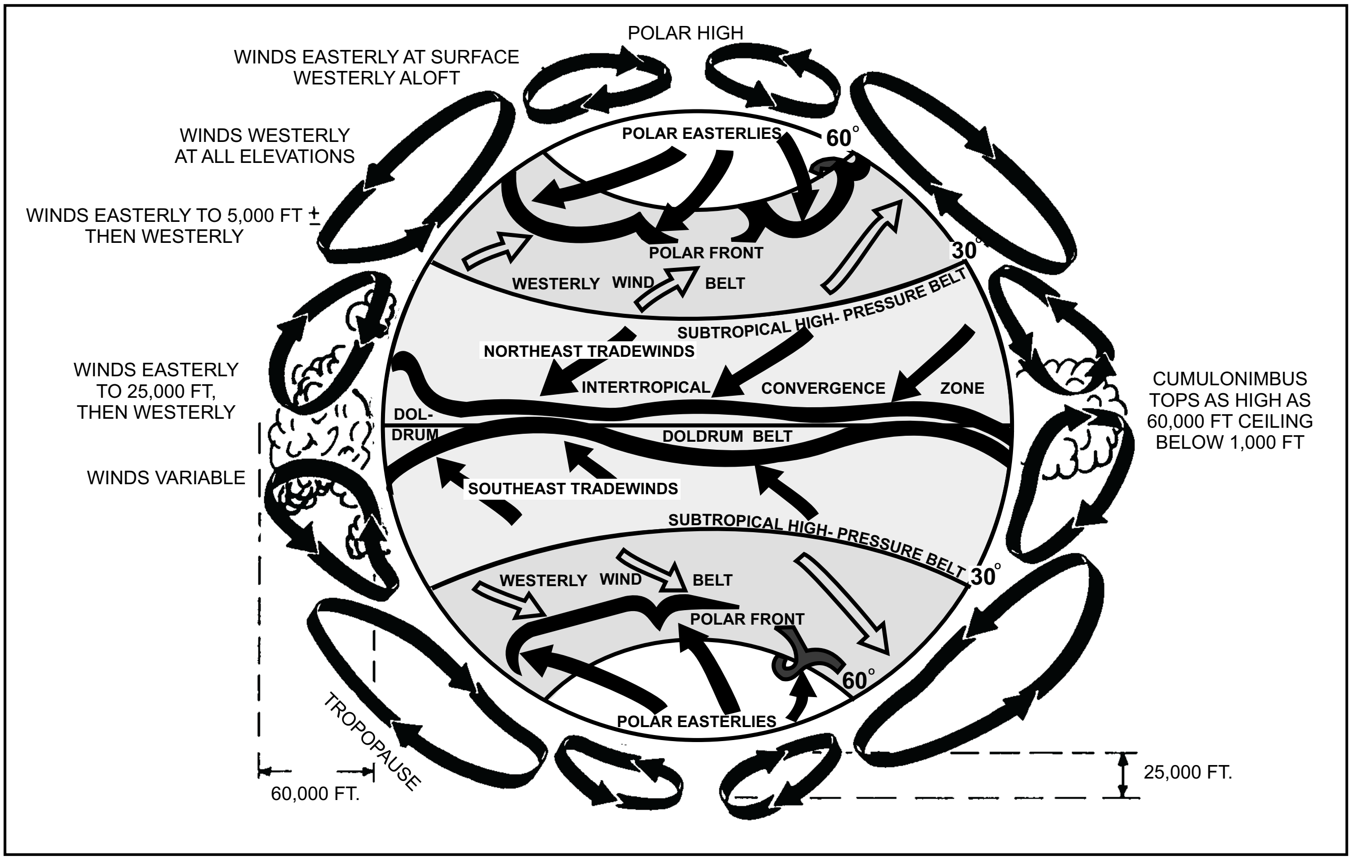

Figure 3-6.—Idealized pattern of the general circulation. (The 3-cell theory).

variable. These areas are referred to as the *horse latitudes*. The prevailing westerlies, which are on the poleward side of the subtropical high-pressure belt, are persistent throughout the mid-latitudes. In the Northern Hemisphere, the direction of the westerlies at the surface is from the southwest. In the Southern Hemisphere, westerlies are from the northwest. This is due to the deflection area resulting from the Coriolis effect as the air moves poleward.

Poleward of the prevailing westerlies, near 60°N and 60°S latitudes, lies the belt of low-pressure basic pressure known as the *polar front zone*. Here, converging winds result in ascending air currents and consequent poor weather.

WIND THEORY

Newton's first two laws of motion indicate that motion tends to be in straight lines and only deviates from such lines when acted upon by another force or by a combination of forces. Air tends to move in a straight line from a high-pressure area to a low-pressure area. However, there are forces that prevent the air from moving in a straight line.

Wind Forces

There are four basic forces that affect the directional movement of air in our atmosphere: pressure gradient force (PGF), the Coriolis effect, centrifugal force, and frictional force. These forces, working together, affect air movement. The forces that are affecting it at that particular time determine the direction that the air moves. Also, the different names given to the movement of the air (geostrophic wind, gradient wind, etc.) depends on what forces are affecting it.

Pressure Gradient

The rate of change in pressure in a direction perpendicular to the isobars is called *pressure gradient*. Pressure applied to a fluid is exerted equally in all

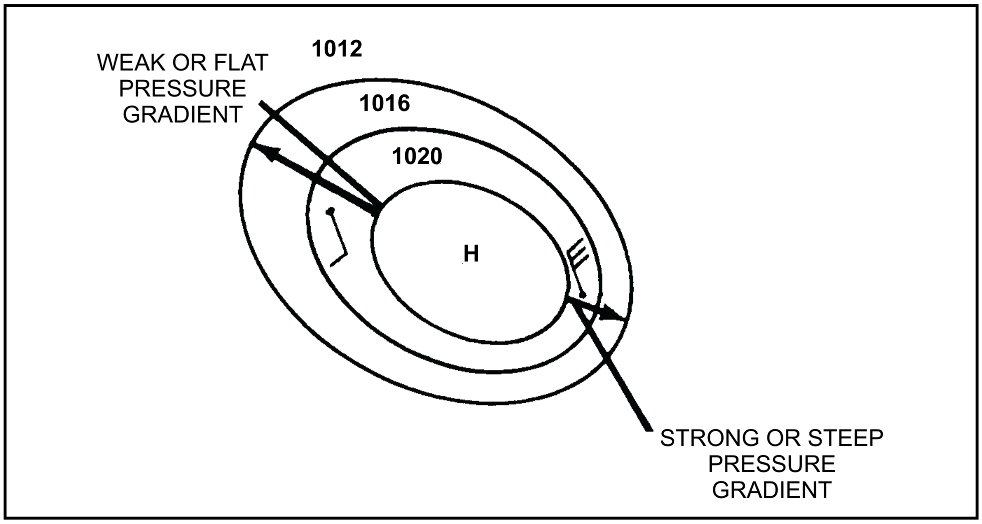

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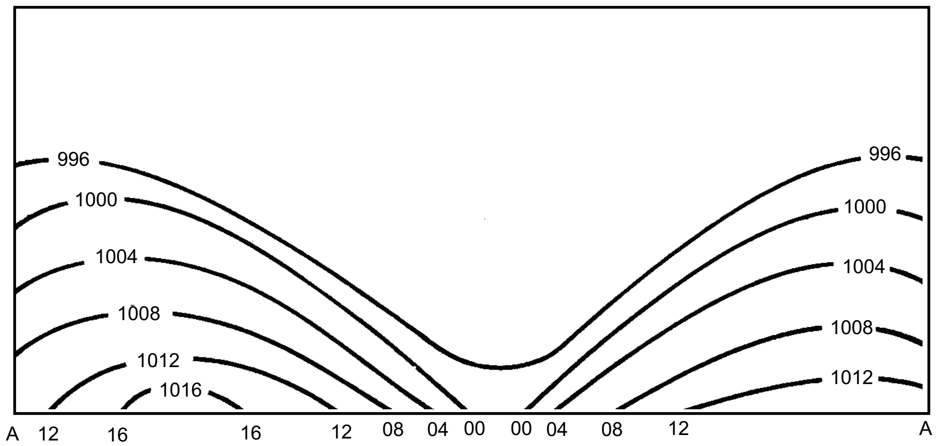
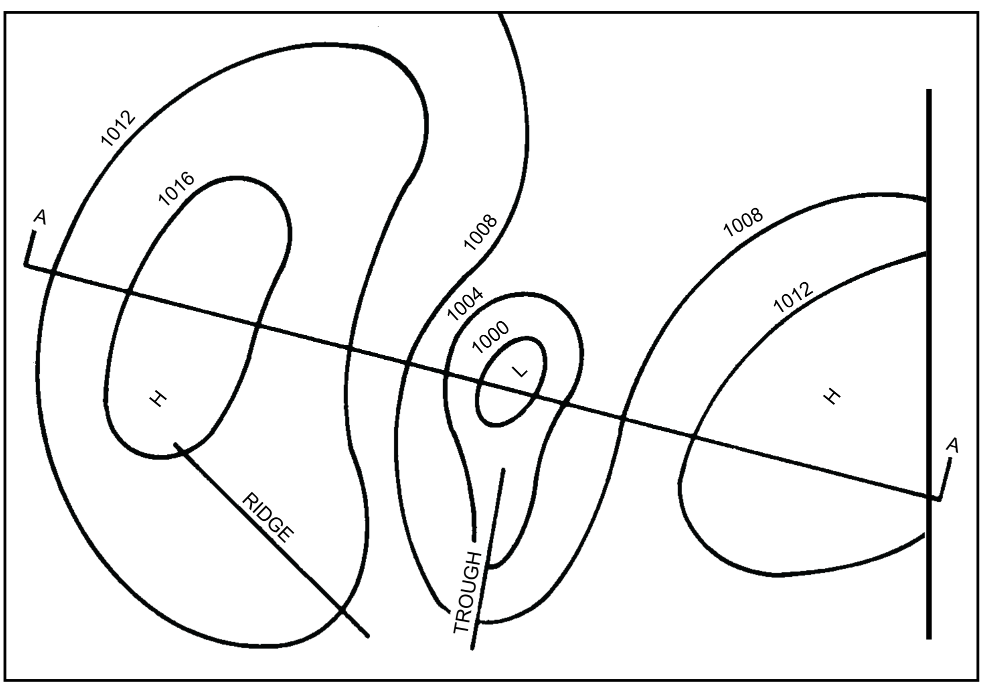
Figure 3-7.—Horizontal pressure gradient.



A. CROSS SECTION OF A VERTICAL PRESSURE GRADIENT ALONG LINE AA



B. HORIZONTAL PRESSURE GRADIENT

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Figure 3-8.—Cross section of a vertical pressure gradient along line AA.

directions throughout the fluid; e.g., if a pressure of 1013.2 millibars is exerted downward by the atmosphere at the surface, this same pressure is also exerted horizontally outward at the surface. Therefore, a pressure gradient exists in the horizontal (along the surface) as well as the vertical plane (with altitude) in the atmosphere.

HORIZONTAL PRESSURE GRADIENT.—The horizontal pressure gradient is *steep* or *strong* when the isobars determining the pressure system (fig. 3-7) are close together. It is *flat* or *weak* when the isobars are far apart.

VERTICAL PRESSURE GRADIENT.—If isobars are considered as depicting atmospheric topography, a high-pressure system represents a hill of air, and a low-pressure system represents a depression or valley of air. The vertical pressure gradient always indicates a decrease in pressure with altitude, but the rate of pressure decrease (gradient) varies directly with changes in air density with altitude. Below 10,000 feet altitude, pressure decreases approximately 1 inch of mercury per 1,000 feet in the standard atmosphere. The vertical cross section through a high and low (view A in fig. 3-8) depicts the vertical pressure gradient. A surface weather map view of the horizontal pressure gradient in the same high and low is illustrated in view B of the figure 3-8.

Pressure Gradient Force

The variation of heating (and consequently the variations of pressure) from one locality to another is the initial factor that produces movement of air or wind. The most direct path from high to low pressure is the path along which the pressure is changing most rapidly. The rate of change is called the pressure gradient. Pressure gradient force is the force that moves air from an area of high pressure to an area of low pressure. The velocity of the wind depends upon the pressure gradient. If the pressure gradient is strong, the wind speed is high. If the pressure gradient is weak, the wind speed is light. (See fig. 3-7.)

Figure 3-9 shows that the flow of air is from the area of high pressure to the area of low pressure, but it does not flow straight across the isobars. Instead the flow is circular around the pressure systems. Pressure gradient force (PGF) causes the air to begin moving from the high-pressure to the low-pressure system. Coriolis (deflective) force and centrifugal force then begin acting on the flow in varying degrees. In this example, frictional force is not a factor.

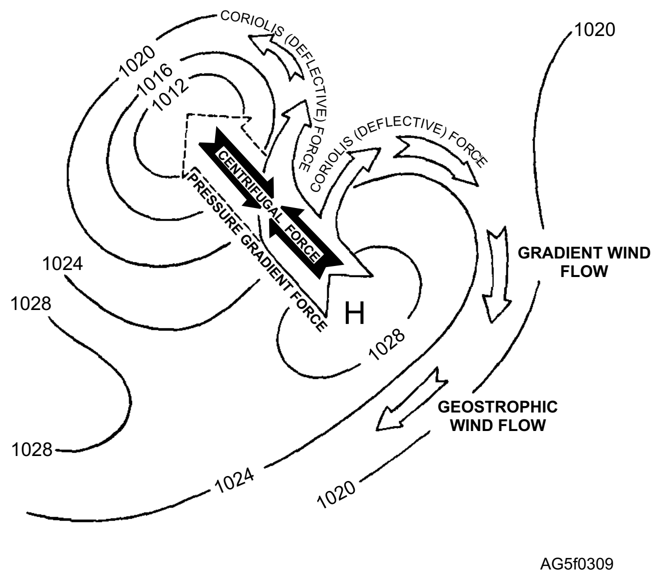

Figure 3-9.—Examples of circulation around high and low pressure systems.

Coriolis Effect

If pressure gradient force were the only force affecting windflow, the wind would blow at right angles across isobars (lines connecting points of equal barometric pressure) from high to low pressure. The wind actually blows parallel to isobars above any frictional level. Therefore, other factors must be affecting the windflow; one of these factors is the rotation of Earth. A particle at rest on Earth's surface is in equilibrium. If the particle starts to move because of a pressure gradient force, its relative motion is affected by the rotation of Earth. If a mass of air from the equator moves northward, it is deflected to the right, so that a south wind tends to become a southwesterly wind.

In the Northern Hemisphere, the result of the Coriolis effect is that moving air is deflected to the right of its path of motion. This deflection to the right is directly proportional to the speed of the wind; the faster the wind speed, the greater the deflection to the right, and conversely, the slower the wind speed, the less the deflection to the right. Finally, this apparent deflective force is stronger at the Polar Regions than at the equator.

Centrifugal Force

According to Newton's first law of motion, a body in motion continues in the same direction in a straight line and with the same speed unless acted upon by some external force. Therefore, for a body to move in a curved path, some force must be continually applied. The force restraining bodies that move in a curved path is called the *centripetal force*; it is always directed *toward* the center of rotation. When a rock is whirled around on a string, the centripetal force is afforded by the tension of the string.

Newton's third law states that for every action there is an equal and opposite reaction. *Centrifugal force* is the reacting force that is equal to and opposite in direction to the centripetal force. Centrifugal force, then, is a force directed *outward* from the center of rotation.

As you know, a bucket of water can be swung over your head at a rate of speed that allows the water to remain in the bucket. This is an example of both centrifugal and centripetal force. The water is held in the bucket by centrifugal force tending to pull it outward. The centripetal force, the force holding the bucket and water to the center, is your arm swinging the bucket. As soon as you cease swinging the bucket, the forces cease and the water falls out of the bucket. Figure 3-10 is a simplified illustration of centripetal and centrifugal force.

High- and low-pressure systems can be compared to rotating discs. Centrifugal effect tends to fling air out from the center of rotation of these systems. This force is directly proportional to the wind speeds, the faster the wind, and the stronger the outward force. Therefore, when winds tend to blow in a circular path, centrifugal effect (in addition to pressure gradient and Coriolis effects) influences these winds.

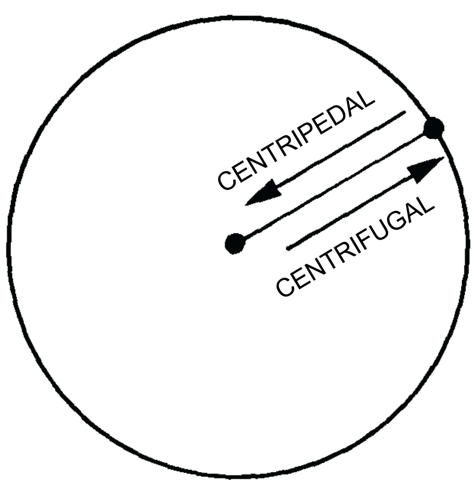

Figure 3-10.—Simplified illustration of centripetal and centrifugal force.

Frictional Force

The actual drag or slowing of air particles in contact with a solid surface is called friction. Friction tends to retard air movement. Since Coriolis force varies with the speed of the wind, a reduction in the wind speed by friction means a reduction of the Coriolis force. This results in a momentary disruption of the balance. When the new balance (including friction) is reached, the air flows at an angle across the isobars from high pressure to low pressure. (Pressure gradient force is the dominant force at the surface.) This angle varies from 10 degrees over the ocean to more than 45 degrees over rugged terrain. Frictional effects on the air are greatest near the ground, but the effects are also carried aloft by turbulence. Surface friction is effective in slowing the wind to an average altitude of 2,000 feet (about 600 meters) above the ground. Above this level, called the gradient wind level or the second standard level the effect of friction decreases rapidly and may be considered negligible. Air above 2,000 feet normally flows parallel to the isobars.

WIND TYPES

Since there is a direct relationship between pressure gradient and wind speed and direction, we have a variety of wind types to deal with. We discuss below the relationship of winds and circulations, the forces involved, and the effect of these factors on the general circulation.

Geostrophic and Gradient Wind

On analyzed surface weather charts, points of equal pressure are connected by drawn lines referred to as isobars, while in upper air analysis, points of equal heights are connected and called isoheights.

The variation of these heights and pressures from one locality to another is the initial factor that produces movement of air, or wind. Assume that at three stations the pressure is lower at each successive point. This means that there is a horizontal pressure gradient (a decrease in pressure in this case) for each unit distance. With this situation, the air moves from the area of greater pressure to the area of lesser pressure.

If the force of the pressure were the only factor acting on the wind, the wind would flow from high to low pressure, perpendicular to the isobars. Since experience shows the wind does not flow perpendicular to isobars, but at a slight angle across them and towards the lower pressure, it is evident that other factors are involved. These other factors are the Coriolis effect, frictional force, and centrifugal effect. When a unit of air moves with no frictional force involved, the movement of air is parallel to the isobars. This wind is called a gradient wind. When the isobars are straight, so only Coriolis and pressure gradient forces are involved, it is termed a geostrophic wind.

Let's consider a parcel of air from the time it begins to move until it develops into a geostrophic wind. As soon as a parcel of air starts to move due to the pressure gradient force, the Coriolis force begins to deflect it from the direction of the pressure gradient force. (See views A and B of fig. 3-11). The Coriolis force is the apparent force exerted upon the parcel of air due to the rotation of Earth. This force acts to the right of the path of motion of the air parcel in the Northern Hemisphere (to the left in the Southern Hemisphere). It always acts at right angles to the direction of motion. In the absence of friction, the Coriolis force changes the direction of motion of the parcel until the Coriolis force and the pressure gradient force are in balance. When the two forces are equal and opposite, the wind blows parallel to the straight isobars (view C in fig. 3-11). The Coriolis force only affects the direction, not the speed of the motion of the air. Normally, Coriolis force is not greater than the pressure gradient force. In the case of super-gradient winds, Coriolis force may be greater than the pressure gradient force. This causes the wind to deflect *more* to the right in the Northern Hemisphere, or toward higher pressure.

Under actual conditions, air moves around high and low pressure centers toward lower pressure. Turn back to figure 3-9. Here, the flow of air is from the area of high pressure to the area of low pressure, but, as we mentioned previously, it does not flow straight across the isobars (or isoheights). Instead, the flow is circular around the pressure systems.

The Coriolis force commences deflecting the path of movement to the right (Northern Hemisphere) or left (Southern Hemisphere) until it reaches a point where a balance exists between the Coriolis and the pressure gradient force. At this point the air is no longer deflected and moves forward around the systems.

Once circular motion around the systems is established, then centrifugal force must be considered. Centrifugal force acts outward from the center of both the highs and the lows with a force dependent upon the velocity of the wind and the degree of curvature of the isobars. However, the pressure gradient force is acting towards the low; therefore, the flow in that direction persists. When the flow is parallel to the curved portion of the analysis in figure 3-9, it is a *gradient wind*. When it is moving parallel to that portion of the analysis showing straight flow, it is a *geostrophic wind*.

We defined pressure gradient as being a change of pressure with distance. This means that if the isobars are closely spaced, then the pressure change is greater over a given distance; it is smaller if they are widely spaced. Therefore, the closer the isobars, the faster the flow. Geostrophic and gradient winds are also dependent, to a certain extent, upon the density of the atmosphere and the latitude. If the density and the pressure gradient remain constant and the latitude increases, the wind speed decreases. If the latitude

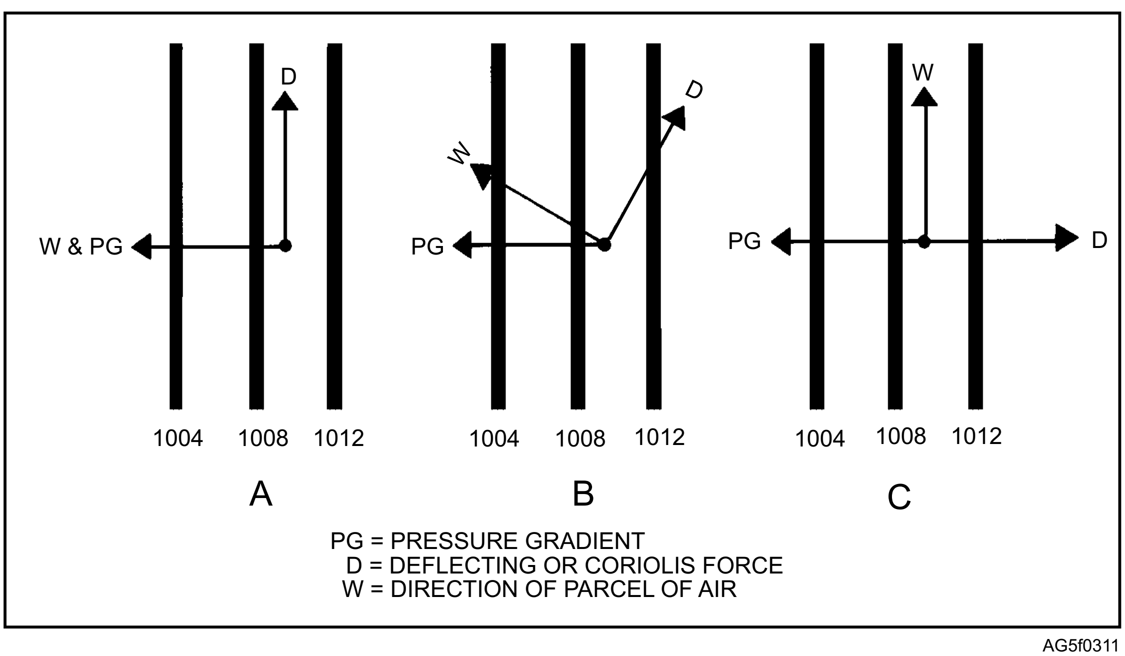

Figure 3-11.—Development cycle of a geostrophic wind.

decreases, the wind speed increases. If the density and the latitude remain constant and the pressure gradient decreases, the wind speed decreases. If the pressure gradient and the latitude remain constant and the density decreases, the wind speed increases. If the density increases, the wind speed decreases. True geostrophic wind is seldom observed in nature, but the conditions are closely approximated on upper-level charts.

Cyclostrophic Wind

In some atmospheric conditions, the radius of rotation becomes so small that the centrifugal force becomes quite strong in comparison with the Coriolis force. This is particularly true in low latitudes where the Coriolis force is quite small to begin with. In this case, the pressure gradient force is nearly balanced by the centrifugal force alone. When this occurs, the wind is said to be cyclostrophic. By definition, a cyclostrophic wind exists when the pressure gradient force is balanced by the centrifugal force alone.

This exact situation rarely exists, but is so nearly reached in some situations that the small Coriolis effect is neglected and the flow is said to be cyclostrophic. Winds in a hurricane or typhoon and the winds around a tornado are considered cyclostrophic.

Movement of Wind around Anticyclones

The movement of gradient winds around anticyclones is affected in a certain manner by the pressure gradient force, the centrifugal force, and the Coriolis force. The pressure gradient force acts from high to low pressure, and the Coriolis force acts opposite to the pressure gradient force and at right angles to the direction of movement of the parcel of air. The centrifugal force acts at right angles to the path of motion and outward from the center about which the parcel is moving. (See fig. 3-12.) In the case of a high-pressure center, the pressure gradient force and the centrifugal force balance the Coriolis force. This phenomenon may be expressed in the following manner:

$$PG + CF = D$$

Movement of Wind around Cyclones

As in the case of anticyclones, the pressure gradient force, the centrifugal force, and the Coriolis force affect gradient winds around cyclones, but the balance of the forces is different. (See fig. 3-12.) In a cyclonic situation the Coriolis force and the centrifugal force balance the pressure gradient force. This balance may be expressed in the following manner:

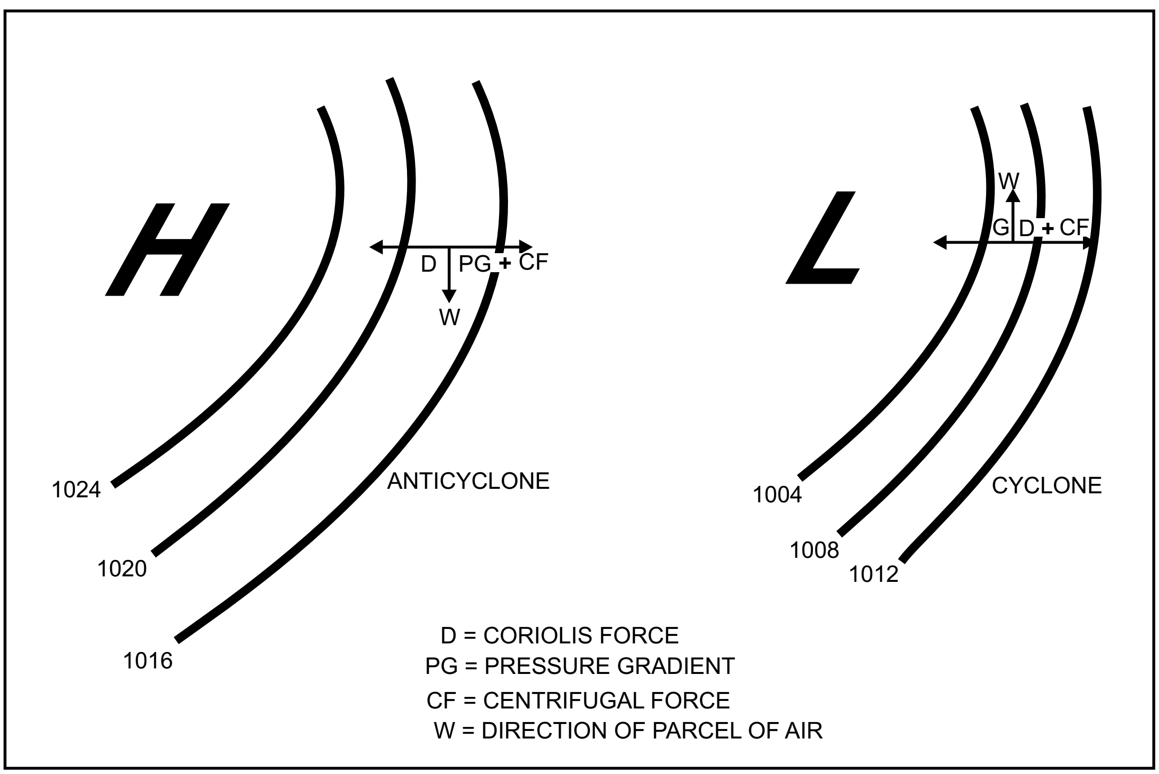

Figure 3-12.—Forces acting on pressure systems.

PG = D+ CF

Centrifugal force acts *with* the pressure gradient force when the circulation is anticyclonic and *against* the pressure gradient force when the circulation is cyclonic. Therefore, wind velocity is greater in an anticyclone than in a cyclone of the same isobaric spacing.

Variations

It has been determined that, given the same density, pressure gradient, and latitude, the wind is weaker around a low-pressure cell than a high-pressure cell. This is also true for gradient and geostrophic winds. The wind we observe on a synoptic chart is usually stronger around low cells than high cells because the pressure gradient is usually stronger around the low-pressure cell.

Geostrophic and Gradient Wind Scales

The geostrophic wind is stronger than the gradient wind around a low and is weaker than a gradient wind around a high. This is why the isobar spacing and contour spacing, for a curved flow, differs from that determined by a geostrophic wind scale. If the flow under consideration is around a high-pressure cell, the isobars are farther apart than indicated by the geostrophic wind scale. If the flow is around a low-pressure cell, the isobars are closer together than indicated by the geostrophic wind scale.

Geostrophic and gradient wind scales are used to determine the magnitude of these winds (based on isobar or contour spacing) and to determine the isobar or contour spacing (based on observed wind speeds). There are a number of scales available for measuring geostrophic and gradient flow of both surface and upper air charts.

Weather plotting charts used by the Naval Oceanography Command has geostrophic wind scales printed on them for both isobaric and contour spacing. The most common scales in general use can be used for both surface and upper air charts. The scales are in 4mb and 60m intervals. An example of a geostrophic wind scale is shown in figure 3-13. Note that latitude

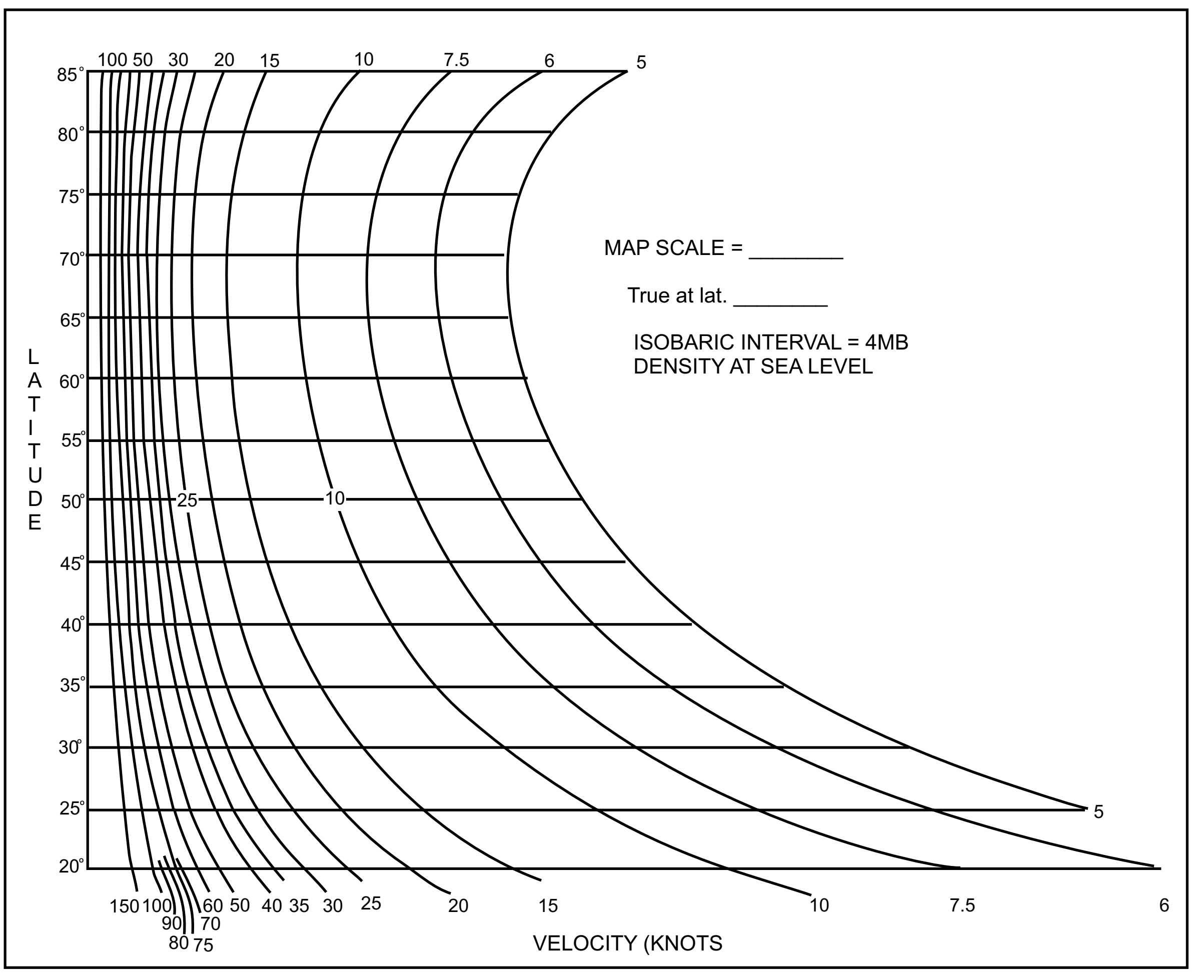

Figure 3-13.—Geostrophic wind scale.

accounts for the increases in gradients. In tropical regions, the geostrophic wind scales become less reliable because pressure gradients are generally rather weak.

REVIEW QUESTIONS

Q3-1. Which two factors influence the Earth's temperature?

Q3-2. What are the major factors that result in the wide distribution of pressure over the earth's surface?

Q3-3. What effect does Coriolis force have on thermal circulation in the Northern Hemisphere?

Q3-4. According to the 3-cell theory, how many circulation belts are there?

Q3-5. According to the 3-cell theory, what type of pressure system would you normally find at 30 degrees north latitude?

Q3-6. What is the predominant wind system in the tropics?

Q3-7. Name two types of pressure gradient.

Q3-8. What is the difference between centrifugal force and centripetal force?

Q3-9. What is the difference between gradient wind and geostrophic wind?

Q3-10. What is the relationship between centrifugal force and pressure gradient force around anticyclones?

SECONDARY CIRCULATION

LEARNING OBJECTIVE: Determine how centers of action, migratory systems, and seasonal variations affect secondary air circulations.

Now that you have a picture of the *general* circulation of the atmosphere over Earth, the next step is to see how land and water areas offset the general circulation. The circulations caused by the effect of Earth’s surfaces, its composition and contour, are known as *secondary circulations*. These secondary circulations give rise to winds that often cancel out the normal effect of the great wind systems.

There are two factors that cause the pressure belts of the primary circulation to break up into closed circulations of the secondary circulations. They are the non-uniform surface of the earth and the difference between heating and cooling of land and water. The surface temperature of oceans changes very little during the year. However, land areas sometimes undergo extreme temperature changes with the seasons. In the winter, large high-pressure areas form over the cold land and the low-pressure areas form over the relatively warm oceans. *The reverse is true in summer when highs are over water and lows form over the warm land areas. The result of this difference in heating and cooling of land and water surfaces is known as the thermal effect.*

Circulation systems are also created by the interaction of wind belts of pressure systems or the variation in wind in combination with certain distributions of temperature and/or moisture. This is known as the dynamic effect. This effect rarely, if ever, operates alone in creating secondary systems, as most of the systems are both created and maintained by a combination of the thermal and dynamic effects.

CENTERS OF ACTION

The pressure belts of the general circulation are rarely continuous. They are broken up into detached areas of high and low pressure cells by the secondary circulation. The breaks correspond with regions showing differences in temperature from land to water surfaces. Turn back to figures 3-2A and 3-2B. Compare the temperature distribution in views A and B of figures 3-2 to the pressure distribution in views A and B of figure 3-3. Note the gradient over the Asian Continent in January. Compare it to the warmer temperature over the ocean and coastal regions. Now look at view A of figure 3-3 and note the strong region of high-pressure corresponding to the area. Now look at the same area in July. Note the way the temperature gradient flattens out and warms. Look at view B of figure 3-3 and see the low-pressure area that has replaced the high-pressure region of winter. These pressure cells tend to persist in a particular area and are called centers of action; that is, they are found at nearly the same location with somewhat similar intensity during the same month each year.

There is a permanent belt of relatively low pressure along the equator and another deeper belt of low-pressure paralleling the coast of the Antarctic Continent. Permanent belts of high pressure largely encircle Earth, generally over the oceans in both the Northern and Southern Hemispheres. The number of centers of action is at a maximum at about 30 to 35 degrees from the equator.

There are also regions where the pressure is predominantly low or high at certain seasons, but not throughout the year. In the vicinity of Iceland, pressure is low most of the time. The water surface is warmer (due to warm ocean currents) than the surface of Iceland or the icecaps of Greenland. The Icelandic low is most intense in winter, when the greatest temperature contrast occurs, but it persists with less intensity through the summer. Near Alaska, a similar situation exists with the Aleutian low. The Aleutian low is most pronounced when the neighboring areas of Alaska and Siberia are snow covered and colder than the adjacent ocean.

These lows are not a continuation of one and the same cyclone. They are, however, regions of low pressure where lows frequently form or arrive from other regions. Here they remain stationary or move sluggishly for a time, then the lows move on or die out and are replaced by others. Occasionally these regions of low pressure are invaded by traveling high-pressure systems.

Two areas of semi permanent high-pressure also exist. There is a semi permanent high-pressure center over the Pacific westward of California and another over the Atlantic, near the Azores and of the coast of Africa. Pressure is also high, but less persistently so, west of the Azores to the vicinity of Bermuda. These subtropical highs are more intense and cover a greater area in summer than winter. They also extend farther northward summer. In winter, these systems move soul toward the equator, following the solar equator.

The largest individual circulation cells in the Northern Hemisphere are the Asiatic high in winter and the Asiatic low in summer. In winter, the Asiatic continent is a region of strong cooling and therefore is dominated by a large high-pressure cell. In summer, strong heating is present and the high-pressure cell becomes a large low-pressure cell. (See fig. 3-3A and fig. 3-3B.) This seasonal change in pressure cells gives rise to the monsoon flow over India and Southeast Asia.

Another cell that is often considered to be a center of action is the polar high. Both Arctic and Antarctic highs have considerable variations in pressure, and these regions have many traveling disturbances in summer. For example, the Greenland high (due to the Greenland icecap) is a persistent feature, but it is not a well-defined high during all seasons of the year. The Greenland high often appears to be an extension of the polar high or vice versa. Other continental regions show seasonal variations, but are generally of small size and their location is variable. Therefore, they are not considered to be centers of action.

An average annual pressure distribution chart (figure 3-14) reveals several important characteristics. First, along the equator there is a belt of relatively low pressure encircling the globe with barometric pressure of about 1,012 millibars. Second, on either side of this belt of low pressure is a belt of high pressure. This high-pressure area in the Northern Hemisphere lies mostly between latitudes 30° and 40°N with three well-defined centers of maximum pressure. One is over the eastern Pacific, the second over the Azores and the third over Siberia; all are about 1,020 millibars. The belt of high pressure in the Southern Hemisphere is roughly parallel to 30°S. It, too, has three centers of maximum pressure. One is in the eastern Pacific, the second in the eastern Atlantic, and the third in the Indian Ocean; again, all are about 1,020 millibars. A third characteristic to be noted from this chart is that, beyond the belt of high pressure in either hemisphere, the pressure diminishes toward the poles. In the Southern Hemisphere, the decrease in pressure toward the South Pole is regular and very marked. The pressure decreases from, an average slightly above 1,016 millibars along latitude 35°S to an average of 992 millibars along latitude 60°S In the Northern Hemisphere, however, the decrease in pressure toward the North Pole is less regular and not as great. This is largely due to the distribution of land and water: note the extensive landmass in the Northern Hemisphere as compared to those of the Southern Hemisphere.

While the pressure belts that stand out on the average annual pressure distribution chart represent average pressure distribution for the year, these belts are rarely continuous on any given day. They are usually broken up into detached areas of high or low pressure by the secondary circulation of the atmosphere. In either hemisphere, the pressure over the land during the winter season is decidedly above the annual average. During the summer season, the pressure is decidedly below the average, with extreme variations occurring such as in the case of continental Asia. Here the mean monthly pressure ranges from about 1,033 millibars during January to about 999 millibars during July. Over the northern oceans, on the other hand, conditions are reversed; the summer pressure there is somewhat higher. Thus in January the Icelandic and Aleutian lows intensify to a depth of about 999 millibars, while in July these lows fill and are almost obliterated.

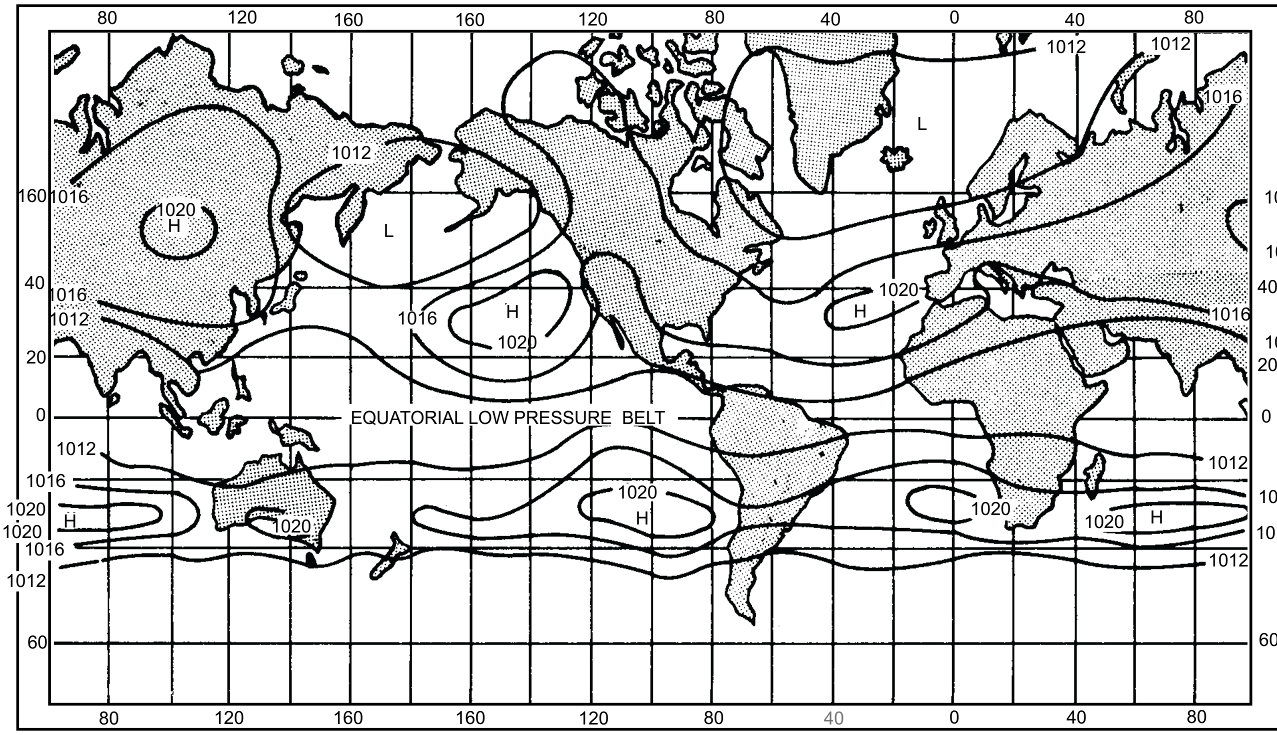

Figure 3-14.—Average annual pressure distribution chart.

The polar high in winter is not a cell centered directly over the North Pole, but appears to be an extension of the Asiatic high and often appears as a wedge extending from the Asiatic continent. The cell is displaced toward the area of coldest temperatures—the Asiatic continent. In summer, this high appears as an extension of the Pacific high and is again displaced toward the area of coolest temperature, which in this case is the extensive water area of the Pacific.

In winter over North America, the most significant feature is the domination by a high-pressure cell. This cell is also due to cooling but is not as intense as the Asiatic cell. In summer, the most significant feature is the so-called heat low over the southwestern part of the continent, which is caused by extreme heating in this region.

MIGRATORY SYSTEMS

General circulation, based on an average of wind conditions, is a more or less quasi-stationary circulation. Likewise, much of the secondary circulation depends on more or less static conditions that, in turn, depend on permanent and semi permanent high and low-pressure areas. Changes in the circulation patterns discussed so far have been largely seasonal. However, secondary circulation also includes wind systems that migrate constantly, producing rapidly changing weather conditions throughout all seasons, especially in the middle latitudes. The migratory circulation systems are associated with air masses, fronts, cyclones, and anticyclones. These are covered in detail in the next unit.

Anticyclones

An anticyclone (high) is an area of relatively high pressure with a clockwise flow (wind circulation) in the Northern Hemisphere and counterclockwise flow in the Southern Hemisphere. The windflow in an anticyclone is slightly across the isobars and away from the center of the anticyclone. (See fig. 3-15.) Anticyclones are commonly called highs or high-pressure areas.

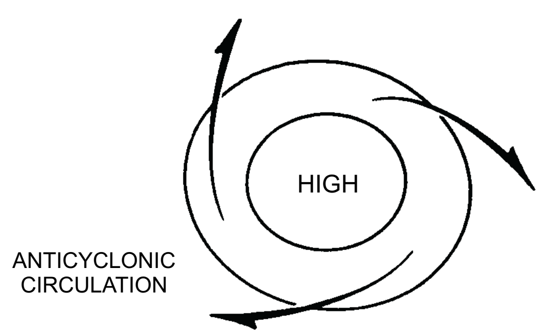

Figure 3-15.—Anticyclone.

The formation of an anticyclone or the intensification of an existing one is called *anticyclogenesis*. Anticyclogenesis refers to the development of anticyclonic circulation as well as the intensification of an existing anticyclonic flow. When a high-pressure center is increasing in pressure, the high is BUILDING or INTENSIFYING. Although a high can build (or intensify) without an increase in anticyclonic flow, it is rare. Normally, building and anticyclogenesis occur simultaneously. The weakening of anticyclonic circulation is *anticyclolysis*. When the pressure of a high is decreasing, we say the high is weakening. Anticyclolysis and weakening can occur separately, but usually occur together.

The vertical extent of pressure greatly depends on the air temperature. Since density increases with a decrease in temperature, pressure decreases more rapidly vertically in colder air than in warmer air. In a cold anticyclone (such as the Siberian high), the vertical extent is shallow; while in a warm anticyclone (such as the subtropical high), the vertical extent reaches high into the upper atmosphere due to the slow decrease in temperature with elevation.

Cyclones

A cyclone (low) is a circular or nearly circular area of low pressure with a counterclockwise flow. The flow is slightly across the isobars toward the center in the Northern Hemisphere and clockwise in the Southern Hemisphere. (See fig. 3-16.) It is commonly called a low or a depression. This use of the word cyclone should be distinguished from the colloquial use of the word as applied to the tornado or tropical cyclone (hurricane).

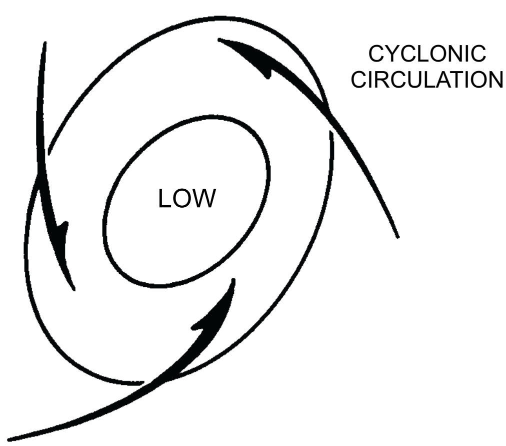

Figure 3-16.—Cyclone.

The formation of a new cyclone or the intensification of the cyclonic flow in an existing one is called *cyclogenesis*. When the pressure in the low is falling, we say the low is deepening. Cyclogenesis and deepening can also occur separately, but usually occur at the same time. The decrease or eventual dissipation of a cyclonic flow is called *cyclolysis*. When the pressure in a low is rising, we say the low is filling. Cyclolysis and filling usually occur simultaneously. Cyclones in middle and high latitudes are referred to as extratropical cyclones. The term *tropical cyclone* refers to hurricanes and typhoons.

VERTICAL STRUCTURE OF SECONDARY CIRCULATIONS (PRESSURE CENTERS)

To better understand the nature of the pressure centers of the secondary circulation, it is necessary to consider them from a three-dimensional standpoint. With the aid of surface and upper air charts, you will be able to see the three dimensions of these pressure systems as well as the circulation patterns of the secondary circulation as established at higher levels in the troposphere and lower stratosphere.

In Chapter 2, the study of gas laws showed that volume is directly proportional to temperature. Stated another way, we might say that the thickness of a layer between two isobaric surfaces is directly proportional to the mean virtual temperature of the layer. Because the atmosphere is always moist to some degree, virtual temperature is used. Mean virtual temperature is defined as the average temperature at which dry air would have the same pressure and density as moist air. Thus, lines representing thickness are also isotherms of mean virtual temperature. The higher the mean virtual temperature, the thicker the layer, or vice versa. The thickness between layers is expressed in geopotential meters. The shift in location, as well as the change of shape and intensity upward of atmospheric pressure systems, is dependent on the temperature distribution.

An example of the effects of virtual temperature can be demonstrated by placing two columns of air next to each other. One air column is cold and the other air column is warm. The constant pressure surfaces in the cold column are closer than the ones in the warm column. Figure 3-17 shows an increase in thickness between two pressure surfaces, resulting in an increase in mean virtual temperature. Note the increase in the distance between the constant pressure surfaces; P, P1, etc., from column A to column B. Using the hypsometric equation can derive the thickness value between two pressure surfaces. Thickness may also be determined from tables, graphs, etc.

VERTICAL STRUCTURE OF HIGH PRESSURE SYSTEMS

The topographic features that indicate the circulation patterns at 500 millibars in the atmosphere correspond in general to those at lower and higher level. However, they may experience a shift in location as well as a change in intensity and shape. For example, a ridge aloft may reflect a closed high on a surface synoptic chart. In addition, upper air circulation patterns may take on a wavelike structure in contrast to the alternate closed lows, or closed high patterns at the surface level. The smoothing of the circulation pattern aloft is typical of atmospheric flow patterns.

Cold Core Highs

A cold core high is one in which the temperatures on a horizontal plane decrease toward the center.

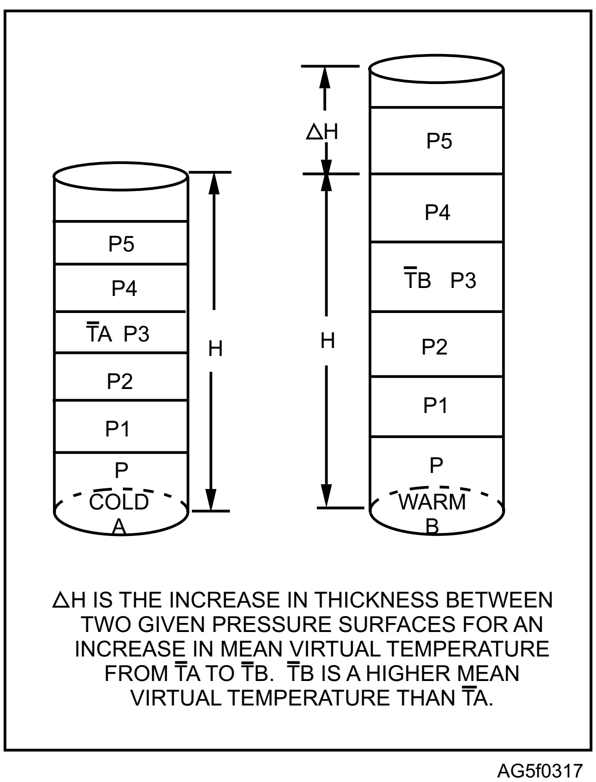

Figure 3-17.—Thickness of two strata as a function of means virtual temperature.

Because the temperature in the center of a cold core high is less than toward the outside of the system, it follows that the vertical spacing of isobars in the center of this system is closer together than on the outside. Although the pressure at the center of these systems on the surface may be high, the pressure decreases rapidly with height. (See fig. 3-18.) Because these highs are often quite shallow, it is common for an upper level low to exist above a cold core high.

NOTE: For the purpose of illustration, figures 3-18 through 3-21 are exaggerated with respect to actual atmospheric conditions.

If the cold core high becomes subjected to warming from below and to subsidence from aloft, as it moves southward from its source and spreads out, it diminishes rapidly in intensity with time (unless some dynamic effect sets in aloft over the high to compensate for the warming). Since these highs decrease in intensity with height, thickness is relatively low. In the vertical, cold core highs slope toward colder air aloft. Anticyclones found in Arctic air are always cold cored, while anticyclones in polar air may be warm or cold core.

Examples of cold core highs are the North American High, the Siberian High and the migratory highs that originate from these anticyclones.

Warm Core Highs

A warm core high is one in which the temperatures on a horizontal level increase toward the center. Because the temperatures in the center of a warm core high are higher than on the outside of the system, it follows that the vertical spacing of isobars in the center is farther apart than toward the outside of the high. For this reason, a warm core high increases in intensity with altitude and has an anticyclonic circulation at all levels (see fig. 3-19). From a vertical view, warm core highs slope toward warmer air aloft. A warm core high is accompanied by a high cold tropopause. Since the pressure surfaces are spaced far apart, the tropopause is reached only at great heights. The temperature continues to decrease with elevation and is cold by the

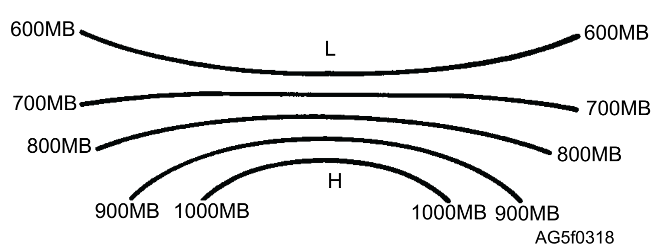

Figure 3-18.—Cold core high.

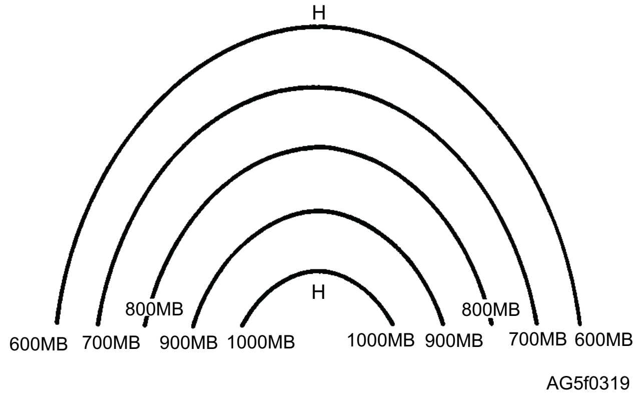

Figure 3-19.—Warm core high.

time the tropopause is reached. The subtropical highs are good examples of this type of high. Therefore, anticyclones found in tropical air are always warm core. Examples of warm core highs are the Azores or Bermuda High and the Pacific High.

VERTICAL STRUCTURE OF LOW-PRESSURE SYSTEMS

Low-pressure systems, like high-pressure systems, are generally a reflection of systems aloft. They, too, experience shifts in location and changes in intensity and shape with height. At times, a surface system may not be evident aloft and a well-developed system aloft may not reflect on a surface analysis.

Cold Core Lows

The cold core low contains the coldest air at its center throughout the troposphere; that is, going outward in any direction at any level in the troposphere, warmer air is encountered. The cold core low (figure 3-20) increases intensity with height. Relative minimums in thickness values, called cold pools, are found in such cyclones. The temperature distribution is almost symmetrical, and the axis of the low is nearly vertical. When they do slope vertically, they slope toward the coldest temperatures aloft. In the cold low, the lowest temperatures coincide with the lowest pressures.

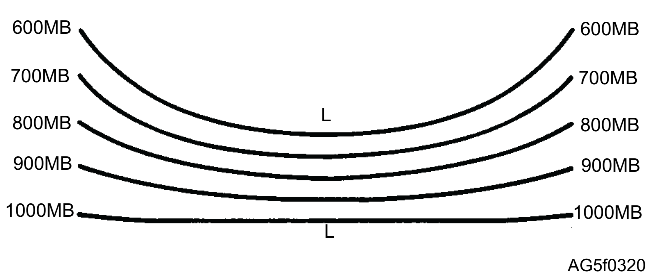

Figure 3-20.—Cold core low.

The cold low has a more intense circulation aloft from 850 to 400 millibars than at the surface. Some cold lows show only slight evidence in the surface pressure field that an intense circulation exists aloft. The cyclonic circulation aloft is usually reflected on the surface in an abnormally low daily mean temperature often accompanied by instability and showery precipitation. A cold core low is accompanied by a low warm tropopause. Since the pressure surfaces are close together, the tropopause is reached at low altitudes where the temperature is relatively warm. Good examples of cold core lows are the Aleutian and Icelandic lows. Occluded cyclones are generally cold core in their later stages, because polar or arctic air has closed in on them.

At high latitudes the cold pools and their associated upper air lows show some tendency for location in the northern Pacific and Atlantic Oceans where, statistically, they contribute to the formation of the Aleutian and Icelandic lows.

Warm Core Lows

A warm core low (figure 3-21) decreases intensity with height and the temperature increases toward the center on a horizontal plane. The warm low is frequently stationary, such as the heat low over the southwestern United States in the summer; this is a result of strong heating in a region usually insulated from intrusions of cold air that tend to fill it or cause it to move. The warm low is also found in its moving form as a stable wave moving along a frontal surface. There is no warm low aloft in the troposphere. The tropical cyclone, however, is believed to be a warm low because its intensity diminishes with height. Because most warm lows are shallow, they have little slope. However, intense warm lows like the heat low over the southwest United States and hurricanes do slope toward warm air aloft.

In general, the temperature field is quite asymmetrical around a warm core cyclone. Usually the southward moving air in the rear of the depression is

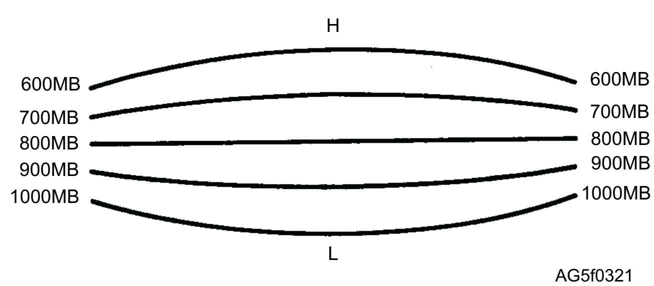

Figure 3-21.—Warm core low.

not as warm as that moving northward in advance of it. A warm core low decreases intensity with height or completely disappears and are often replaced by anticyclones aloft. The heat lows of the southwestern United States, Asia, and Africa are good examples of warm core lows. Newly formed waves are generally warm core because of the wide-open warm sector.

DYNAMIC LOW

Systems that retain their closed circulations to appreciable altitudes and are migratory are called dynamic lows or highs. A dynamic low is a combination of a warm surface low and a cold upper low or trough, or a warm surface low in combination with a dynamic mechanism aloft for producing a cold upper low or trough. It has an axis that slopes toward the coldest tropospheric air. (See figure 3-22.) In the final stage, after occlusion of the surface warm low is complete, the dynamic low becomes a cold low with the axis of the low becoming practically vertical.

DYNAMIC HIGH

The dynamic high is a combination of a surface cold high and an upper-level warm high or well-developed ridge, or a combination of a surface cold high with a dynamic mechanism aloft for producing high-level anticyclogenesis. Dynamic highs have axes that slope toward the warmest tropospheric air. (See fig. 3-22.) In the final stages of warming the cold surface high, the dynamic high becomes a warm high with its axis practically vertical.

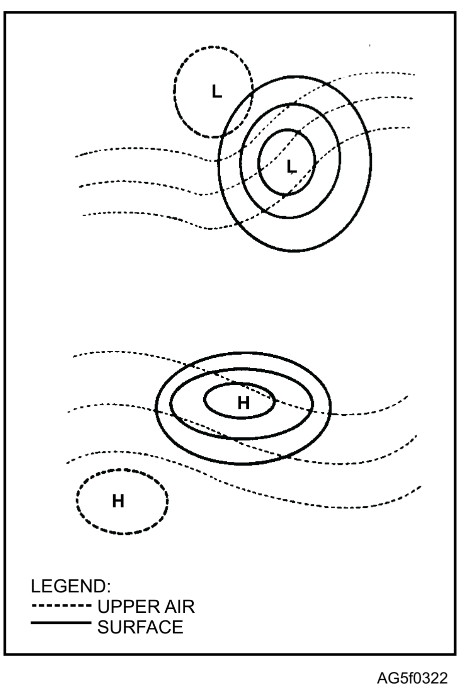

Figure 3-22.—Vertical slope of low -pressure systems.

REVIEW QUESTIONS

Q3-11. What is the term that defines the formation of an anticyclone or the intensification of an existing anticyclone?

Q3-12. What is the direction of the windflow around a cyclone?

Q3-13. How do temperatures change within a cold core low?

Q3-14. Low pressure due to intense heating over the southwestern United States is an example of which type of low-pressure system?

TERTIARY CIRCULATION

LEARNING OBJECTIVE: Define tertiary circulation and describe how tertiary circulations affect local weather and wind direction and speed.

Tertiary (third order) circulations are localized circulations directly attributable to one of the following causes or a combination of them: local cooling, local heating, adjacent heating or cooling, and induction (dynamics).

Many regions have local weather phenomena caused by temperature differences between land and water surfaces or by local topographical features. These weather phenomena show up as circulations. These tertiary circulations can result in dramatic local weather conditions and wind flows. The most common tertiary circulations are discussed in this lesson. However, there are numerous other circulations and related phenomena in existence around the world.

MONSOON WINDS

The term *monsoon* is of Arabic origin and means season. The monsoon wind is a seasonal wind that blows from continental interiors (or large land areas) to the ocean in the winter; they blow in the opposite direction during the summer. The monsoon wind is most pronounced over India, although there are other regions with noticeable monsoon winds.

Monsoon winds are a result of unequal heating and cooling of land and water surfaces. During winter a massive area of cold high pressure develops over the extensive Asiatic continent. This high pressure is due primarily to cold arctic air and long-term radiation cooling. To the south, the warm equatorial waters exist and, in contrast, the area has relatively lower surface pressures. The combination of high pressure over Asia

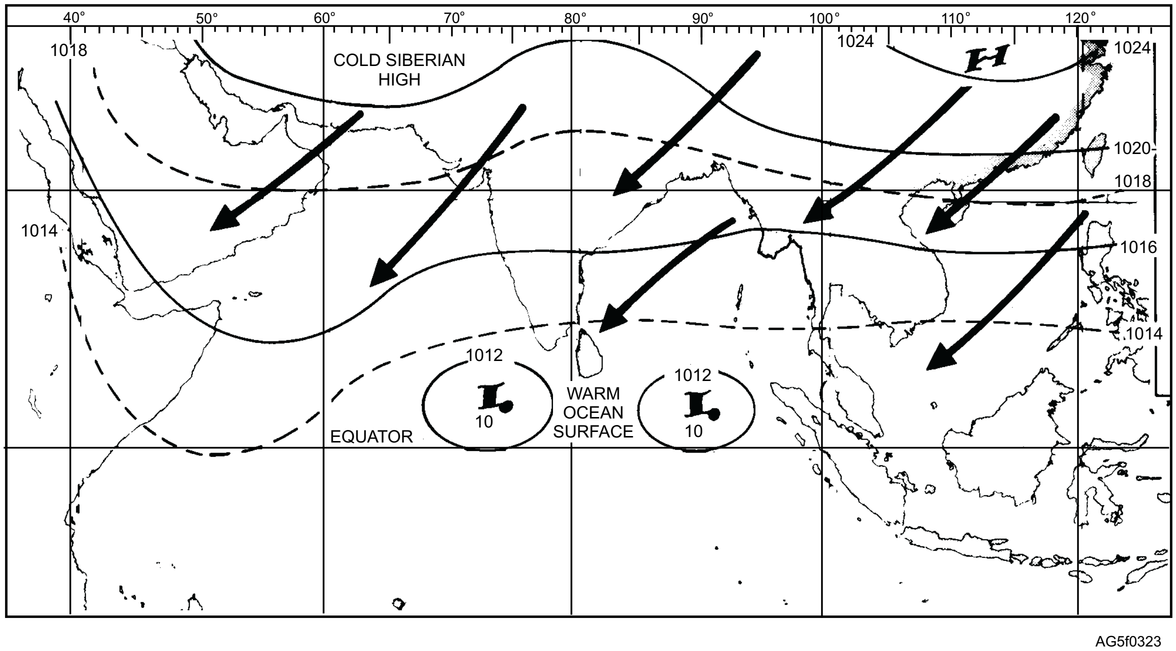

Figure 3-23.—Northeast monsoon (January).

and low pressure over the Equatorial Belt sets up a pressure gradient directed from north to south. Because of the flow around the massive Siberian high, northeast winds begin to dominate the regions from India to the Philippines. (See fig. 3-23).

During the winter months, clear skies predominate over most of the region. This is caused by the mass motion of air from a high-pressure area over land to an area of lower pressure over the ocean. As the air leaves the high-pressure area over land, it is cold and dry. As it travels over land toward the ocean, there is no source of moisture to induce precipitation. The air is also traveling from a higher altitude to a lower altitude; consequently, this downslope motion causes the air to be warmed at the adiabatic lapse rate. This warming process has a still further clearing effect on the skies.

During the summer the airflow over the region is completely reversed. The large interior of Asia is heated to the point where the continent is much warmer than the ocean areas to the south. This induces relatively low pressure over Asia and higher pressure over the equatorial region. This situation produces a southwesterly flow as shown in figure 3-24.

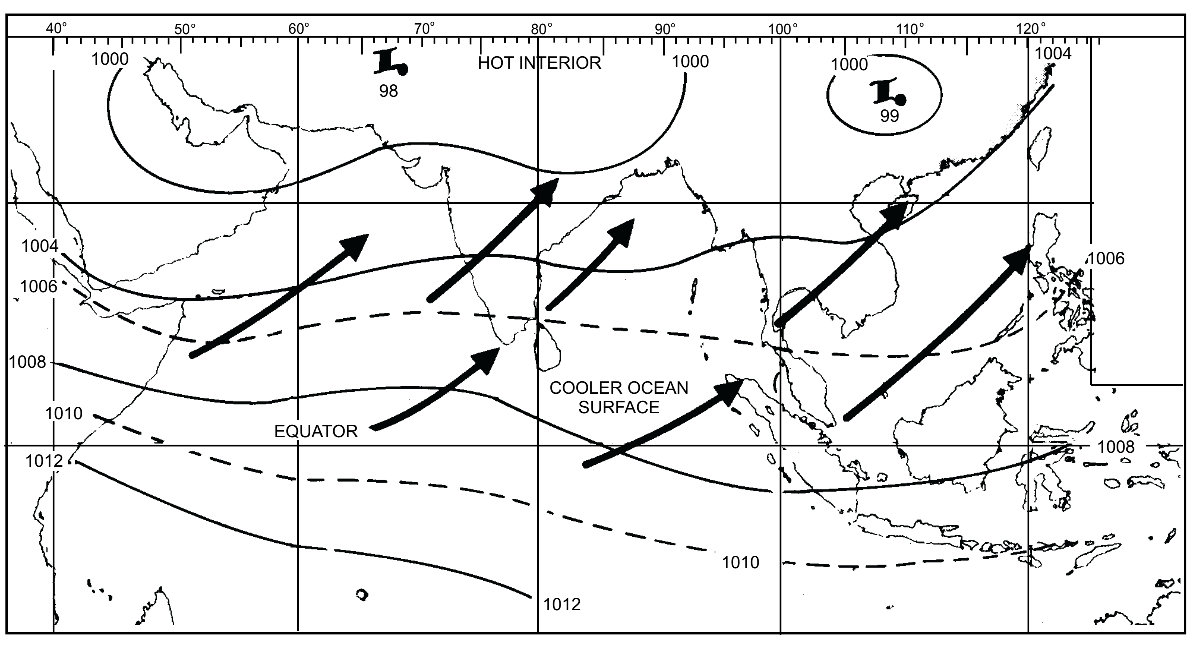

Figure 3-24.—Southwest monsoon (July).

The weather associated with the summer monsoon winds is thunderstorms, almost constant heavy rain, rain showers, and gusty surface winds. This condition is caused by mass motion of air from the relatively high-pressure area over the ocean to a low-pressure area over land. When the air leaves the ocean, it is warm and moist. As the air travels over land toward the low-pressure area, it is also traveling from a lower altitude to a higher altitude. The air is lifted by a mechanical force and cooled to its condensation point by this upslope motion (pseudo adiabatic process).

LAND AND SEA BREEZES

There is a diurnal (daily) contrast in the heating of local water and land areas similar to the seasonal variation of the monsoon. During the day, the land is warmer than the water area; at night the land area is cooler than the water area. A slight variation in pressure is caused by this temperature contrast. At night the wind blows from land to sea and is called a land breeze. During the day, the wind blows from water areas to land areas and is called a sea breeze.

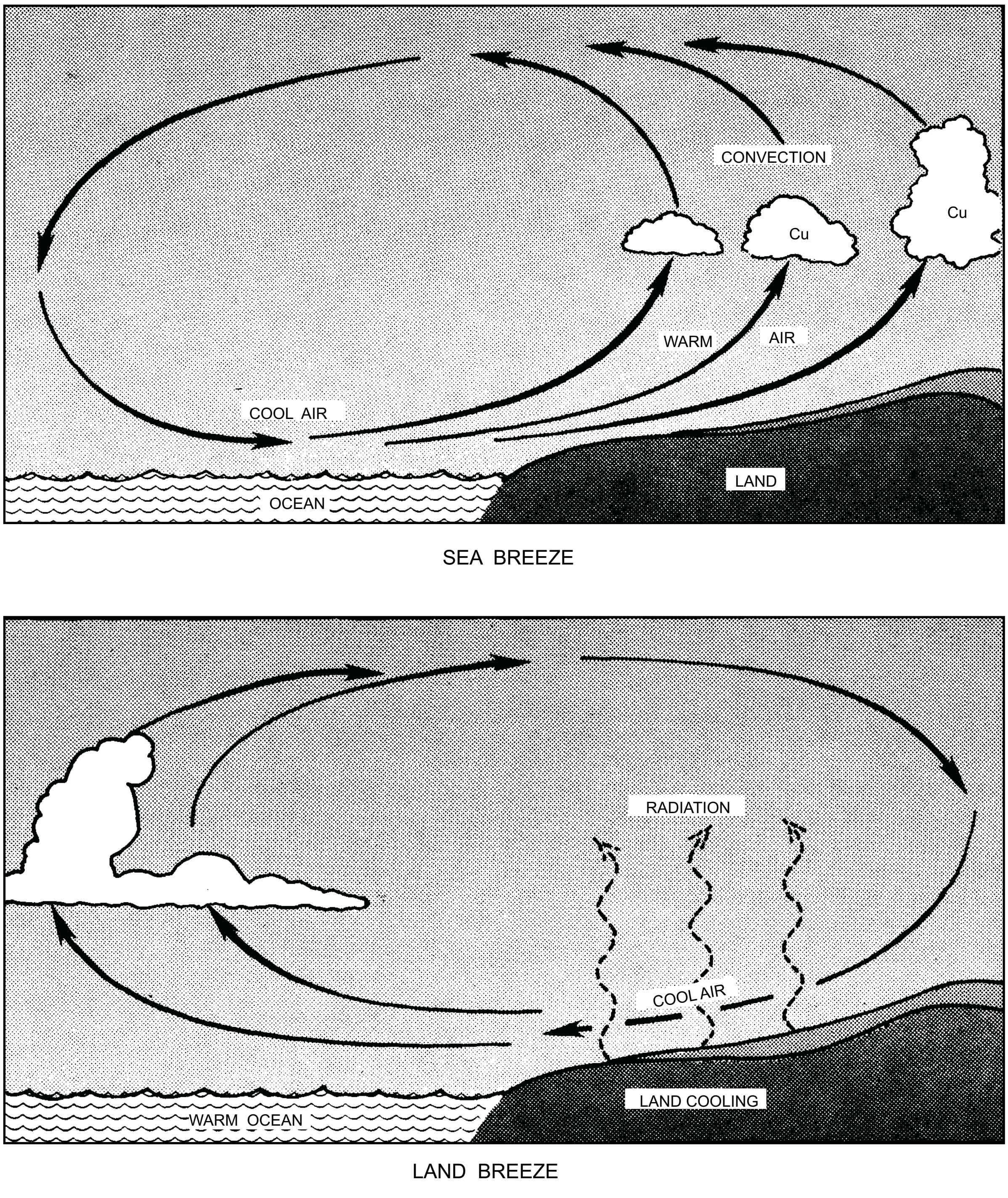

Figure 3-25.—Circulation of land and sea breezes.

The sea breeze usually begins during midmorning (0900-1100 local time) when the land areas become warmer than adjacent ocean waters (see fig. 3-25). This temperature difference creates an area of slightly lower surface pressures over land compared to the now cooler waters. The result is a wind flow from water to land. The sea breeze starts with a shallow flow along the surface; however, as maximum heating occurs, the flow increases with height. The height varies from an average of 3,000 feet in moderately warm climates to 4,500 (or more) in tropical regions. The effects of the sea breeze can be felt as far as 30 miles both onshore and offshore. In extreme cases, the sea breeze is felt 100 miles inland depending upon terrain. By mid afternoon (maximum heating) the sea breeze will reach its maximum speed and may be strong enough to be influenced by the Coriolis force, which causes it to flow at an angle to the shore. The sea breeze is most pronounced in late spring, summer, and early fall when maximum temperature differences occur between land and water surfaces. A decrease in temperature and an increase in humidity and wind speed mark the start of a sea breeze.

The sea breeze continues until the land area cools sufficiently to dissipate the weak low pressure. After sunset, the land cools at a faster rate than the adjacent waters and eventually produces a complete reversal of the winds. As the land continues to cool through the evening hours, a weak area of high pressure forms over the land. The water area, with its warmer temperatures, has slightly lower pressure and again a flow is established; however, the flow is now from land to water (offshore). (See fig. 3-25.)

The land breezes, when compared to the sea breezes, are less extensive and not as strong (usually less than 10 knots and less than 10 miles offshore). This is because there is less temperature contrast at night between land and water surfaces as compared to the temperature contrast during daytime heating. Land breezes are at maximum development late at night, in late fall and early winter. In the tropical land regions, the land and sea breezes are repeated day after day with great regularity. In high latitudes the land and sea breezes are often masked by winds of synoptic features.

WINDS DUE TO LOCAL COOLING AND HEATING

In the next sections we discuss tertiary circulations due to local cooling and heating effects. Under normal circumstances, these winds attain only light to moderate wind speeds; however, winds often occur in and near mountain areas that have undergone dramatic changes in normal character. At times, mountain areas tend to funnel winds through valleys and mountain passes. This funneling effect produces extremely dangerous wind speeds.

FUNNEL EFFECT

Winds blowing against mountain barriers tend to flatten out and go around or over them. If a pass or a valley breaks the barrier, the air is forced through the break at considerable speed. When wind is forced through narrow valleys it is known as the funnel effect and is explained by Bernoulli's theorem. According to Bernoulli's theorem, pressures are least where velocities are greatest; likewise, pressures are greatest where velocities are least. This observation is true for both liquids and gases. (See fig. 3-26.)

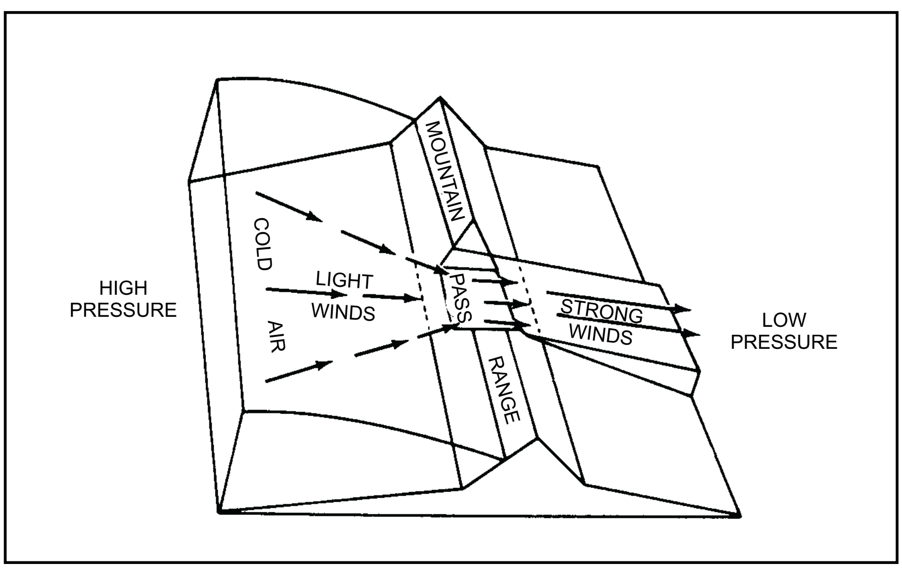

Figure 3-26.—Strong wind produced by funneling.

Bernoulli's theorem is frequently used to forecast tertiary winds in the mountainous western United States. The famous Santa Ana winds of southern California are a prime example. Winds associated with high pressure situated over Utah are funneled through the valley leading into the town of Santa Ana near the California coast. Low pressure develops at the mouth of the valley and the end result is hot, dry, gusty and extremely dangerous winds. When the Santa Ana is strong enough, the effects are felt in virtually every valley located along the coast of southern California. Visibility is often restricted due to blowing sand. It is common to see campers, trailers, and trucks turned over by the force of these winds. When funneled winds reach this magnitude, they are called jet-effect winds, canyon winds, or mountain-gap winds.

Winds Due to Local Cooling and Heating

There are two types of tertiary circulations produced by local cooling—glacier winds and drainage winds.

GLACIER WINDS.—Glacier winds, or fall winds (as they are sometimes called) occur in many varieties in all parts of the world where there are glaciers or elevated land masses that become covered by snow and ice during winter. During winter, the area of snow cover becomes most extensive. Weak pressure results in a maximum of radiation cooling. Consequently the air coming in contact with the cold snow cools. The cooling effect makes the overlying air more dense, therefore, heavier than the surrounding air. When set in motion, the cold dense air flows down the sides of the glacier or plateau. If it is funneled through a pass or valley, it may become very strong. This type of wind may form during the day or night due to radiation cooling. The glacier wind is most common during the winter when more snow and ice are present.

When a changing pressure gradient moves a large cold air mass over the edge of a plateau, this action sets in motion the strongest, most persistent, and most extensive of the glacier or fall winds. When this happens, the fall velocity is added to the pressure gradient force causing the cold air to rush down to sea level along a front that may extend for hundreds of miles. This condition occurs in winter on a large scale along the edge of the Greenland icecap. In some places along the icecap, the wind attains a velocity in excess of 90 knots for days at a time and reaches more than 150 nautical miles out to sea.

Glacier winds are cold katabatic (downhill) winds. Since all katabatic winds are heated adiabatically in their descent, they are predominantly dry. Occasionally, the glacier winds pick up moisture from falling precipitation when they underride warm air. Even with the adiabatic heating they undergo, all glacier or fall winds are essentially cold winds because of the extreme coldness of the air in their source region. Contrary to all other descending winds that are warm and dry, the glacier wind is cold and dry. It is colder, level for level, than the air mass it is displacing. In the Northern Hemisphere, the glacier winds descend frequently from the snow-covered plateaus and glaciers of Alaska, Canada, Greenland, and Norway.

DRAINAGE WINDS.—Drainage winds (also called mountain or gravity winds) are caused by the cooling air along the slopes of a mountain. Consequently, the air becomes heavy and flows downhill, producing the MOUNTAIN BREEZE.

Drainage winds are katabatic winds and like glacier winds; a weak or nonexistent pressure gradient is required to start the downward flow. As the air near the top of a mountain cools through radiation or contact with colder surfaces, it becomes heavier than the surrounding air and gradually flows downward (fig. 3-27). Initially this flow is light (2 to 4 knots) and only a few feet thick. As cooling continues, the flow increases achieving speeds up to 15 knots at the base of the mountain and a depth of 200 feet or more. Winds in excess of 15 knots are rare and only occur when the mountain breeze is severely funneled.

Drainage winds are cold and dry. Adiabatic heating does not sufficiently heat the descending air because of the relative coldness of the initial air and because the distance traveled by the air is normally short. Drainage winds have a very localized circulation. As the cold air enters the valley below, it displaces the warm air. Temperatures continue to fall. If the flow achieves speeds of 8 knots or more, mixing results between the warm valley air and the cold descending air that results in a slight temperature increase. Campers often prefer to make summer camps at the base of mountains to take advantage of the cooling effect of the mountain breeze.

Funnel Effects

VALLEY BREEZES.—The valley breeze is the anabatic (uphill) counterpart of the mountain breeze. When the sun heats the valley walls and mountain slopes during the morning hours, the air next to the ground is heated until it rises along the slopes. Rocky or

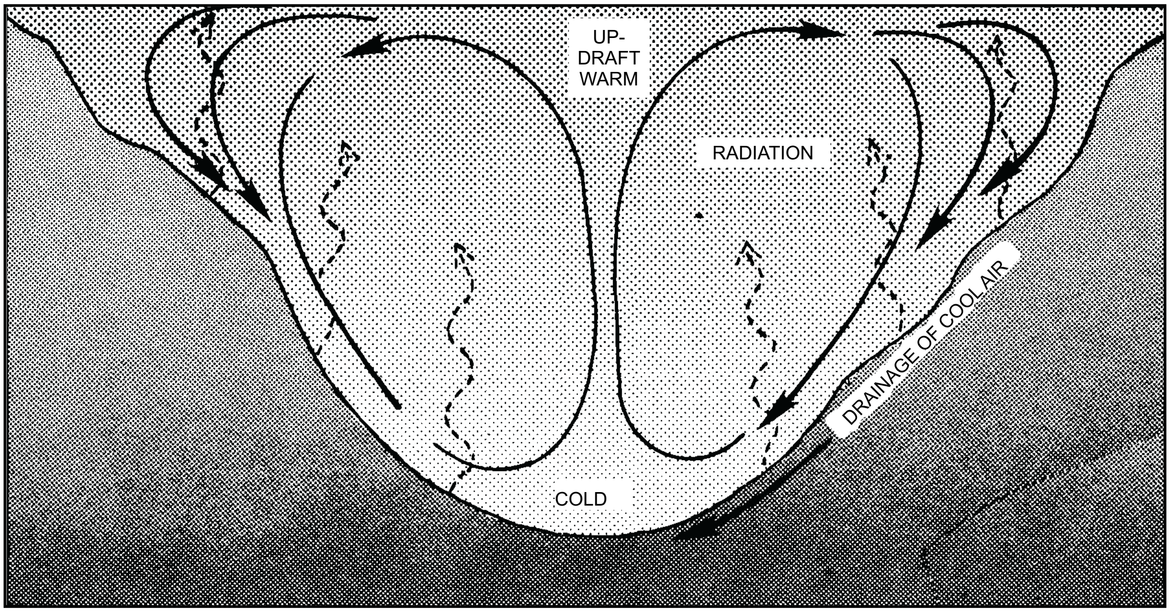

Figure 3-27.—Mountain breeze or katabatic wind. During the night outgoing radiation cools air along hillsides below free air temperature. The cooled air drains to lowest point of the terrain.

sandy slopes devoid of vegetation are the most effective heating surfaces. If the slopes are steep, the ascending breeze tends to move up the valley walls. The expansion of the heated air next to the surface produces a slight local pressure gradient against the ground surface. As the heating becomes stronger, convective currents begin to rise vertically from the valleys (figure 3-28). The updrafts along the valley walls continue to be active, particularly at the head of the valley. The valley breeze usually reaches its maximum strength in the early afternoon. It is a stronger and deeper wind than the mountain breeze. It is difficult to isolate the valley breeze effect because of the prevailing gradient winds. Consequently, the valley breeze is much more likely to be superposed as a prevailing wind than is the mountain breeze, which by its very nature can develop only in the absence of any appreciable gradient wind. The valley breezes are generally restricted to slopes facing south or the more direct rays of the sun, and they are more pronounced in southern latitudes. They are diurnally strongest in the late afternoon and are seasonally strongest in summer.

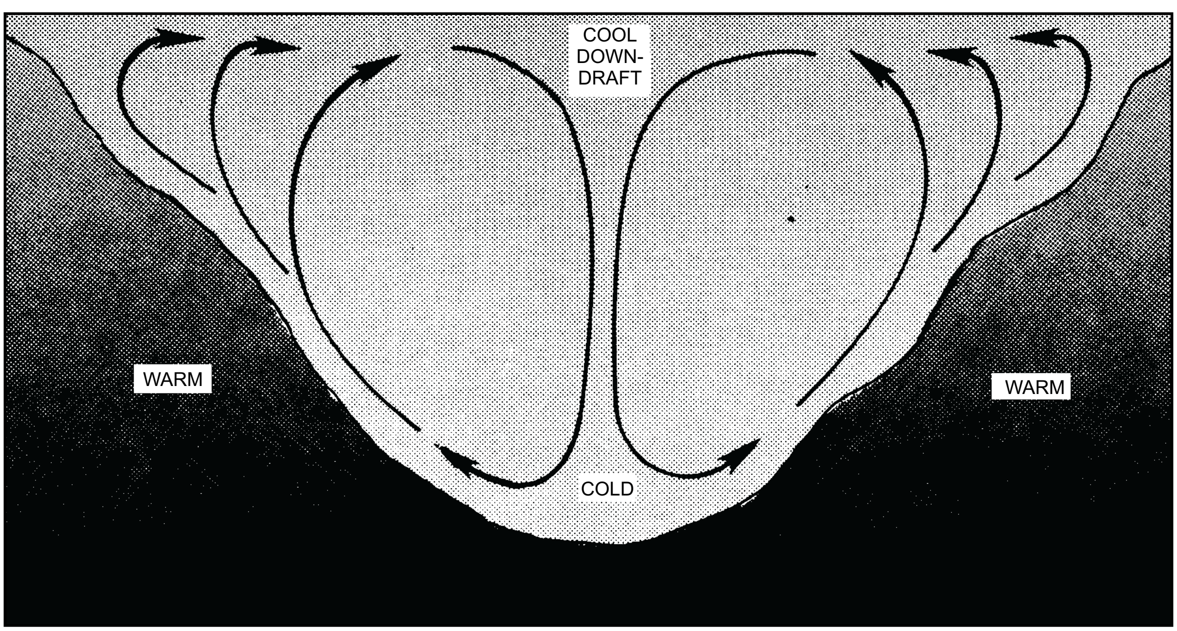

Figure 3-28.—Valley breeze or anabatic wind. During the daytime hillsides heat quickly. This heating effect causes updrafts along slopes—downdrafts in the center.

THERMALS.—Thermals are vertical convective currents that result from local heating. They stop short of the condensation level. Thermal convection is the usual result of strong heating of the lower atmosphere by the ground surface. A superadiabatic lapse rate immediately above the ground is necessary to the development of strong thermals. They form most readily over areas of bare rock or sand and in particular over sand dunes or bare rocky hills. In the presence of a moderate or fresh breeze, especially in a hilly terrain, it is impossible to distinguish between turbulent and thermal convection currents. Pure thermal convection normally occurs on clear summer days with very light prevailing wind. In the eastern United States, dry thermals are usually of only moderate intensity, seldom reaching an elevation in excess of 5,000 feet above the surface. The high moisture content of the air masses in this section in summer reduces the intensity of surface heating to some extent. This moisture content usually keeps the condensation level of the surface air near or even below a height of 5,000 feet above the ground. In the dry southwestern part of the country, where ground heating during clear summer days is extreme, dry thermal convection may extend to a height of 10,000 feet or more. Under these conditions, extremely turbulent air conditions can occur locally up to whatever heights the thermals extend, frequently without a cloud in the sky.

One variation of the dry thermal is seen in the dust or sand whirls, sometimes called dust devils. They are formed over heated surfaces when the winds are very light. Dust whirls are seldom more than two or three hundred feet high and they last only a few minutes at most. Over the desert on clear hot days as many as a dozen columns of whirling sand may be visible at once. The large desert sand whirls can become several hundred feet in diameter, extend to heights of 4,000 feet or higher, and in some cases last for an hour or more. They have been observed to rotate both anticyclonically and cyclonically, the same as tornadoes.

An almost identical phenomenon is observed over water in the form of the waterspout. Waterspouts occur frequently in groups and form in relatively cool humid air over a warm water surface when the wind is light. The waterspout is visible due to the condensed water vapor, or cloud formation, within the vortex. The condensation is the result of dynamic cooling by expansion within the vortex. In this respect it differs from the sand whirl, which is always dry. Both the sand whirl and the waterspout represent simple thermal convection of an extreme type. They are not to be confused with the more violent tornado.

When dry thermal convection extends to an elevation where the dry thermals reach the condensation level, then cumulus convection takes the place of the dry convection. A cumulus cloud, whose base is at the condensation level of the rising air, tops each individual thermal current. Beneath every building cumulus cloud a vigorous rising current or updraft is observed. Thus the local thermal convection pattern becomes visible in the cumulus cloud pattern. The cumulus clouds form first over the hills where the strongest thermals develop. Under stable atmospheric conditions, little convective cloud development occurs. However, under unstable conditions these thermals may develop cumulonimbus clouds.

INDUCED OR DYNAMIC TERTIARY CIRCULATIONS

There are four types of induced or dynamic tertiary circulations. They are eddies, turbulence, large-scale vertical waves, and Foehn winds.

Eddies

An eddy is a circulation that develops when the wind flows over or adjacent to rough terrain, buildings, mountains or other obstructions. They generally form on the lee (downwind or sheltered) side of these obstructions. The size of the eddy is directly proportional to the size of the obstruction and speed of the wind. Eddies may have horizontal or vertical circulations that can be either cyclonic or anticyclonic.

Horizontal eddies form in sheltered areas downwind of rough coastlines or mountain chains. An example of a horizontal eddy is the weak cyclonic circulation that develops in the channel off the coast of Santa Barbara, California. The winds frequently blow parallel to the northern California coastline during the winter fog and stratus season. The Santa Barbara channel often remains fog-free because the waters are protected from winds that transport the fog inland. However, when the winds are sufficiently strong, friction along the tough coastal range produces a weak cyclonic eddy over the channel. This cyclonic flow, though weak, is sufficient to advect fog into the region.

Vertical eddies are generally found on the lee side of mountains, but with low wind speeds, stationary eddies or rotating pockets of air are produced and remain on both the windward and leeward sides of obstructions. (See figure 3-29.) When wind speeds exceed about 20 knots, the flow may be broken up into irregular eddies that are carried along with a wind some distance downstream from the obstruction. These eddies may cause extreme and irregular variations in the wind and may disturb aircraft landing areas sufficiently to be a hazard.

A similar and much disturbed wind condition occurs when the wind blows over large obstructions such as mountain ridges. In such cases the wind blowing up the slope on the windward side is usually relatively smooth. On the leeward side the wind spills rapidly down the slope, setting up strong downdrafts and causing the air to be very turbulent. This condition is illustrated in figure 3-30. These downdrafts can be very violent. Aircraft caught in these eddies could be forced to collide with the mountain peaks. This effect is also noticeable in the case of hills and bluffs, but is not as pronounced.

Turbulence

Turbulence is the irregular motion of the atmosphere caused by the air flowing over an uneven surface or by two currents of air flowing past each other in different directions or at different speeds. The main source of turbulence is the friction along the surface of Earth. This is called mechanical turbulence. Turbulence is also caused by irregular temperature distribution. The warmer air rises and the colder air descends, causing an irregular vertical motion of air; this is called thermal turbulence.

Mechanical turbulence is intensified in unstable air and is weakened in stable air. These influences cause fluctuations in the wind with periods ranging from a few minutes to more than an hour. If these wind variations are strong, they are called wind squalls and are usually associated with convective clouds. They are an indication of approaching towering cumulus or cumulonimbus clouds.

Gustiness and turbulence are more or less synonymous. Gustiness is an irregularity in the wind speed that creates eddy currents disrupting the smooth airflow. Thus, the term *gust* is usually used in conjunction with sudden intermittent increases in the wind speed near the surface levels. Turbulence, on the other hand, is used with reference to levels above the surface. Gustiness can be measured; turbulence, however, unless encountered by aircraft equipped with a gust probe or an accelerometer, is usually estimated.

Large-Scale Vertical Waves (Mountain Waves)

Mountain waves occur on the lee side of topographical barriers and occur when the wind-flow is strong, 25 knots or more, and the flow is roughly perpendicular to the mountain range. The structure of the barrier and the strength of the wind determines the

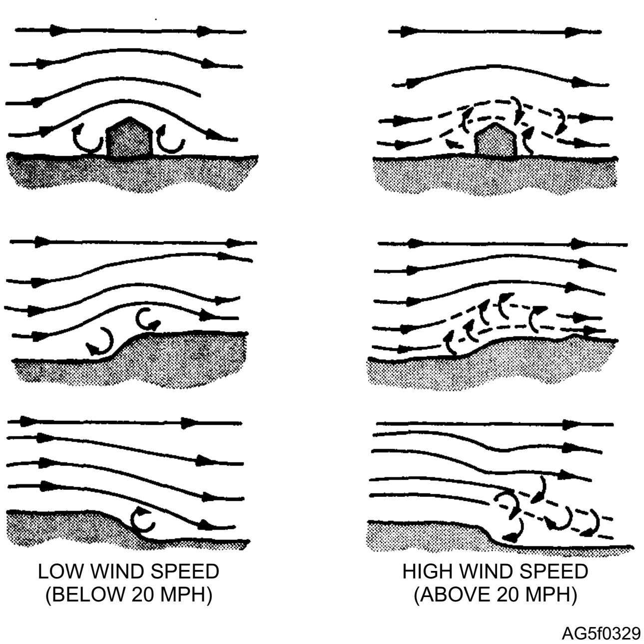

Figure 3-29.—Eddy currents formed when wind flows over uneven ground or obstructions.

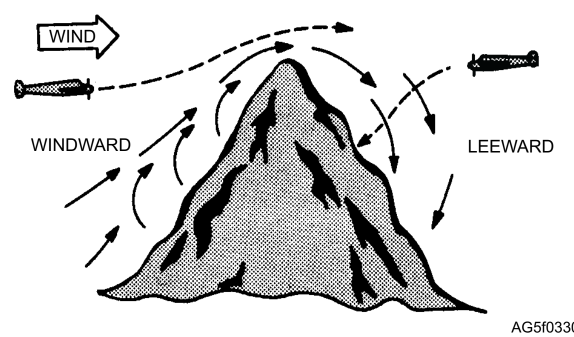

Figure 3-30.—Effect of windflow over mountains.

amplitude and the type of the wave. The characteristics of a typical mountain wave are shown in figure 3-31.

Figure 3-31 shows the cloud formations normally found with wave development and illustrates schematically the airflow in a similar situation. The illustration shows that the air flows fairly smoothly with a lifting component as it moves along the windward side of the mountain. The wind speed gradually increases, reaching a maximum near the summit. On passing the crest, the flow breaks down into a much more complicated pattern with downdrafts predominating. An indication of the possible intensities can be gained from verified records of sustained downdrafts (and also updrafts) of at least 5,000 feet per minute with other reports showing drafts well in excess of this figure. Turbulence in varying degrees can be expected and is particularly severe in the lower levels; however, it can extend to the tropopause to a lesser degree. Proceeding downwind, some 5 to 10 miles from the summit, the airflow begins to ascend in a definite wave pattern. Additional waves, generally less intense than the primary wave, may form downwind (in some cases six or more have been reported). These are similar to the series of ripples that form downstream from a submerged rock in a swiftly flowing river. The distance between successive waves usually ranges from 2 to 10 miles, depending largely on the existing wind speed and the atmospheric stability. However, wavelengths up to 20 miles have been reported.

It is important to know how to identify a wave situation. Pilots must be briefed on this condition so they can avoid the wave hazards. Characteristic cloud forms peculiar to wave action provide the best means of visual identification. The lenticular (lens shaped) clouds in the upper right of figure 3-31 are smooth in contour. These clouds may occur singly or in layers at heights usually above 20,000 feet, and may be quite ragged when the airflow at that level is turbulent. The roll cloud (also called rotor cloud) forms at a lower level, generally near the height of the mountain ridge, and can be seen extending across the center of the figure. The cap cloud, shown partially covering the mountain slope, must always be avoided in flight because of turbulence, concealed mountain peaks, and strong downdrafts on the lee side. The lenticular, like the roll clouds and cap clouds, are stationary, constantly forming on the windward side and dissipating on the lee side of the wave. The actual cloud forms can be a guide to the degree of turbulence. Smooth clouds generally show smoother airflow in or near them with light turbulence. Clouds appearing ragged or irregular indicate more turbulence.

While clouds are generally present to forewarn the presence of wave activity, it is possible for wave action to take place when the air is too dry to form clouds. This makes the problem of identifying and forecasting more difficult.

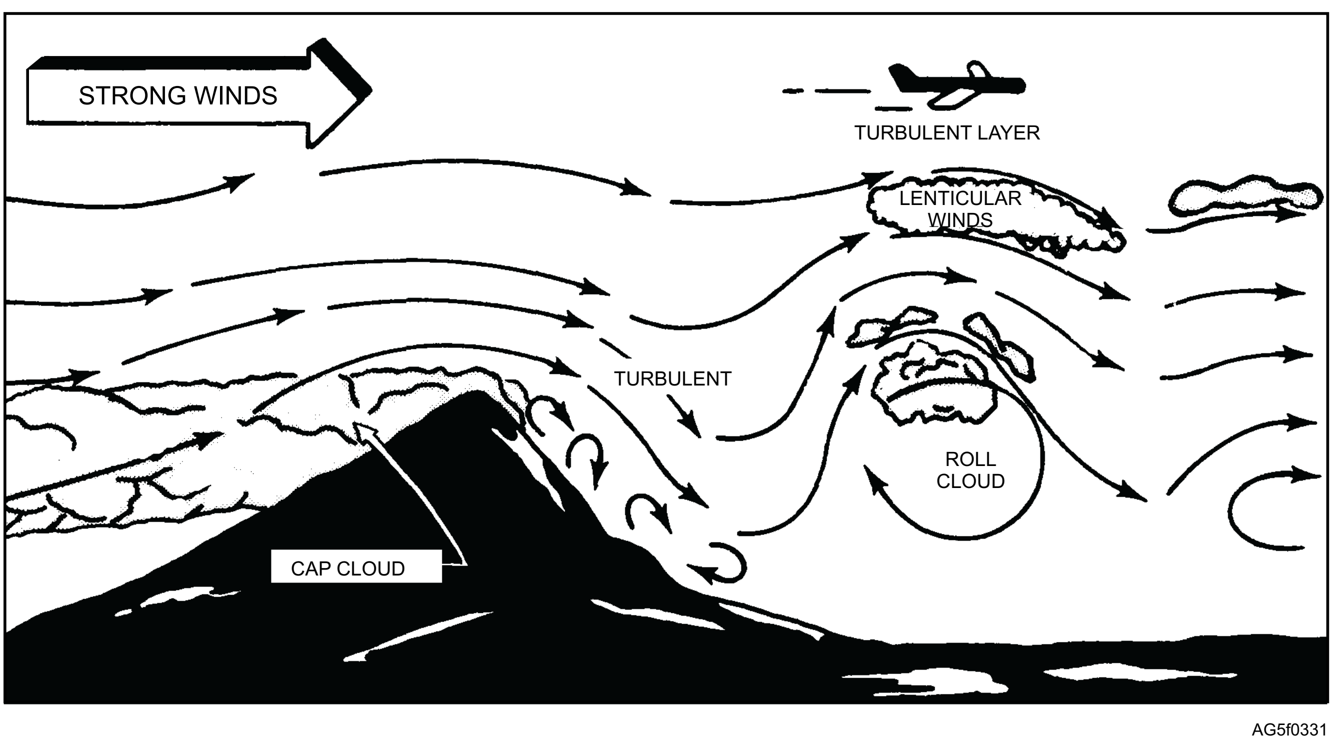

Figure 3-31.—Schematic diagram showing airflow and clouds in a mountain wave.

Foehn Winds

When air flows downhill from a high elevation, its temperature is raised by adiabatic compression. Foehn winds are katabatic winds caused by adiabatic heating of air as it descends on the lee sides of mountains. Foehn winds occur frequently in our western mountain states and in Europe in the late fall and winter. In Montana and Wyoming, the Chinook is a well-known phenomenon; in southern California, the Santa Ana is known particularly for its high-speed winds that easily exceed 50 knots. For the purpose of illustrating a Foehn wind, the Santa Ana is used.

The condition producing the Foehn wind is a high-pressure area with a strong pressure gradient situated near Salt Lake City, Utah. This gradient directs the wind flow into a valley leading to the town of Santa Ana near the coast of California. As the wind enters the valley, its flow is sharply restricted by the funneling effect of the mountainsides. This restriction causes the wind speed to increase, bringing about a drop in pressure in and near the valley. The Bernoulli effect causes this pressure drop in and near a valley.

Generally speaking, when the Santa Ana blows through the Santa Ana Canyon, a similar wind simultaneously affects the entire southern California area. Thus, when meteorological conditions are favorable, this dry northeast wind blows through the many passes and canyons, over all the mountainous area, including the highest peaks, and quite often at exposed places along the entire coast from Santa Barbara to San Diego. Therefore, the term Santa Ana refers to the general condition of a dry northeast wind over southern California.

In the Rocky Mountain states, the onset of Foehn winds have accounted for temperature rises of 50°F or more in only a few minutes. In southern California, the temperature, though less dramatically, also rises rapidly and is accompanied by a rapid decrease in humidity (to 20 percent or less) and a strong shift and increase in wind speeds. Although these winds may on occasion reach destructive velocities, one beneficial aspect is that these winds quickly disperse the severe air pollutants that plague the Los Angeles Basin.

REVIEW QUESTIONS

Q3-15. What is the cause of monsoon winds?

Q3-16. What causes land and sea breezes?

Q3-17. Describe Bernoulli's theorem.

Q3-18. When does a valley breeze usually reach its maximum?

Q3-19. What causes eddies?

Q3-20. What causes Foehn winds?

SUMMARY

In this chapter, we studied the primary, secondary and tertiary circulation of the atmosphere. We learned about large-scale circulations, worldwide locations of major pressure systems, horizontal and vertical pressure systems. We studied how pressure systems, temperature, and world winds relate to each other, and finally we studied small-scale effects, due to local features. A good understanding of atmospheric circulation is essential in order to understand the characteristics of air masses and fronts.