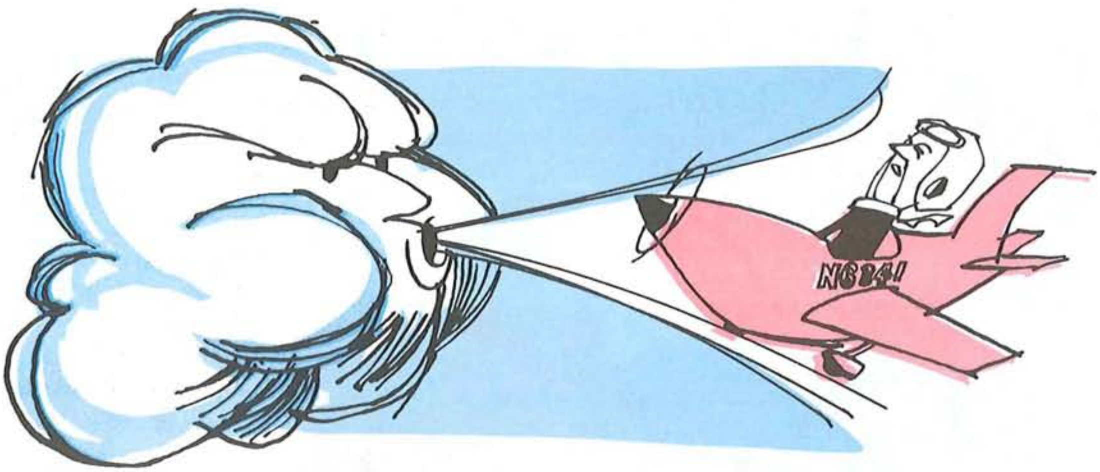

Chapter 4
WIND

Differences in temperature create differences in pressure. These pressure differences drive a complex system of winds in a never ending attempt to reach equilibrium. Wind also transports water vapor and spreads fog, clouds, and precipitation. To help you relate wind to pressure patterns and the movement of weather systems, this chapter explains convection and the pressure gradient force, describes the effects of the Coriolis and frictional forces, relates convection and these forces to the general circulation, discusses local and small-scale wind systems, introduces you to wind shear, and associates wind with weather.

CONVECTION

When two surfaces are heated unequally, they heat the overlying air unevenly. The warmer* air expands and becomes lighter or less dense than the cool* air. The more dense, cool air is drawn to the ground by its greater gravitational force lifting or forcing the warm air upward much as oil is forced to the top of water when the two are mixed. Figure 18 shows the convective process. The rising air spreads and cools, eventually descending to com-

*Frequently throughout this book, we refer to air as *warm, cool,* or *cold.* These terms refer to relative temperatures and not to any fixed temperature reference or to temperatures as they may affect our comfort. For example, compare air at $-10°$ F to air at $0°$ F; relative to each other, the $-10°$ F air is *cool* and the $0°$ F, *warm.* $90°$ F would be *cool* or *cold* relative to $100°$ F.

plete the convective circulation. As long as the uneven heating persists, convection maintains a continuous "convective current."

The horizontal air flow in a convective current is "wind." Convection of both large and small scales accounts for systems ranging from hemispheric circulations down to local eddies. This horizontal flow, wind, is sometimes called "advection." However, the term "advection" more commonly applies to the transport of atmospheric properties by the wind, i.e., warm advection; cold advection; advection of water vapor, etc.

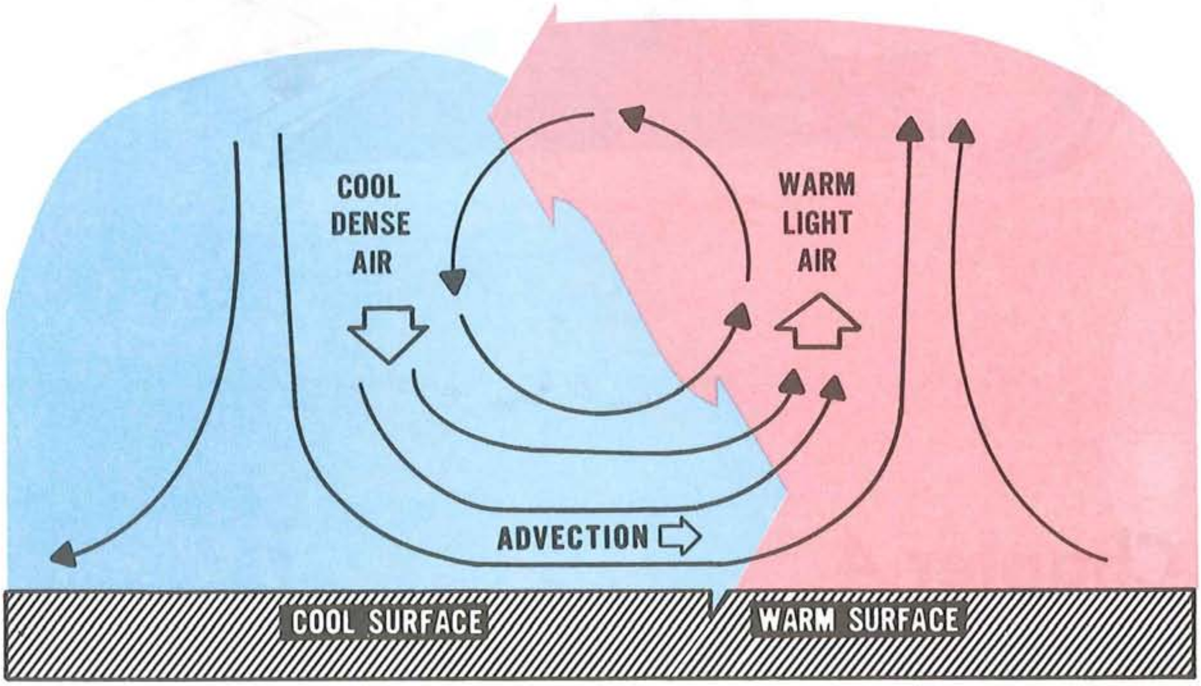

FIGURE 18. Convective current resulting from uneven heating of air by contrasting surface temperatures. The cool, heavier air forces the warmer air aloft establishing a convective cell. Convection continues as long as the uneven heating persists.

PRESSURE GRADIENT FORCE

Pressure differences must create a force in order to drive the wind. This force is the *pressure gradient force*. The force is from higher pressure to lower pressure and is perpendicular to isobars or contours. Whenever a pressure difference develops over an area, the pressure gradient force begins moving the air directly across the isobars. The closer the spacing of isobars, the stronger is the pressure gradient force. The stronger the pressure gradient force, the stronger is the wind. Thus, closely spaced isobars mean strong winds; widely spaced isobars mean lighter wind. From a pressure analysis, you can get a general idea of wind speed from contour or isobar spacing.

Because of uneven heating of the Earth, surface pressure is low in warm equatorial regions and high in cold polar regions. A pressure gradient develops from the poles to the Equator. If the Earth did not rotate, this pressure gradient force would be the only force acting on the wind. Circulation would be two giant hemispheric convective currents as shown in figure 19. Cold air would sink at the poles; wind would blow straight from the poles to the Equator; warm air at the Equator would be forced upward; and high level winds would blow directly toward the poles. However, the Earth does rotate; and because of its rotation, this simple circulation is greatly distorted.

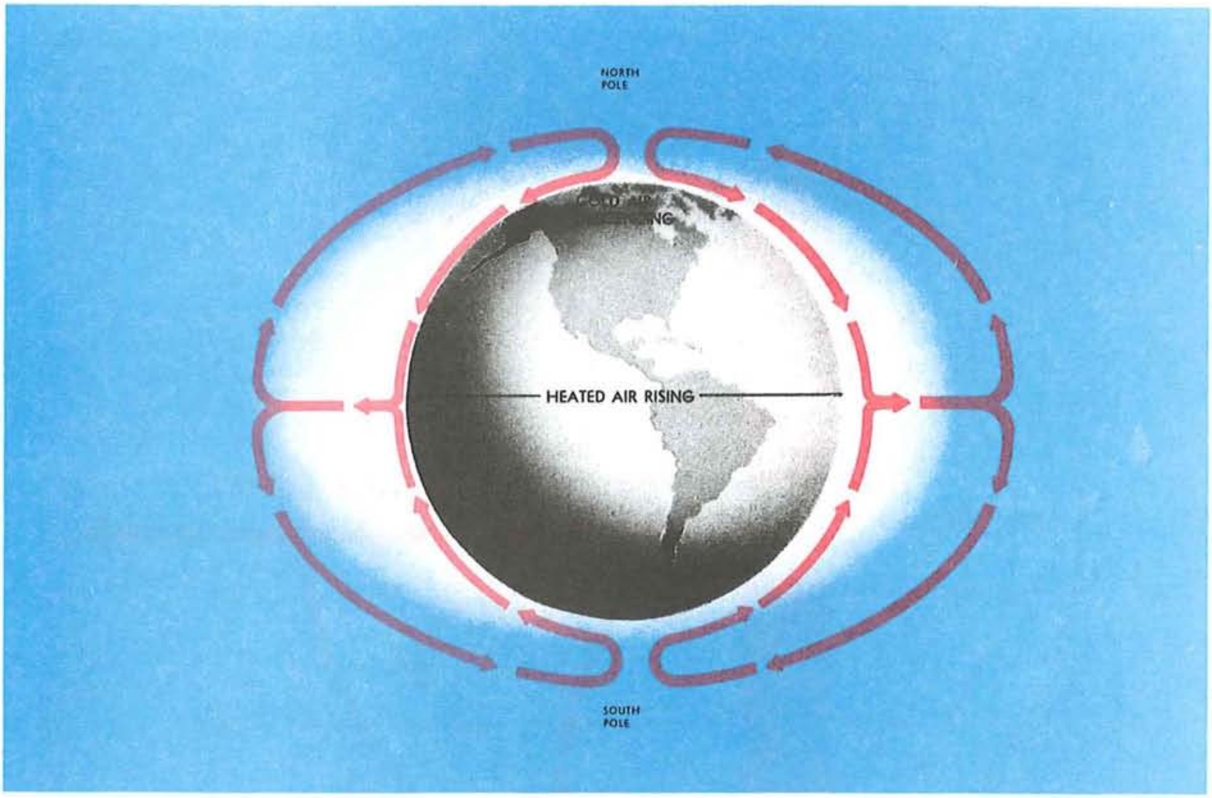

FIGURE 19. Circulation as it would be on a nonrotating globe. Intense heating at the Equator lowers the density. More dense air flows from the poles toward the Equator forcing the less dense air aloft where it flows toward the poles. The circulation would be two giant hemispherical convective currents.

CORIOLIS FORCE

A moving mass travels in a straight line until acted on by some outside force. However, if one views the moving mass from a rotating platform, the path of the moving mass relative to his platform appears to be deflected or curved. To illustrate, start rotating the turntable of a record player. Then using a piece of chalk and a ruler, draw a "straight" line from the center to the outer edge of the turntable. To you, the chalk traveled in a straight line. Now stop the turntable; on it, the line spirals outward from the center as shown in figure 20. To a viewer on the turntable, some "apparent" force deflected the chalk to the right.

A similar apparent force deflects moving particles on the earth. Because the Earth is spherical, the deflective force is much more complex than the simple turntable example. Although the force is termed "apparent," to us on Earth, it is very real. The principle was first explained by a Frenchman, Coriolis, and carries his name—the Coriolis force.

The Coriolis force affects the paths of aircraft; missiles; flying birds; ocean currents; and, most important to the study of weather, air currents. The force deflects air to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. This book concentrates mostly on deflection to the right in the Northern Hemisphere.

Coriolis force is at a right angle to wind direction and directly proportional to wind speed. That is, as wind speed increases, Coriolis force increases. At a given latitude, double the wind speed and you double the Coriolis force. Why at a given latitude?

Coriolis force varies with latitude from zero at the Equator to a maximum at the poles. It influences wind direction everywhere except immediately at the Equator; but the effects are more pronounced in middle and high latitudes.

Remember that the pressure gradient force drives the wind and is perpendicular to isobars. When a pressure gradient force is first established, wind be-

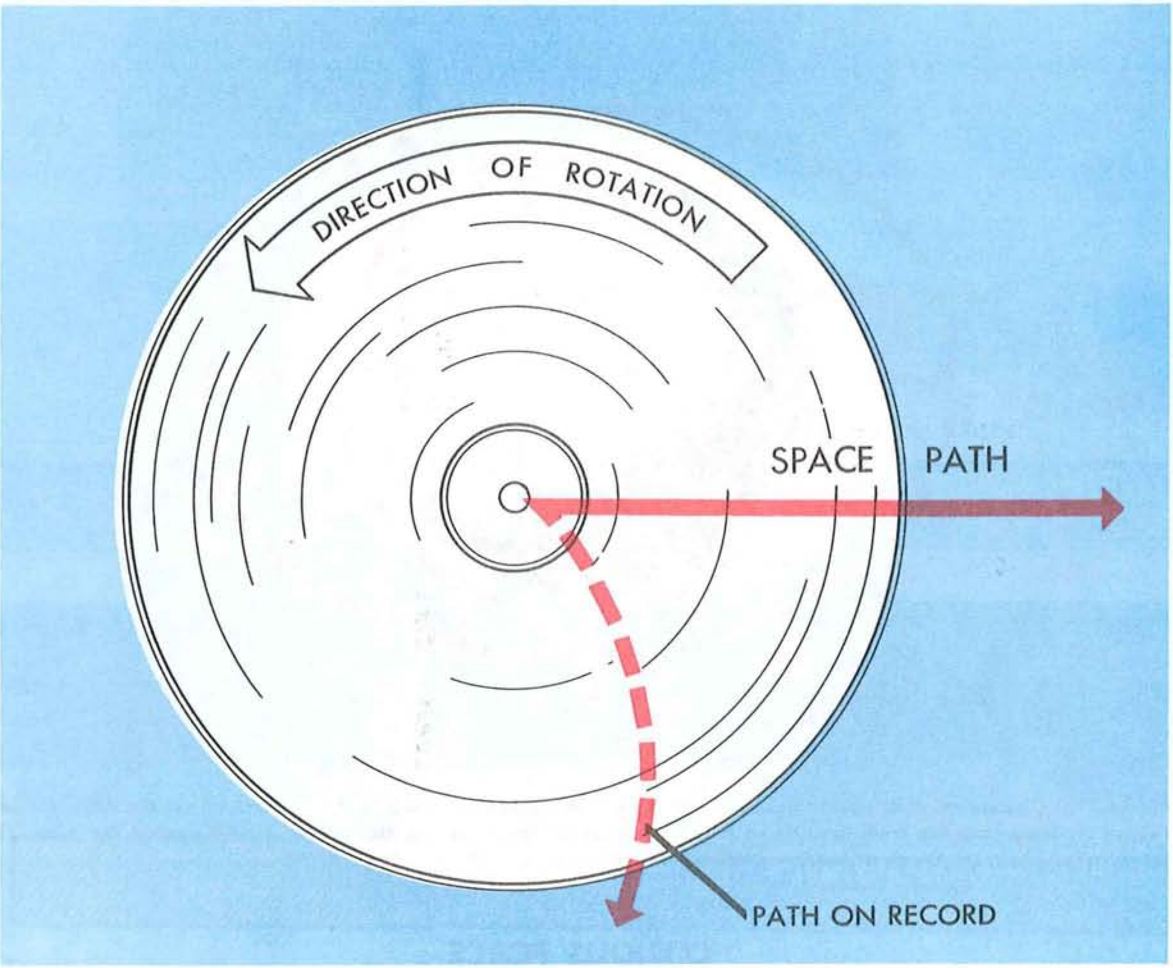

FIGURE 20. Apparent deflective force due to rotation of a horizontal platform. The "space path" is the path taken by a piece of chalk. The "path on the record" is the line traced on the rotating record. Relative to the record, the chalk appeared to curve; in space, it traveled in a straight line.

gins to blow from higher to lower pressure directly across the isobars. However, the instant air begins moving, Coriolis force deflects it to the right. Soon the wind is deflected a full 90° and is parallel to the isobars or contours. At this time, Coriolis force exactly balances pressure gradient force as shown in figure 21. With the forces in balance, wind will remain parallel to isobars or contours. Surface friction disrupts this balance as we discuss later; but first let's see how Coriolis force distorts the fictitious global circulation shown in figure 19.

THE GENERAL CIRCULATION

As air is forced aloft at the Equator and begins its high-level trek northward, the Coriolis force turns it to the right or to the east as shown in figure 22. Wind becomes westerly at about 30° latitude temporarily blocking further northward movement. Similarly, as air over the poles begins its low-level journey southward toward the Equator, it likewise is deflected to the right and becomes an east wind, halting for a while its southerly progress—also shown in figure 22. As a result, air literally "piles up" at about 30° and 60° latitude in both hemispheres. The added weight of the air in-

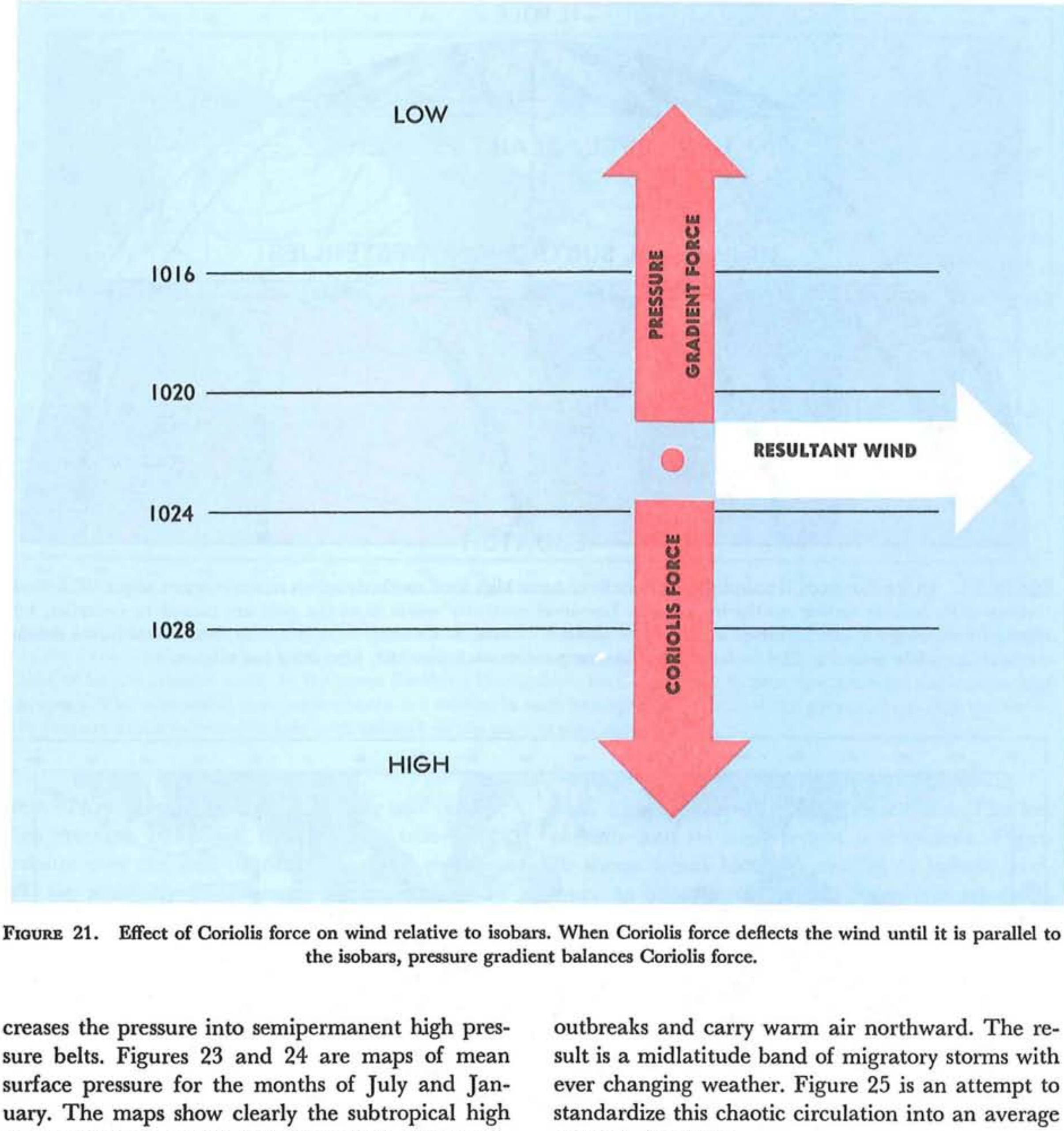

FIGURE 21. Effect of Coriolis force on wind relative to isobars. When Coriolis force deflects the wind until it is parallel to the isobars, pressure gradient balances Coriolis force.

creases the pressure into semipermanent high pressure belts. Figures 23 and 24 are maps of mean surface pressure for the months of July and January. The maps show clearly the subtropical high pressure belts near 30° latitude in both the Northern and Southern Hemispheres.

The building of these high pressure belts creates a temporary impasse disrupting the simple convective transfer between the Equator and the poles. The restless atmosphere cannot live with this impasse in its effort to reach equilibrium. Something has to give. Huge masses of air begin overturning in middle latitudes to complete the exchange.

Large masses of cold air break through the northern barrier plunging southward toward the Tropics. Large midlatitude storms develop between cold outbreaks and carry warm air northward. The result is a midlatitude band of migratory storms with ever changing weather. Figure 25 is an attempt to standardize this chaotic circulation into an average general circulation.

Since pressure differences cause wind, seasonal pressure variations determine to a great extent the areas of these cold air outbreaks and midlatitude storms. But, seasonal pressure variations are largely due to seasonal temperature changes. We have learned that, at the surface, warm temperatures to a great extent determine low pressure and cold temperatures, high pressure. We have also learned that seasonal temperature changes over continents are much greater than over oceans.

During summer, warm continents tend to be

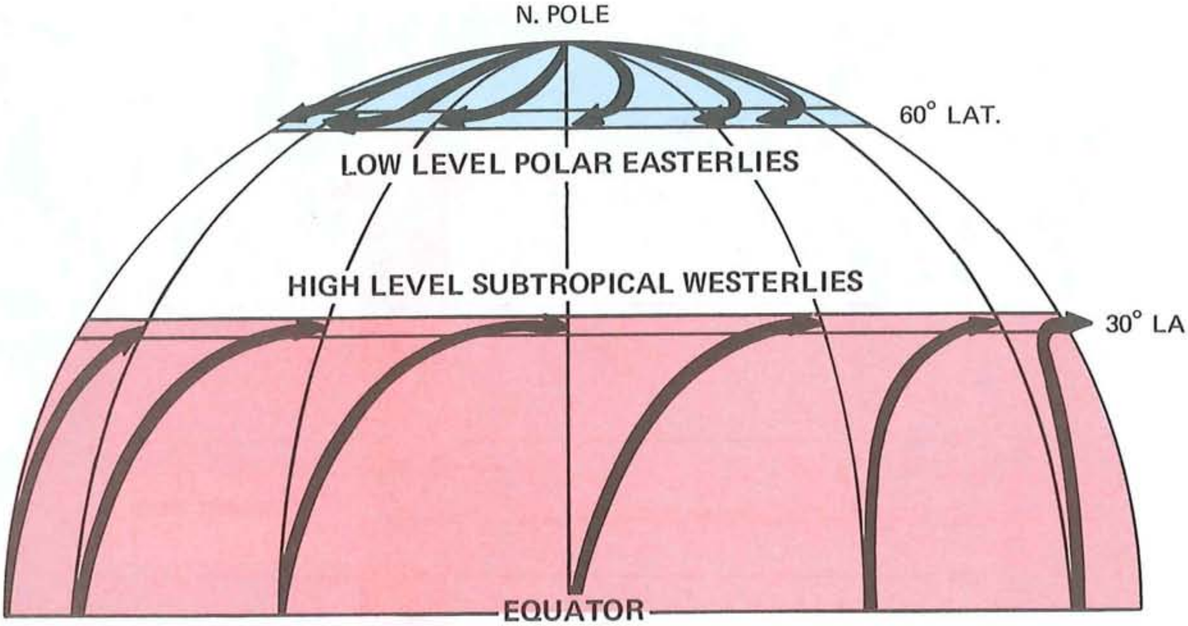

FIGURE 22. In the Northern Hemisphere, Coriolis force turns high level southerly winds to westerlies at about 30° latitude, temporarily halting further northerly progress. Low-level northerly winds from the pole are turned to easterlies, temporarily stopping further southward movement at about 60° latitude. Air tends to "pile up" at these two latitudes creating a void in middle latitudes. The restless atmosphere cannot live with this void; something has to give.

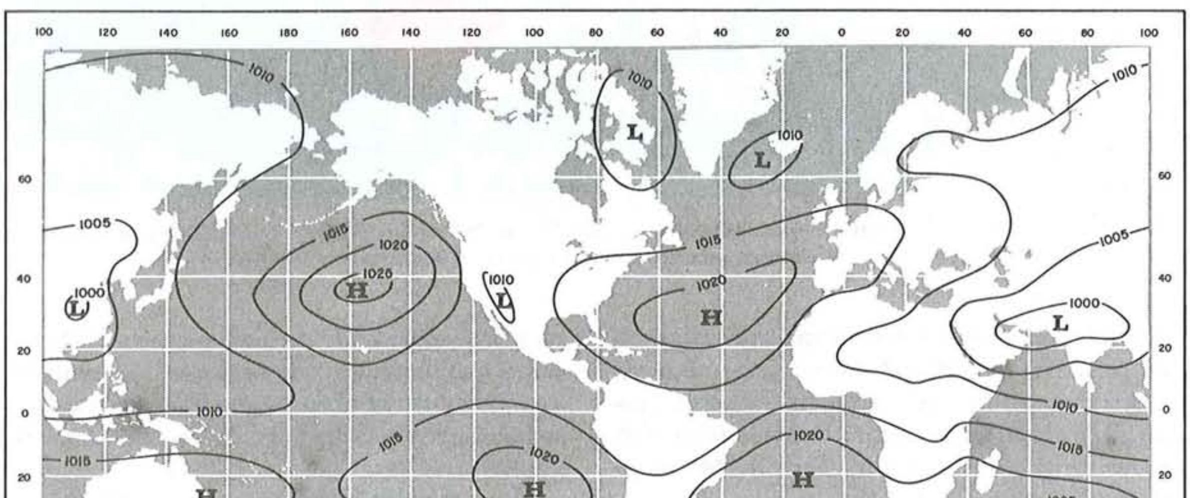

FIGURE 23. Mean world-wide surface pressure distribution in July. In the warm Northern Hemisphere, warm land areas tend to have low pressure, and cool oceanic areas tend to have high pressure. In the cool Southern Hemisphere, the pattern is reversed; cool land areas tend to have high pressure; and water surfaces, low pressure. However, the relationship is not so evident in the Southern Hemisphere because of relatively small amounts of land. The subtropical high pressure belts are clearly evident at about 30° latitude in both hemispheres.

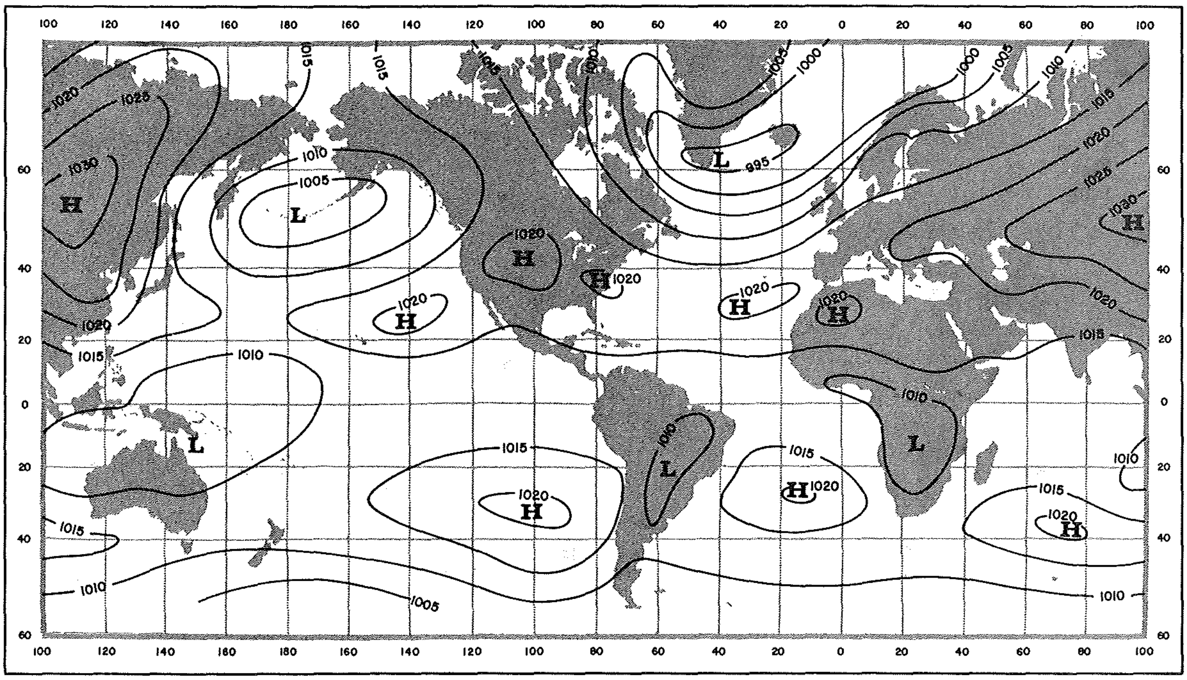

FIGURE 24. Mean world-wide surface pressure distribution in January. In this season, the pattern in figure 23 is reversed. In the cool Northern Hemisphere, cold continental areas are predominantly areas of high pressure while warm oceans tend to be low pressure areas. In the warm Southern Hemisphere, land areas tend to have low pressure; and oceans, high pressure. The subtropical high pressure belts are evident in both hemispheres. Note that the pressure belts shift southward in January and northward in July with the shift in the zone of maximum heating.

areas of low pressure and the relatively cool oceans, high pressure. In winter, the reverse is true—high pressure over the cold continents and low pressure over the relatively warm oceans. Figures 23 and 24 show this seasonal pressure reversal. The same pressure variations occur in the warm and cold seasons of the Southern Hemisphere, although the effect is not as pronounced because of the much larger water areas of the Southern Hemisphere.

Cold outbreaks are strongest in the cold season and are predominantly from cold continental areas. Summer outbreaks are weaker and more likely to originate from cool water surfaces. Since these outbreaks are masses of cool, dense air, they characteristically are high pressure areas.

As the air tries to blow outward from the high pressure, it is deflected to the right by the Coriolis force. Thus, the wind around a high blows clockwise. The high pressure with its associated wind system is an *anticyclone*.

The storms that develop between high pressure systems are characterized by low pressure. As winds try to blow inward toward the center of low pressure, they also are deflected to the right. Thus, the wind around a low is counterclockwise. The low pressure and its wind system is a *cyclone*. Figure 26 shows winds blowing parallel to isobars (contours on upper level charts). The winds are clockwise around highs and counterclockwise around lows.

The high pressure belt at about 30° north latitude forces air outward at the surface to the north and to the south. The northbound air becomes entrained into the midlatitude storms. The southward moving air is again deflected by the Coriolis force becoming the well-known subtropical northeast trade winds. In midlatitudes, high level winds are predominantly from the west and are known as the prevailing westerlies. Polar easterlies dominate low-level circulation north of about 60° latitude.

These three major wind belts are shown in figure 25. Northeasterly trade winds carry tropical storms from east to west. The prevailing westerlies drive midlatitude storms generally from west to east. Few major storm systems develop in the comparatively small Arctic region; the chief influence of the polar easterlies is their contribution to the development of midlatitude storms.

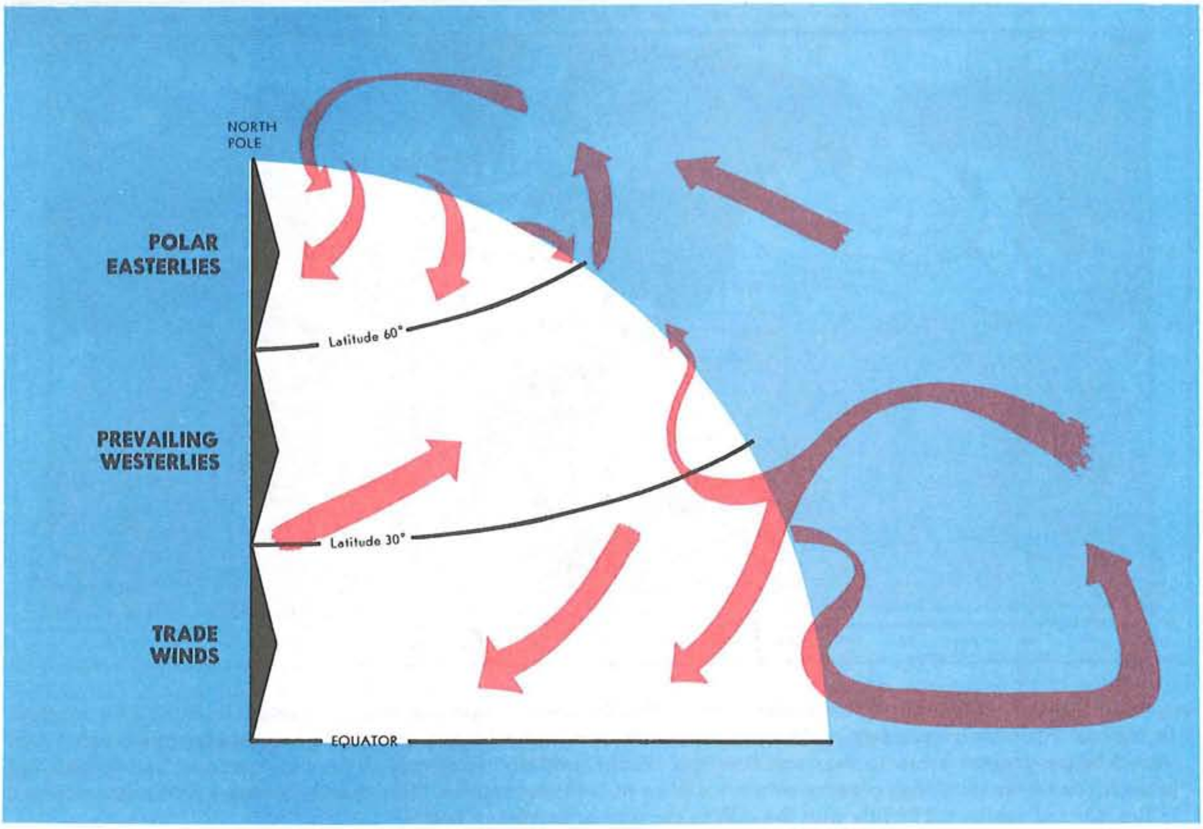

FIGURE 25. General average circulation in the Northern Hemisphere. Note the three belts of prevailing winds, the polar easterlies, the prevailing westerlies in middle latitudes, and the northeasterly "trade" winds. The belt of prevailing westerlies is a mixing zone between the North Pole and the Equator characterized by migrating storms.

Our discussion so far has said nothing about friction. Wind flow patterns aloft follow isobars or contours where friction has little effect. We cannot, however, neglect friction near the surface.

FRICTION

Friction between the wind and the terrain surface slows the wind. The rougher the terrain, the greater is the frictional effect. Also, the stronger the wind speed, the greater is the friction. One may not think of friction as a force, but it is a very real and effective force always acting opposite to wind direction.

As frictional force slows the windspeed, Coriolis force decreases. However, friction does not affect pressure gradient force. Pressure gradient and Coriolis forces are no longer in balance. The stronger pressure gradient force turns the wind at an angle across the isobars toward lower pressure until the three forces balance as shown in figure 27. Frictional and Coriolis forces combine to just balance pressure gradient force. Figure 28 shows how surface wind spirals outward from high pressure into low pressure crossing isobars at an angle.

The angle of surface wind to isobars is about 10° over water increasing with roughness of terrain. In mountainous regions, one often has difficulty relating surface wind to pressure gradient because of immense friction and also because of local terrain effects on pressure.

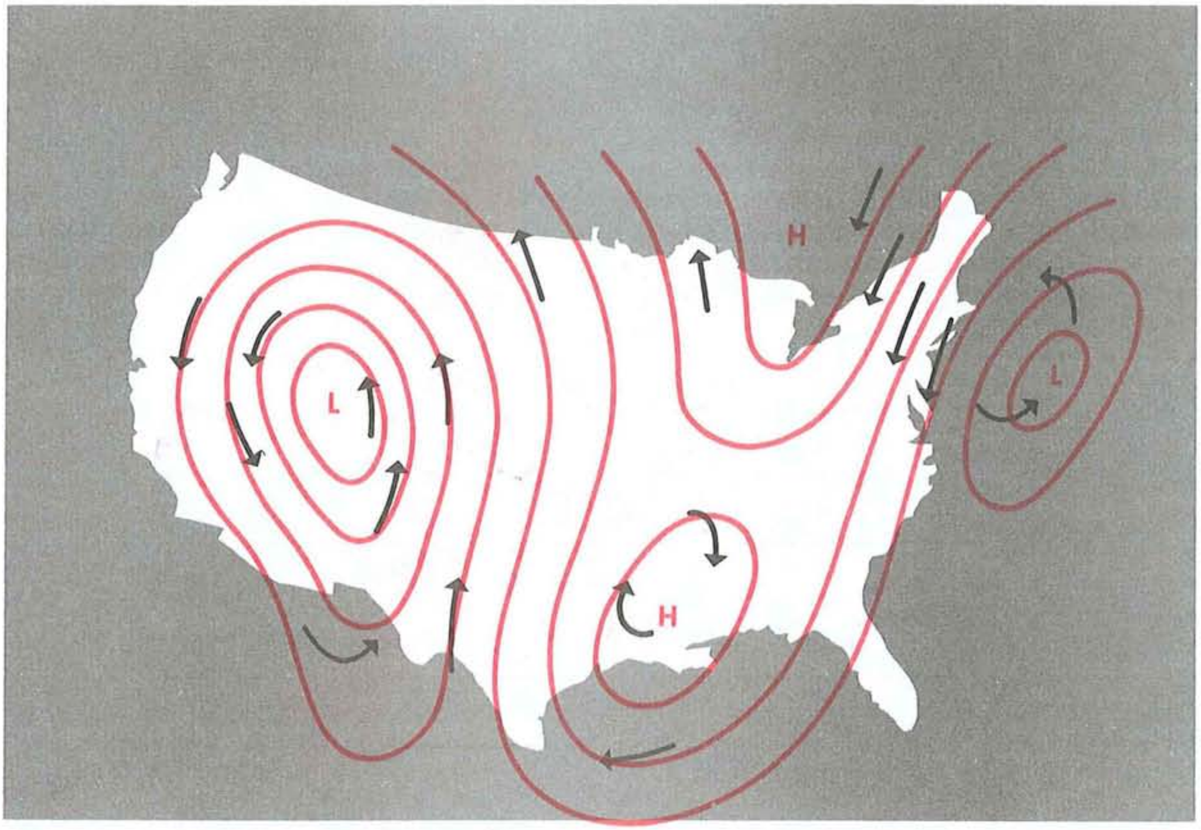

FIGURE 26. Air flow around pressure systems above the friction layer. Wind (black arrows) is parallel to contours and circulates clockwise around high pressure and counterclockwise around low pressure.

THE JET STREAM

A discussion of the general circulation is incomplete when it does not mention the "jet stream." Winds on the average increase with height throughout the troposphere culminating in a maximum near the level of the tropopause. These maximum winds tend to be further concentrated in narrow bands. A jet stream, then, is a narrow band of strong winds meandering through the atmosphere at a level near the tropopause. Since it is of interest primarily to high level flight, further discussion of the jet stream is reserved for chapter 13, "High Altitude Weather."

LOCAL AND SMALL SCALE WINDS

Until now, we have dealt only with the general circulation and major wind systems. Local terrain features such as mountains and shore lines influence local winds and weather.

MOUNTAIN AND VALLEY WINDS

In the daytime, air next to a mountain slope is heated by contact with the ground as it receives radiation from the sun. This air usually becomes warmer than air at the same altitude but farther from the slope.

Colder, denser air in the surroundings settles downward and forces the warmer air near the ground up the mountain slope. This wind is a "valley wind" so called because the air is flowing up out of the valley.

At night, the air in contact with the mountain slope is cooled by terrestrial radiation and becomes heavier than the surrounding air. It sinks along the

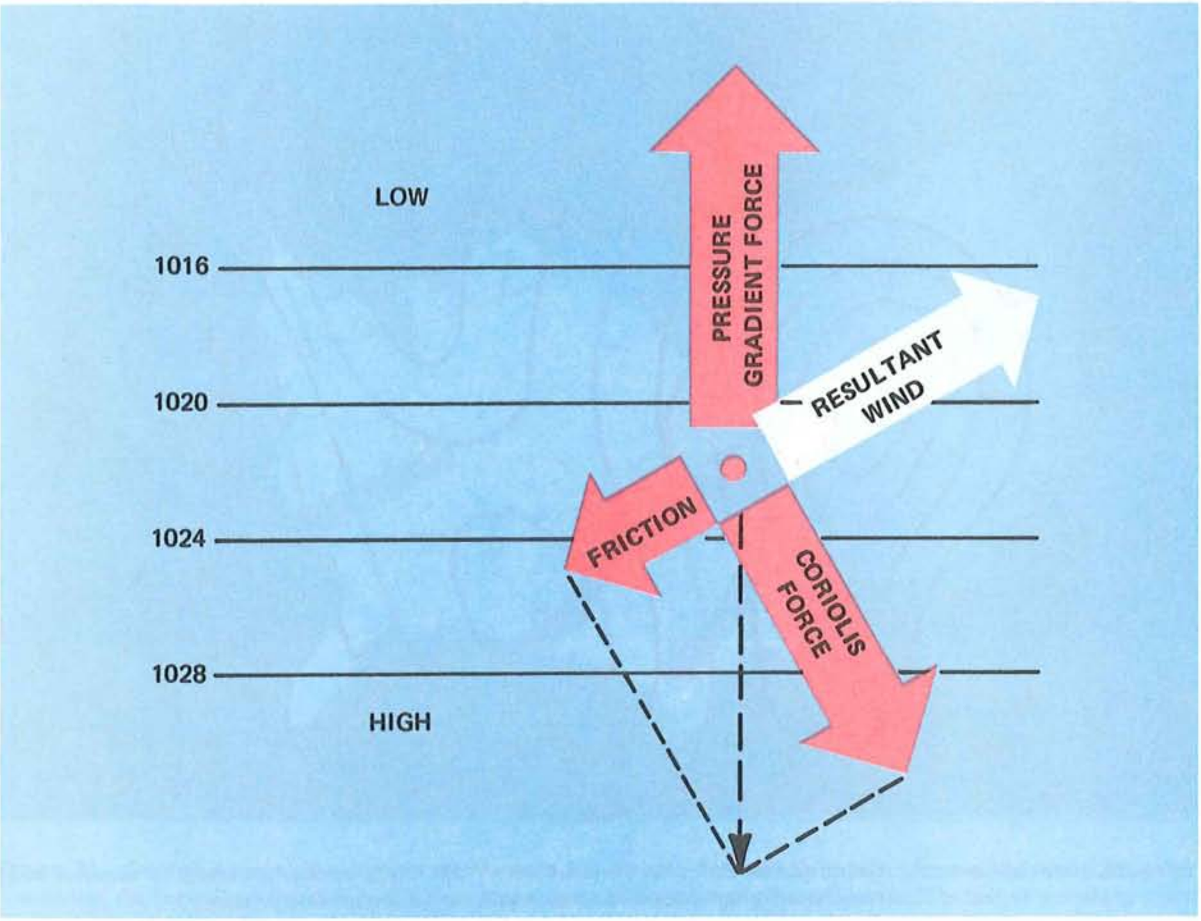

FIGURE 27. Surface friction slows the wind and reduces Coriolis force but does not affect pressure gradient force; winds near the surface are deflected across the isobars toward lower pressure.

slope, producing the "mountain wind" which flows like water down the mountain slope. Mountain winds are usually stronger than valley winds, especially in winter. The mountain wind often continues down the more gentle slopes of canyons and valleys, and in such cases takes the name "drainage wind." It can become quite strong over some terrain conditions and in extreme cases can become hazardous when flowing through canyon restrictions as discussed in chapter 9.

KATABATIC WIND

A katabatic wind is any wind blowing down an incline when the incline is influential in causing the wind. Thus, the mountain wind is a katabatic wind. Any katabatic wind originates because cold, heavy air spills down sloping terrain displacing warmer, less dense air ahead of it. Air is heated and dried as it flows down slope as we will study in later chapters. Sometimes the descending air becomes warmer than the air it replaces.

Many katabatic winds recurring in local areas have been given colorful names to highlight their dramatic, local effect. Some of these are the Bora, a cold northerly wind blowing from the Alps to the Mediterranean coast; the Chinook, figure 29, a warm wind down the east slope of the Rocky Mountains often reaching hundreds of miles into the high plains; the Taku, a cold wind in Alaska blowing off the Taku glacier; and the Santa Ana, a warm wind descending from the Sierras into the Santa Ana Valley of California.

LAND AND SEA BREEZES

As frequently stated earlier, land surfaces warm and cool more rapidly than do water surfaces; therefore, land is warmer than the sea during the

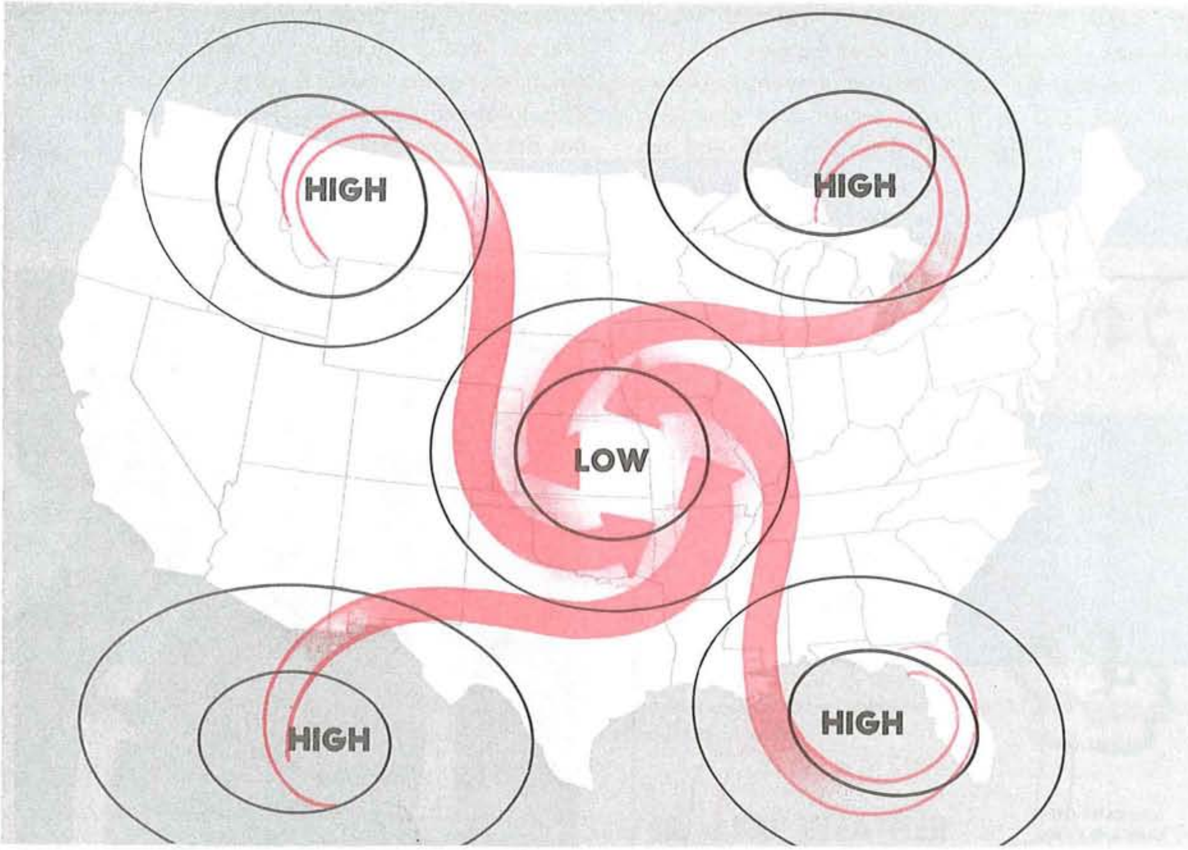

FIGURE 28. Circulation around pressure systems at the surface. Wind spirals outward from high pressure and inward to low pressure, crossing isobars at an angle.

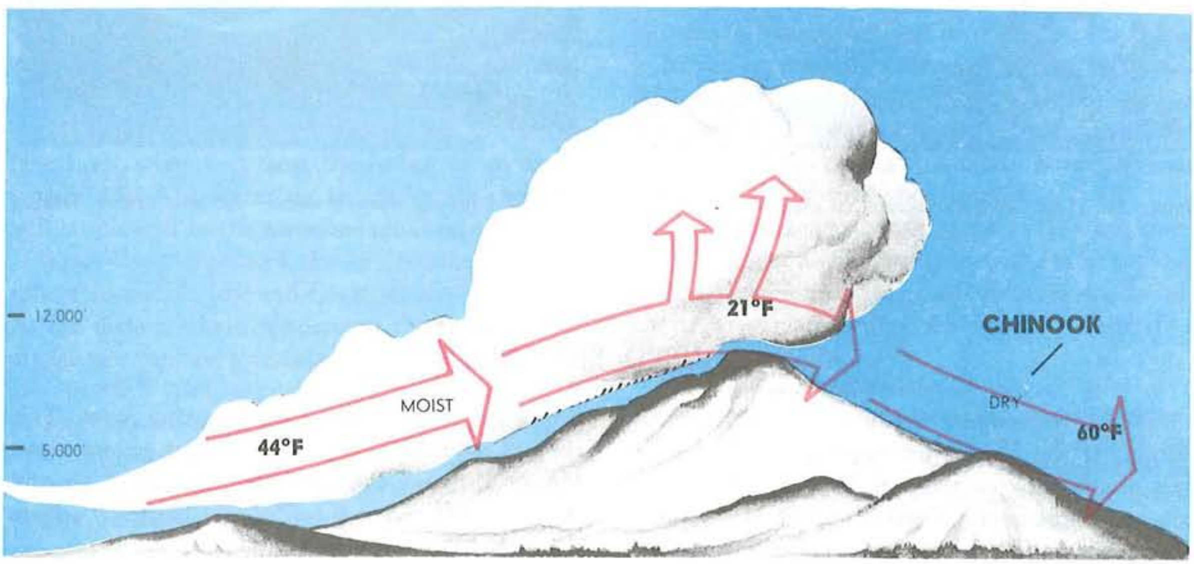

FIGURE 29. The "Chinook" is a katabatic (downslope) wind. Air cools as it moves upslope and warms as it blows downslope. The Chinook occasionally produces dramatic warming over the plains just east of the Rocky Mountains.

day; wind blows from the cool water to warm land—the "sea breeze" so called because it blows from the sea. At night, the wind reverses, blows from cool land to warmer water, and creates a "land breeze." Figure 30 diagrams land and sea breezes.

Land and sea breezes develop only when the overall pressure gradient is weak. Wind with a stronger pressure gradient mixes the air so rapidly that local temperature and pressure gradients do not develop along the shore line.

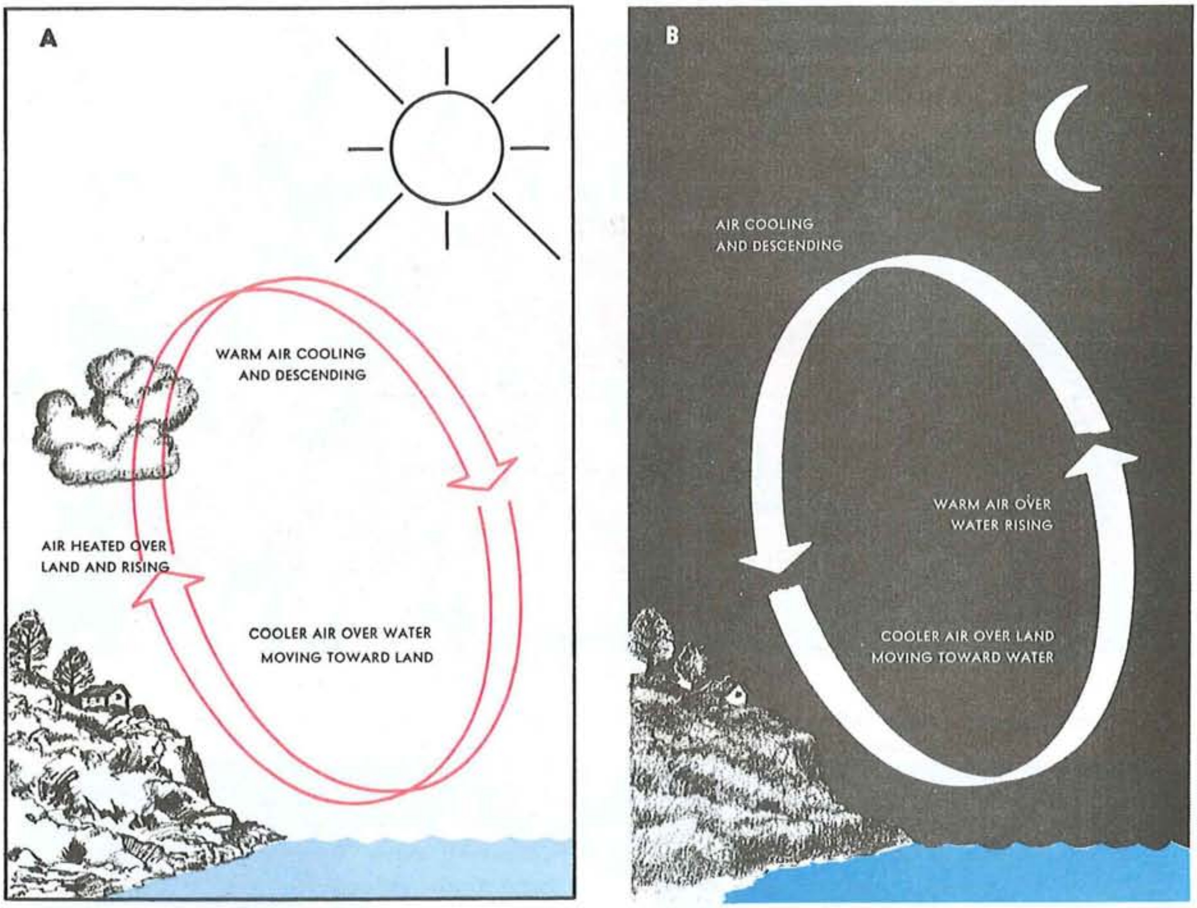

FIGURE 30. Land and sea breezes. At night, cool air from the land flows toward warmer water—the land breeze. During the day, wind blows from the water to the warmer land—the sea breeze.

WIND SHEAR

Rubbing two objects against each other creates friction. If the objects are solid, no exchange of mass occurs between the two. However, if the objects are fluid currents, friction creates eddies along a common shallow mixing zone, and a mass transfer takes place in the shallow mixing layer. This zone of induced eddies and mixing is a shear zone. Figure 31 shows two adjacent currents of air and their accompanying shear zone. Chapter 9 relates wind shear to turbulence.

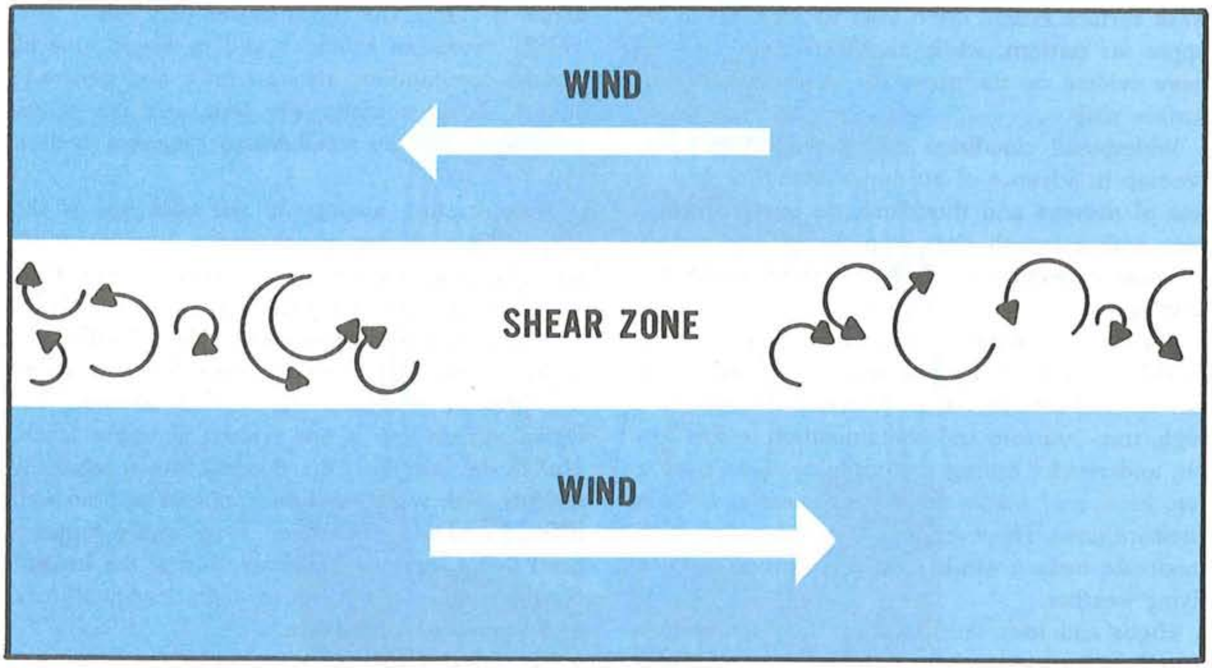

FIGURE 31. Wind shear. Air currents of differing velocities create friction or "shear" between them. Mixing in the shear zone results in a snarl of eddies and whirls.

WIND, PRESSURE SYSTEMS, AND WEATHER

We already have shown that wind speed is proportional to the spacing of isobars or contours on a weather map. However, with the same spacing, wind speed at the surface will be less than aloft because of surface friction.

You also can determine wind direction from a weather map. If you face along an isobar or contour with lower pressure on your left, wind will be blowing in the direction you are facing. On a surface map, wind will cross the isobar at an angle toward lower pressure; on an upper air chart, it will be parallel to the contour.

Wind blows counterclockwise (Northern Hemisphere) around a low and clockwise around a high. At the surface where winds cross the isobars at an angle, you can see a transport of air from high to low pressure. Although winds are virtually parallel to contours on an upper air chart, there still is a slow transport of air from high to low pressure.

At the surface when air converges into a low, it cannot go outward against the pressure gradient, nor can it go downward into the ground; it must go upward.* Therefore, a low or trough is an area of rising air.

Rising air is conducive to cloudiness and precipitation; thus we have the general association of low pressure–bad weather. Reasons for the inclement weather are developed in later chapters.

By similar reasoning, air moving out of a high or ridge depletes the quantity of air. Highs and ridges, therefore, are areas of descending air. Descending air favors dissipation of cloudiness; hence the association, high pressure–good weather.

Many times weather is more closely associated with an upper air pattern than with features shown by the surface map. Although features on the two charts are related, they seldom are identical. A

*You may recall that earlier we said air "piles up" in the vicinity of 30° latitude increasing pressure and forming the subtropical high pressure belt. Why, then, does not air flowing into a low or trough increase pressure and fill the system? Dynamic forces maintain the low or trough; and these forces differ from the forces that maintain the subtropical high.

weak surface system often loses its identity in the upper air pattern, while another system may be more evident on the upper air chart than on the surface map.

Widespread cloudiness and precipitation often develop in advance of an upper trough or low. A line of showers and thunderstorms is not uncommon with a trough aloft even though the surface pressure pattern shows little or no cause for the development.

On the other hand, downward motion in a high or ridge places a "cap" on convection, preventing any upward motion. Air may become stagnant in a high, trap moisture and contamination in low levels, and restrict ceiling and visibility. Low stratus, fog, haze, and smoke are not uncommon in high pressure areas. However, a high or ridge aloft with moderate surface winds most often produces good flying weather.

Highs and lows tend to *lean* from the surface into the upper atmosphere. Due to this slope, winds aloft often blow across the associated surface systems. Upper winds tend to steer surface systems in the general direction of the upper wind flow.

An intense, cold, low pressure vortex *leans less* than does a weaker system. The intense low becomes oriented almost vertically and is clearly evident on both surface and upper air charts. Upper winds encircle the surface low and do not blow across it. Thus, the storm moves very slowly and usually causes an extensive and persistent area of clouds, precipitation, strong winds, and generally adverse flying weather. The term *cold low* sometimes used by the weatherman describes such a system.

A contrasting analogy to the cold low is the *thermal low*. A dry, sunny region becomes quite warm from intense surface heating thus generating a surface low pressure area. The warm air is carried to high levels by convection, but cloudiness is scant because of lack of moisture. Since in warm air, pressure decreases slowly with altitude, the warm surface low is not evident at upper levels. Unlike the cold low, the thermal low is relatively shallow with weak pressure gradients and no well defined cyclonic circulation. It generally supports good flying weather. However, during the heat of the day, one must be alert for high density altitude and convective turbulence.

We have cited three exceptions to the low pressure–bad weather, high pressure–good weather rule: (1) cloudiness and precipitation with an upper air trough or low not evident on the surface chart; (2) the contaminated high; and (3) the thermal low. As this book progresses, you can further relate weather systems more specifically to flight operations.

Chapter 5
MOISTURE, CLOUD FORMATION, AND PRECIPITATION

Imagine, if you can, how easy flying would be if skies everywhere were clear! But, flying isn't always that easy; moisture in the atmosphere creates a variety of hazards unmatched by any other weather element. Within Earth's climatic range, water is in the frozen, liquid, and gaseous states.

WATER VAPOR

Water evaporates into the air and becomes an ever-present but variable constituent of the atmosphere. Water vapor is invisible just as oxygen and other gases are invisible. However, we can readily measure water vapor and express it in different ways. Two commonly used terms are (1) relative humidity, and (2) dew point.

RELATIVE HUMIDITY

Relative humidity routinely is expressed in percent. As the term suggests, *relative humidity* is "relative." It relates *the actual water vapor present to that which could be present.*

Temperature largely determines the maximum amount of water vapor air can hold. As figure 32 shows, warm air can hold more water vapor than cool air. Figure 33 relates water vapor, temperature, and relative humidity. Actually, relative humidity expresses the degree of saturation. Air with 100% relative humidity is saturated; less than 100% is unsaturated.

If a given volume of air is cooled to some specific temperature, it can hold no more water vapor than is actually present, relative humidity becomes 100%, and saturation occurs. What is that temperature?

DEW POINT

Dew point is the temperature to which air must be cooled to become saturated by the water vapor already present in the air. Aviation weather reports normally include the air temperature and dew point temperature. Dew point when related to air temperature reveals qualitatively how close the air is to saturation.

TEMPERATURE–DEW POINT SPREAD

The difference between air temperature and dew point temperature is popularly called the "spread." As spread becomes less, relative humidity increases, and it is 100% when temperature and dew point are the same. Surface temperature–dew point spread is important in anticipating fog but has little bearing on precipitation. To support precipitation, air must be saturated through thick layers aloft.

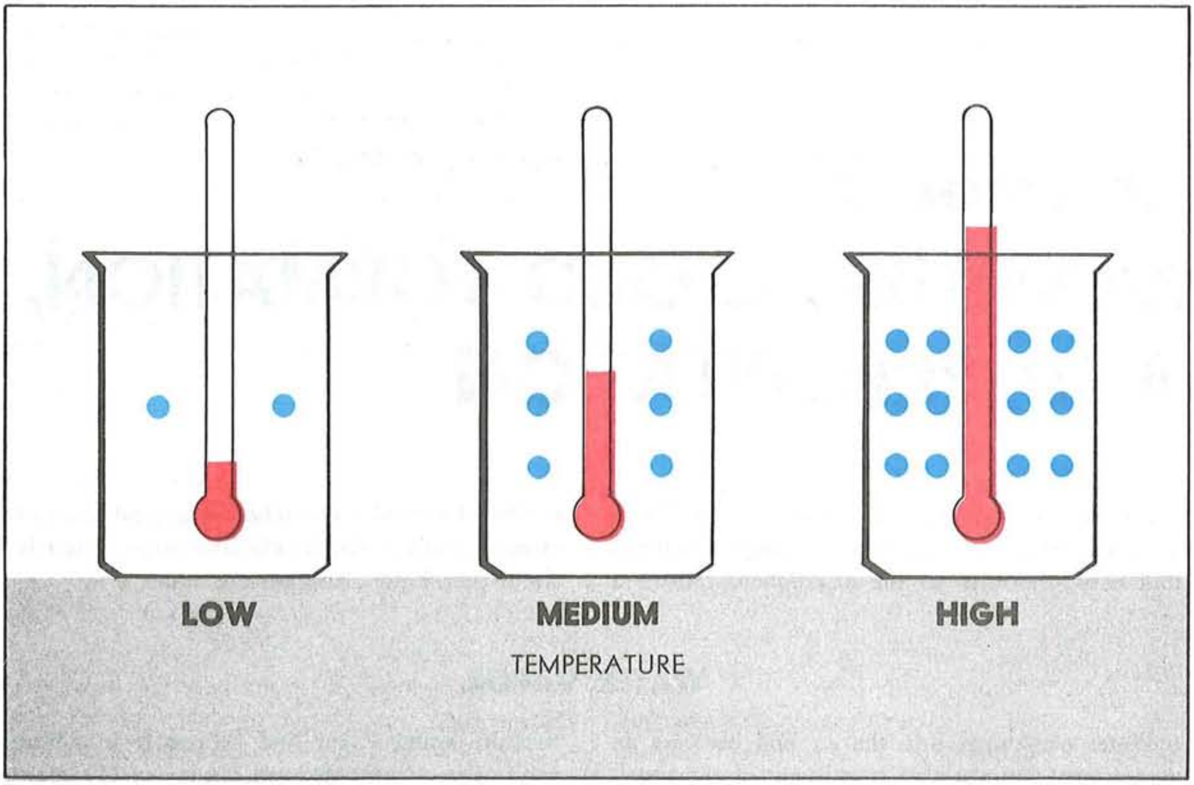

FIGURE 32. Blue dots illustrate the increased water vapor capacity of warmer air. At each temperature, air can hold a specific amount of water vapor—no more.

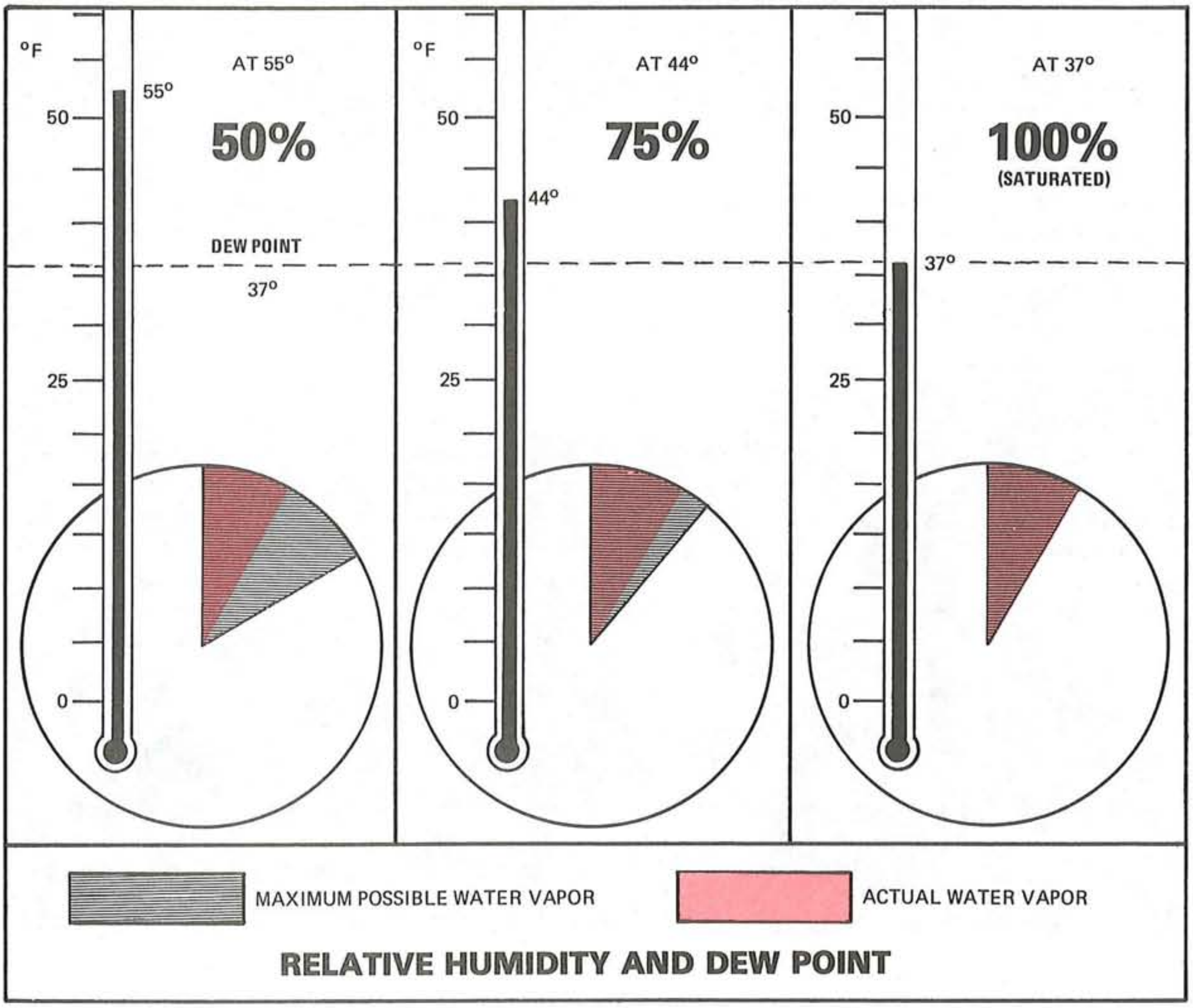

FIGURE 33. Relative humidity depends on both temperature and water vapor. In this figure, water vapor is constant but temperature varies. On the left, relative humidity is 50%; the warmer air could hold twice as much water vapor as is actually present. As the air cools, center and right, relative humidity increases. As the air cools to 37° F, its capacity to hold water vapor is reduced to the amount actually present. Relative humidity is 100% and the air is now "saturated." Note that at 100% humidity, temperature and dew point are the same. The air cooled to saturation, i.e., it cooled to the dew point.

Sometimes the spread at ground level may be quite large, yet at higher altitudes the air is saturated and clouds form. Some rain may reach the ground or it may evaporate as it falls into the drier air. Figure 34 is a photograph of "virga"—streamers of precipitation trailing beneath clouds but evaporating before reaching the ground. Our never ending weather cycle involves a continual reversible change of water from one state to another. Let's take a closer look at change of state.

CHANGE OF STATE

Evaporation, condensation, sublimation, freezing, and melting are changes of state. Evaporation is the changing of liquid water to invisible water vapor. Condensation is the reverse process. Sublimation is the changing of ice directly to water vapor, or water vapor to ice, bypassing the liquid

FIGURE 34. Virga. Precipitation from the cloud evaporates in drier air below and does not reach the ground.

state in each process. Snow or ice crystals result from the sublimation of water vapor directly to the solid state. We are all familiar with freezing and melting processes.

LATENT HEAT

Any change of state involves a heat transaction with no change in temperature. Figure 35 diagrams the heat exchanges between the different states. Evaporation requires heat energy that comes from the nearest available heat source. This heat energy is known as the "latent heat of vaporization," and its removal cools the source it comes from. An example is the cooling of your body by evaporation of perspiration.

What becomes of this heat energy used by evaporation? Energy cannot be created or destroyed, so it is hidden or stored in the invisible water vapor. When the water vapor condenses to liquid water or sublimates directly to ice, energy originally used in the evaporation reappears as heat and is released to the atmosphere. This energy is "latent heat" and is quite significant as we learn in later chapters. Melting and freezing involve the exchange of "latent heat of fusion" in a similar manner. The latent heat of fusion is much less than that of condensation and evaporation; however, each in its own way plays an important role in aviation weather.

As air becomes saturated, water vapor begins to condense on the nearest available surface. What surfaces are in the atmosphere on which water vapor may condense?

CONDENSATION NUCLEI

The atmosphere is never completely clean; an abundance of microscopic solid particles suspended in the air are condensation surfaces. These particles, such as salt, dust, and combustion byproducts

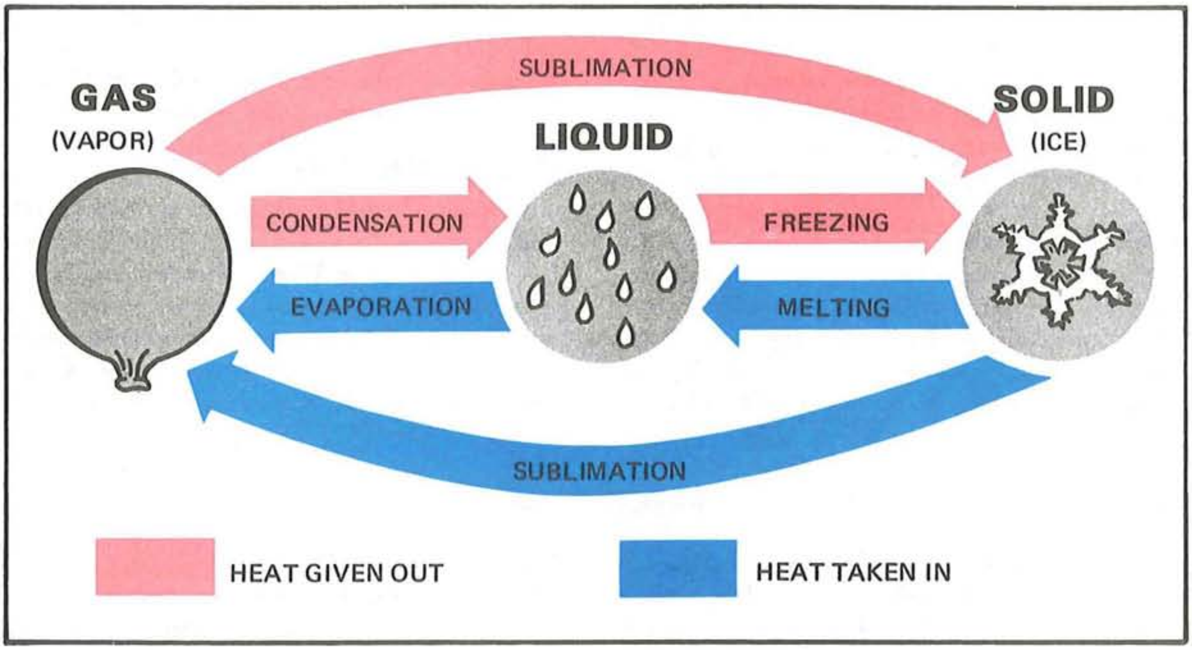

FIGURE 35. Heat transactions when water changes state. Blue arrows indicate changes that absorb heat. The absorbed heat remains hidden, or "latent" until a reverse change occurs. The red arrows show changes that release latent heat back to the surroundings. The heat exchange occurs whenever water changes state even when there is no change in temperature. These heat exchanges play important roles in suppressing temperature changes and in developing instability.

are "condensation nuclei." Some condensation nuclei have an affinity for water and can induce condensation or sublimation even when air is almost but not completely saturated.

As water vapor condenses or sublimates on condensation nuclei, liquid or ice particles begin to grow. Whether the particles are liquid or ice does not depend entirely on temperature. Liquid water may be present at temperatures well below freezing.

SUPERCOOLED WATER

Freezing is complex and liquid water droplets often condense or persist at temperatures colder than 0° C. Water droplets colder than 0° C are supercooled. When they strike an exposed object, the impact induces freezing. Impact freezing of supercooled water can result in aircraft icing.

Supercooled water drops very often are in abundance in clouds at temperatures between 0° C and −15° C with decreasing amounts at colder temperatures. Usually, at temperatures colder than −15° C, sublimation is prevalent; and clouds and fog may be mostly ice crystals with a lesser amount of supercooled water. However, strong vertical currents may carry supercooled water to great heights where temperatures are much colder than −15° C. Supercooled water has been observed at temperatures colder than −40° C.

DEW AND FROST

During clear nights with little or no wind, vegetation often cools by radiation to a temperature at or below the dew point of the adjacent air. Moisture then collects on the leaves just as it does on a pitcher of ice water in a warm room. Heavy dew often collects on grass and plants when none collects on pavements or large solid objects. These more massive objects absorb abundant heat during the day, lose it slowly during the night, and cool below the dew point only in rather extreme cases.

Frost forms in much the same way as dew. The difference is that the dew point of surrounding air must be colder than freezing. Water vapor then sublimates directly as ice crystals or frost rather than condensing as dew. Sometimes dew forms and later freezes; however, frozen dew is easily distinguished from frost. Frozen dew is hard and transparent while frost is white and opaque.

To now, we have said little about clouds. What brings about the condensation or sublimation that results in cloud formation?

CLOUD FORMATION

Normally, air must become saturated for condensation or sublimation to occur. Saturation may result from cooling temperature, increasing dew point, or both. Cooling is far more predominant.

COOLING PROCESSES

Three basic processes may cool air to saturation. They are (1) air moving over a colder surface, (2) stagnant air overlying a cooling surface, and (3) expansional cooling in upward moving air. Expansional cooling is the major cause of cloud formation. Chapter 6, "Stable and Unstable Air," discusses expansional cooling in detail.

CLOUDS AND FOG

A cloud is a visible aggregate of minute water or ice particles suspended in air. If the cloud is on the ground, it is fog. When entire layers of air cool to saturation, fog or sheet-like clouds result. Saturation of a localized updraft produces a towering cloud. A cloud may be composed entirely of liquid water, of ice crystals, or a mixture of the two.

PRECIPITATION

Precipitation is an all inclusive term denoting drizzle, rain, snow, ice pellets, hail, and ice crystals. Precipitation occurs when these particles grow in size and weight until the atmosphere no longer can suspend them and they fall. These particles grow primarily in two ways.

PARTICLE GROWTH

Once a water droplet or ice crystal forms, it continues to grow by added condensation or sublimation directly onto the particle. This is the slower of the two methods and usually results in drizzle or very light rain or snow.

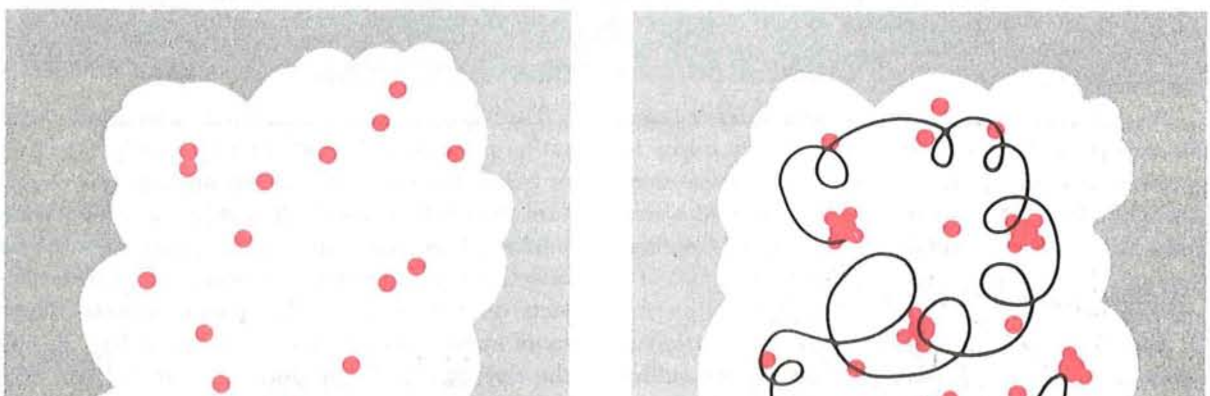

FIGURE 36. Growth of raindrops by collision of cloud droplets.

Cloud particles collide and merge into a larger drop in the more rapid growth process. This process produces larger precipitation particles and does so more rapidly than the simple condensation growth process. Upward currents enhance the growth rate and also support larger drops as shown in figure 36. Precipitation formed by merging drops with mild upward currents can produce light to moderate rain and snow. Strong upward currents support the largest drops and build clouds to great heights. They can produce heavy rain, heavy snow, and hail.

LIQUID, FREEZING, AND FROZEN

Precipitation forming and remaining liquid falls as rain or drizzle. Sublimation forms snowflakes, and they reach the ground as snow if temperatures remain below freezing.

Precipitation can change its state as the temperature of its environment changes. Falling snow may melt in warmer layers of air at lower altitudes to form rain. Rain falling through colder air may become supercooled, freezing on impact as freezing rain; or it may freeze during its descent, falling as ice pellets. Ice pellets always indicate freezing rain at higher altitude.

Sometimes strong upward currents sustain large supercooled water drops until some freeze; subsequently, other drops freeze to them forming hailstones.

PRECIPITATION VERSUS CLOUD THICKNESS

To produce significant precipitation, clouds usually are 4,000 feet thick or more. The heavier the precipitation, the thicker the clouds are likely to be. When arriving at or departing from a terminal reporting precipitation of light or greater intensity, expect clouds to be more than 4,000 feet thick.

LAND AND WATER EFFECTS

Land and water surfaces underlying the atmosphere greatly affect cloud and precipitation development. Large bodies of water such as oceans and large lakes add water vapor to the air. Expect the greatest frequency of low ceilings, fog, and precipitation in areas where prevailing winds have an over-water trajectory. Be especially alert for these hazards when moist winds are blowing upslope.

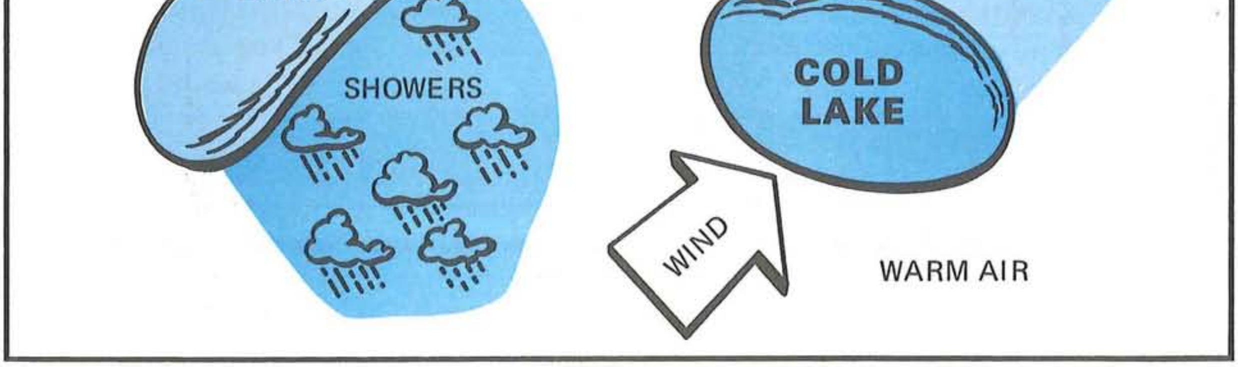

FIGURE 37. Lake effects. Air moving across a sizeable lake absorbs water vapor. Showers may appear on the leeward side if the air is colder than the water. When the air is warmer than the water, fog often develops on the lee side.

In winter, cold air frequently moves over relatively warm lakes. The warm water adds heat and water vapor to the air causing showers to the lee of the lakes. In other seasons, the air may be warmer than the lakes. When this occurs, the air may become saturated by evaporation from the water while also becoming cooler in the low levels by contact with the cool water. Fog often becomes extensive and dense to the lee of a lake. Figure 37 illustrates movement of air over both warm and cold lakes. Strong cold winds across the Great Lakes often carry precipitation to the Appalachians as shown in figure 38.

A lake only a few miles across can influence convection and cause a diurnal fluctuation in cloudiness. During the day, cool air over the lake blows toward the land, and convective clouds form over the land as shown in figure 39, a photograph of Lake Okeechobee in Florida. At night, the pattern reverses; clouds tend to form over the lake as cool air from the land flows over the lake creating convective clouds over the water.

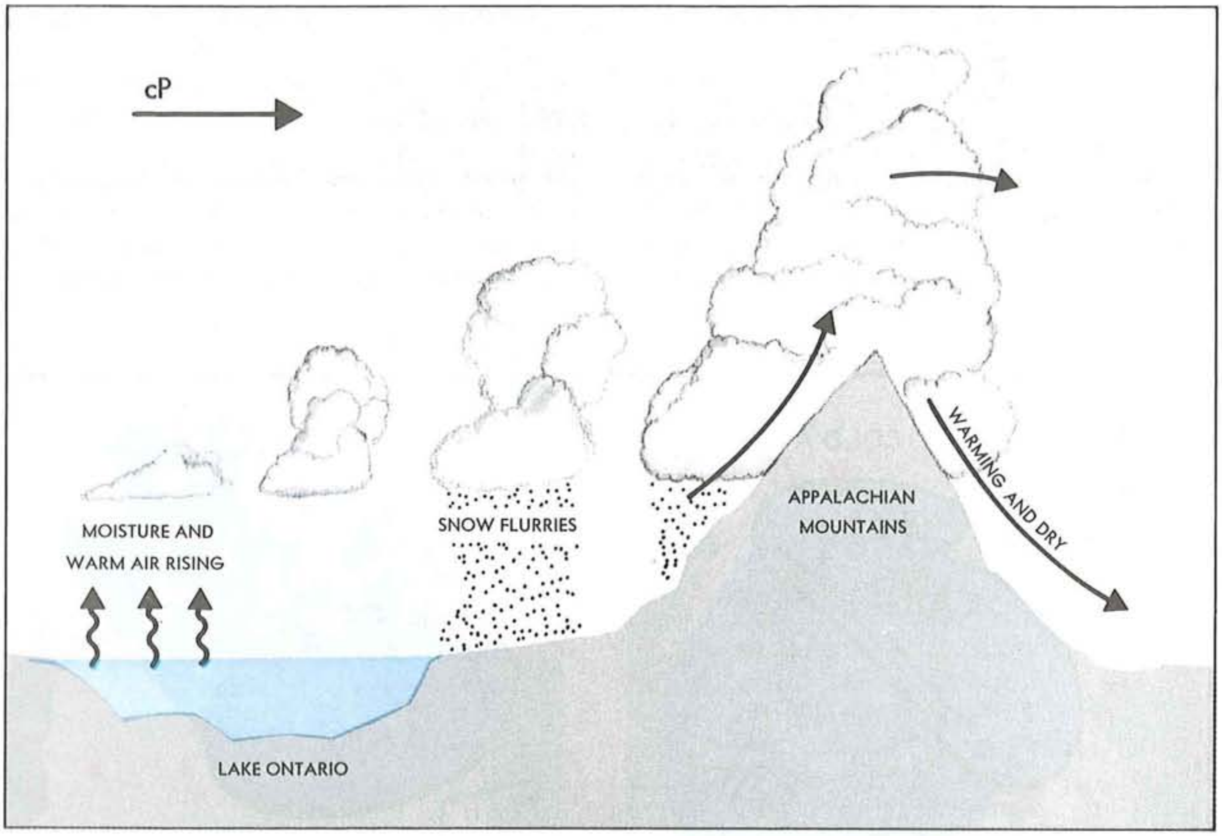

FIGURE 38. Strong cold winds across the Great Lakes absorb water vapor and may carry showers as far eastward as the Appalachians.

FIGURE 39. A view of clouds from 27,000 feet over Lake Okeechobee in southern Florida. Note the lake effect. During daytime, cool air from the lake flows toward the warmer land forming convective clouds over the land.

IN CLOSING

Water exists in three states—solid, liquid, and gaseous. Water vapor is an invisible gas. Condensation or sublimation of water vapor creates many common aviation weather hazards. You may anticipate:

1. Fog when temperature-dew point spread is 5° F or less and decreasing.
2. Lifting or clearing of low clouds and fog when temperature-dew point spread is increasing.
3. Frost on a clear night when temperature-dew point spread is 5° F or less, is decreasing, and dew point is colder than 32° F.
4. More cloudiness, fog, and precipitation when wind blows from water than when it blows from land.
5. Cloudiness, fog, and precipitation over higher terrain when moist winds are blowing uphill.
6. Showers to the lee of a lake when air is cold and the lake is warm. Expect fog to the lee of the lake when the air is warm and the lake is cold.
7. Clouds to be at least 4,000 feet thick when significant precipitation is reported. The heavier the precipitation, the thicker the clouds are likely to be.
8. Icing on your aircraft when flying through liquid clouds or precipitation with temperature freezing or colder.

Chapter 6
STABLE AND UNSTABLE AIR

To a pilot, the stability of his aircraft is a vital concern. A stable aircraft, when disturbed from straight and level flight, returns by itself to a steady balanced flight. An unstable aircraft, when disturbed, continues to move away from a normal flight attitude.

So it is with the atmosphere. A *stable* atmosphere resists any upward or downward displacement. An *unstable* atmosphere allows an upward or downward disturbance to grow into a vertical or convective current.

This chapter first examines fundamental changes in upward and downward moving air and then relates stable and unstable air to clouds, weather, and flying.

CHANGES WITHIN UPWARD AND DOWNWARD MOVING AIR

Anytime air moves upward, it expands because of decreasing atmospheric pressure as shown in figure 40. Conversely, downward moving air is compressed by increasing pressure. But as pressure and volume change, temperature also changes.

When air expands, it cools; and when compressed, it warms. These changes are *adiabatic,* meaning that no heat is removed from or added to the air. We frequently use the terms *expansional* or *adiabatic cooling* and *compressional* or *adiabatic*

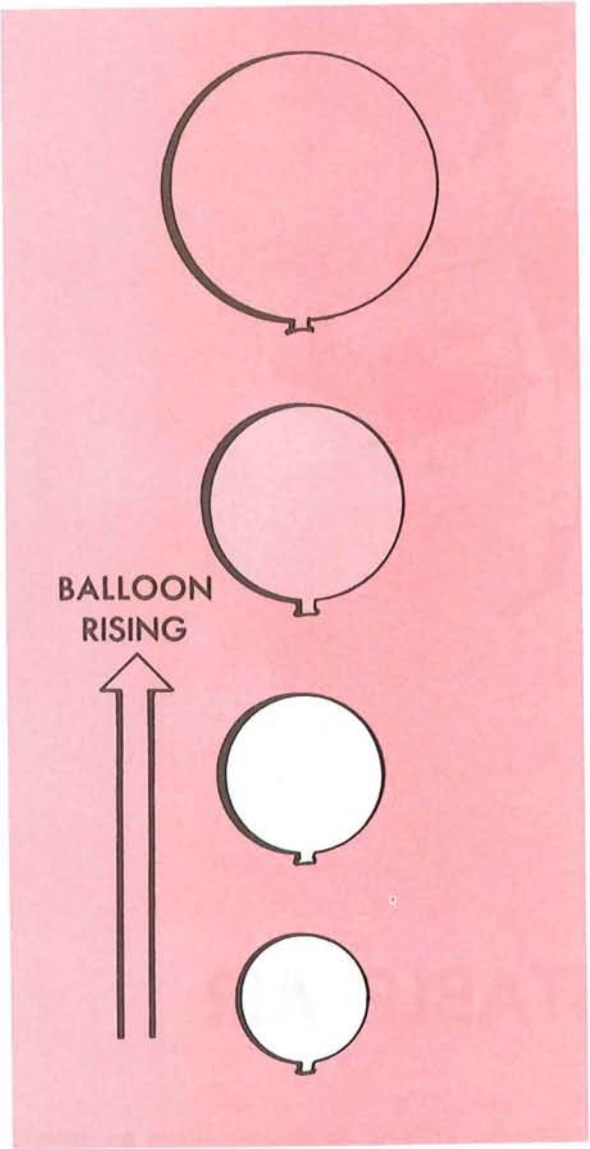

FIGURE 40. Decreasing atmospheric pressure causes the balloon to expand as it rises. Anytime air moves upward, it expands.

heating. The adiabatic rate of change of temperature is virtually fixed in unsaturated air but varies in saturated air.

UNSATURATED AIR

Unsaturated air moving upward and downward cools and warms at about 3.0° C (5.4° F) per 1,000 feet. This rate is *the "dry adiabatic rate of temperature change"* and *is independent of the temperature of the mass of air through which the vertical movements occur*. Figure 41 illustrates a "Chinook Wind"—an excellent example of dry adiabatic warming.

SATURATED AIR

Condensation occurs when *saturated* air moves upward. Latent heat released through condensation (chapter 5) partially offsets the expansional cooling. Therefore, *the saturated adiabatic rate of cooling is slower than the dry adiabatic rate*. The saturated rate depends on saturation temperature or dew point of the air. Condensation of copious moisture in saturated warm air releases more latent heat to offset expansional cooling than does the scant moisture in saturated cold air. Therefore, *the saturated adiabatic rate of cooling is less in warm air than in cold air*.

When saturated air moves downward, it heats at the same rate as it cools on ascent *provided* liquid water evaporates rapidly enough to maintain saturation. Minute water droplets evaporate at virtually this rate. Larger drops evaporate more slowly and complicate the moist adiabatic process in downward moving air.

ADIABATIC COOLING AND VERTICAL AIR MOVEMENT

If we force a sample of air upward into the atmosphere, we must consider two possibilities:

(1) The air may become colder than the surrounding air, or

(2) Even though it cools, the air may remain warmer than the surrounding air.

If the upward moving air becomes colder than surrounding air, it sinks; but if it remains warmer, it is accelerated upward as a convective current. Whether it sinks or rises depends on the ambient or existing temperature lapse rate (chapter 2).

Do not confuse existing lapse rate with adiabatic rates of cooling in vertically moving air.* The difference between the existing lapse rate of a given mass of air and the adiabatic rates of cooling in upward moving air determines if the air is stable or unstable.

*Sometimes you will hear the dry and moist adiabatic rates of cooling called the dry adiabatic lapse rate and the moist adiabatic lapse rate. In this book, *lapse rate* refers exclusively to the existing, or actual, decrease of temperature with height in a real atmosphere. The dry or moist adiabatic lapse rate signifies a prescribed rate of expansional cooling or compressional heating. An adiabatic lapse rate becomes real *only* when it becomes a condition brought about by vertically moving air.

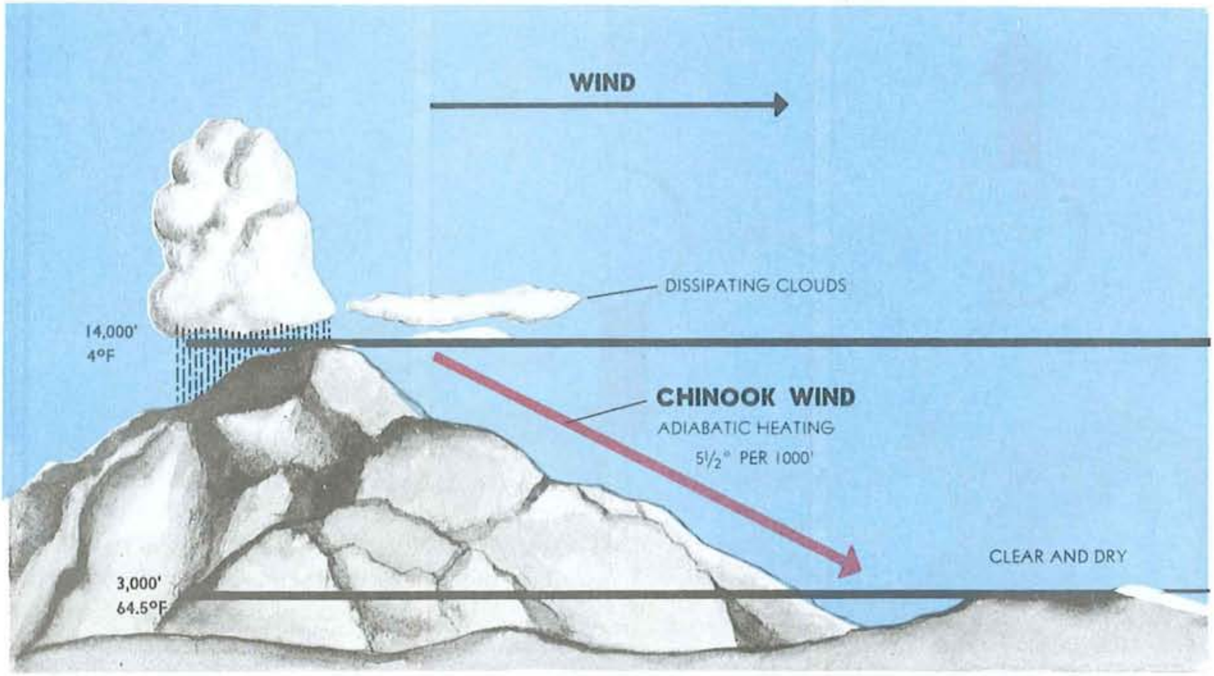

FIGURE 41. Adiabatic warming of downward moving air produces the warm Chinook wind.

STABILITY AND INSTABILITY

Let's use a balloon to demonstrate stability and instability. In figure 42 we have, for three situations, filled a balloon at sea level with air at 31° C —the same as the ambient temperature. We have carried the balloon to 5,000 feet. In each situation, the air in the balloon expanded and cooled at the dry adiabatic rate of 3° C for each 1,000 feet to a temperature of 16° C at 5,000 feet.

In the first situation (left), air inside the balloon, even though cooling adiabatically, remains warmer than surrounding air. Vertical motion is favored. The colder, more dense surrounding air forces the balloon on upward. This air is unstable, and a convective current develops.

In situation two (center) the air aloft is warmer. Air inside the balloon, cooling adiabatically, now becomes colder than the surrounding air. The balloon sinks under its own weight returning to its original position when the lifting force is removed. The air is stable, and spontaneous convection is impossible.

In the last situation, temperature of air inside the balloon is the same as that of surrounding air. The balloon will remain at rest. This condition is neutrally stable; that is, the air is neither stable nor unstable.

Note that, *in all three situations, temperature of air in the expanding balloon cooled at a fixed rate.* The differences in the three conditions depend, therefore, on the temperature differences between the surface and 5,000 feet, that is, on the ambient lapse rates.

HOW STABLE OR UNSTABLE?

Stability runs the gamut from absolutely stable to absolutely unstable, and the atmosphere usually is in a delicate balance somewhere in between. A change in ambient temperature lapse rate of an air mass can tip this balance. For example, surface heating or cooling aloft can make the air more unstable; on the other hand, surface cooling or warming aloft often tips the balance toward greater stability.

Air may be stable or unstable in layers. A stable layer may overlie and cap unstable air; or, conversely, air near the surface may be stable with unstable layers above.

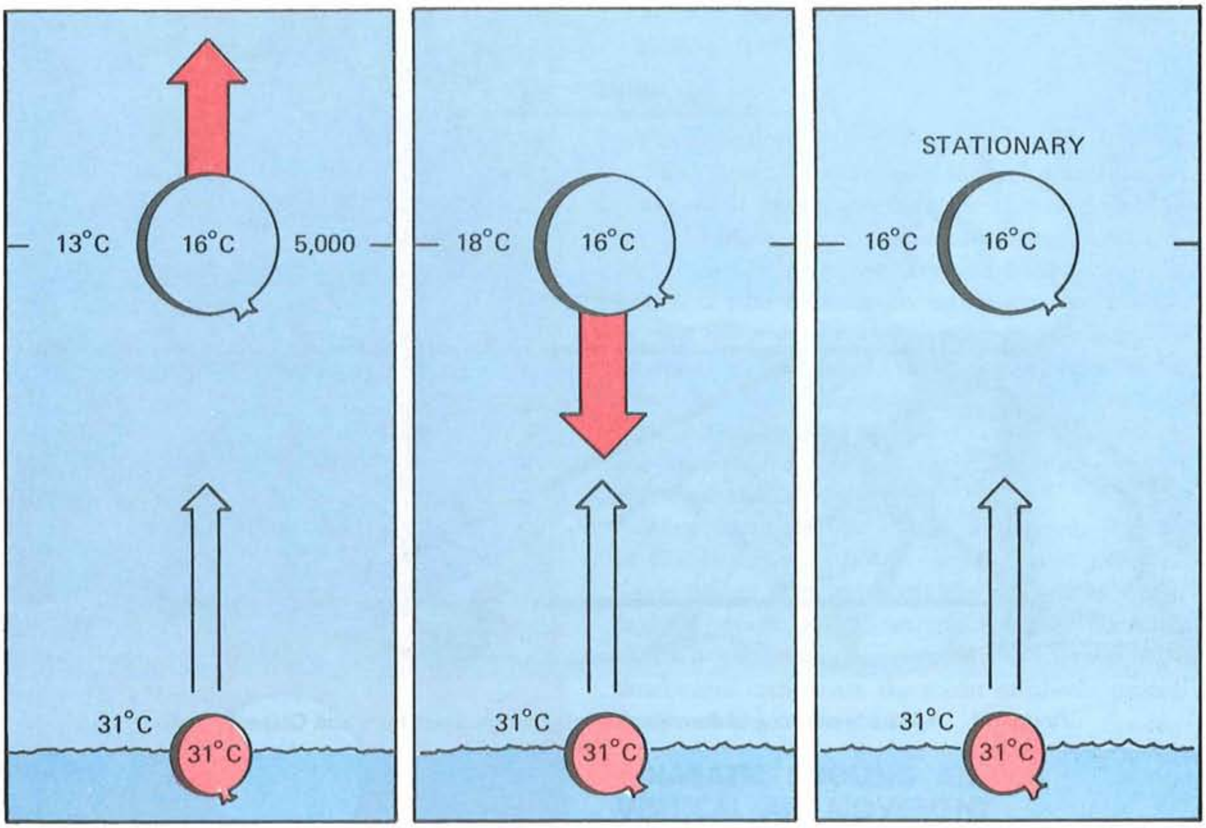

FIGURE 42. Stability related to temperatures aloft and adiabatic cooling. In each situation, the balloon is filled at sea level with air at 31° C, carried manually to 5,000 feet, and released. In each case, air in the balloon expands and cools to 16° C (at the dry adiabatic rate of 3° C per 1,000 feet). But, the temperature of the surrounding air aloft in each situation is different. The balloon on the left will rise. Even though it cooled adiabatically, the balloon remains warmer and lighter than the surrounding cold air; when released, it will continue upward spontaneously. The air is unstable; it favors vertical motion. In the center, the surrounding air is warmer. The cold balloon will sink. It resists our forced lifting and cannot rise spontaneously. The air is stable—it resists upward motion. On the right, surrounding air and the balloon are at the same temperature. The balloon remains at rest since no density difference exists to displace it vertically. The air is neutrally stable, i.e., it neither favors nor resists vertical motion. A mass of air in which the temperature decreases rapidly with height favors instability; but, air tends to be stable if the temperature changes little or not at all with altitude.

CLOUDS—STABLE OR UNSTABLE?

Chapter 5 states that when air is cooling and first becomes saturated, condensation or sublimation begins to form clouds. Chapter 7 explains cloud types and their significance as "signposts in the sky." Whether the air is stable or unstable within a layer largely determines cloud structure.

Stratiform Clouds

Since stable air resists convection, clouds in stable air form in horizontal, sheet-like layers or "strata." Thus, within a *stable* layer, clouds are *stratiform.* Adiabatic cooling may be by upslope flow as illustrated in figure 43; by lifting over cold, more dense air; or by converging winds. Cooling by an underlying cold surface is a stabilizing process and may produce fog. If clouds are to remain stratiform, the layer must remain stable after condensation occurs.

Cumuliform Clouds

Unstable air favors convection. A "cumulus" cloud, meaning "heap," forms in a convective updraft and builds upward, also shown in figure 43. Thus, within an *unstable* layer, clouds are *cumuliform;* and the vertical extent of the cloud depends on the depth of the unstable layer.

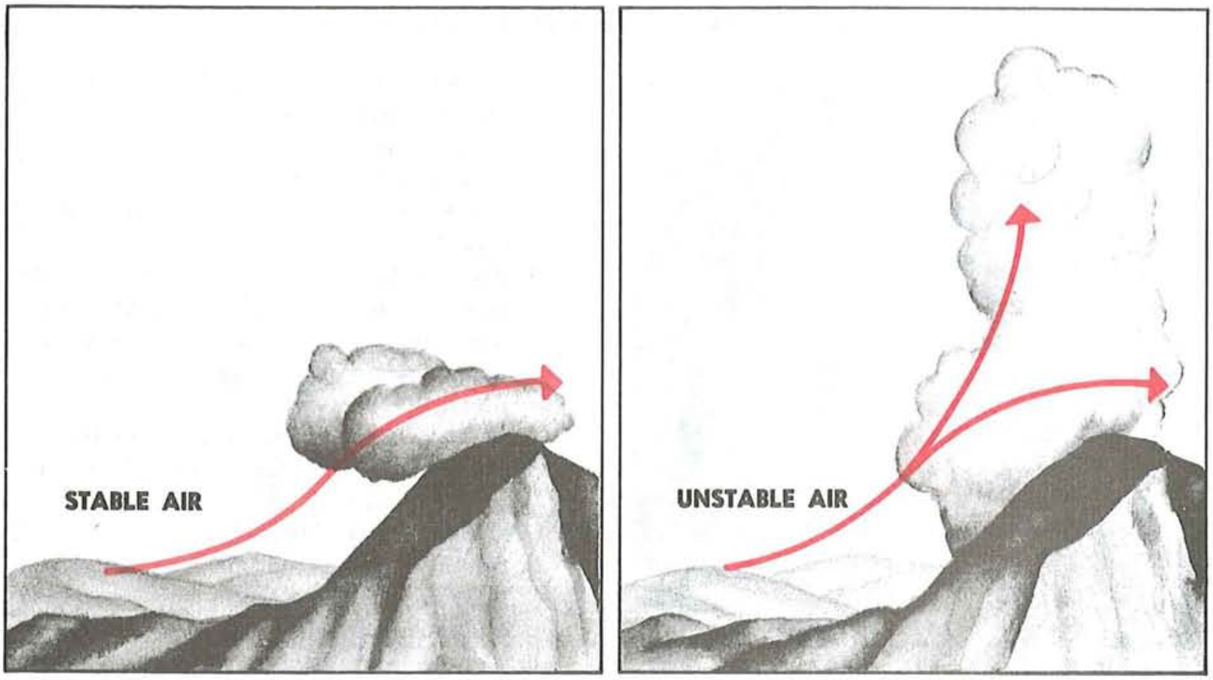

FIGURE 43. When stable air (left) is forced upward, the air tends to retain horizontal flow, and any cloudiness is flat and stratified. When unstable air is forced upward, the disturbance grows, and any resulting cloudiness shows extensive vertical development.

Initial lifting to trigger a cumuliform cloud may be the same as that for lifting stable air. In addition, convection may be set off by surface heating (chapter 4). Air may be unstable or slightly stable before condensation occurs; but for convective cumuliform clouds to develop, it must be unstable after saturation. Cooling in the updraft is now at the slower moist adiabatic rate because of the release of latent heat of condensation. Temperature in the saturated updraft is warmer than ambient temperature, and convection is spontaneous. Updrafts accelerate until temperature within the cloud cools below the ambient temperature. This condition occurs where the unstable layer is capped by a stable layer often marked by a temperature inversion. Vertical heights range from the shallow fair weather cumulus to the giant thunderstorm cumulonimbus—the ultimate in atmospheric instability capped by the tropopause.

You can estimate height of cumuliform cloud bases using surface temperature-dew point spread. Unsaturated air in a convective current cools at about 5.4° F (3.0° C) per 1,000 feet; dew point decreases at about 1° F (5/9° C). Thus, in a convective current, temperature and dew point converge at about 4.4° F (2.5° C) per 1,000 feet as illustrated in figure 44. We can get a quick *estimate* of a convective cloud base in thousands of feet by rounding these values and dividing into the spread or by multiplying the spread by their reciprocals. When using Fahrenheit, divide by 4 or multiply by .25; when using Celsius, divide by 2.2 or multiply by .45. This method of estimating is reliable only with instability clouds and during the warmer part of the day.

When unstable air lies above stable air, convective currents aloft sometimes form middle and high level cumuliform clouds. In relatively shallow layers they occur as altocumulus and ice crystal cirrocumulus clouds. Altocumulus castellanus clouds develop in deeper midlevel unstable layers.

Merging Stratiform and Cumuliform

A layer of stratiform clouds may sometimes form in a mildly stable layer while a few ambitious convective clouds penetrate the layer thus merging stratiform with cumuliform. Convective clouds may be almost or entirely embedded in a massive stratiform layer and pose an unseen threat to instrument flight.

WHAT DOES IT ALL MEAN?

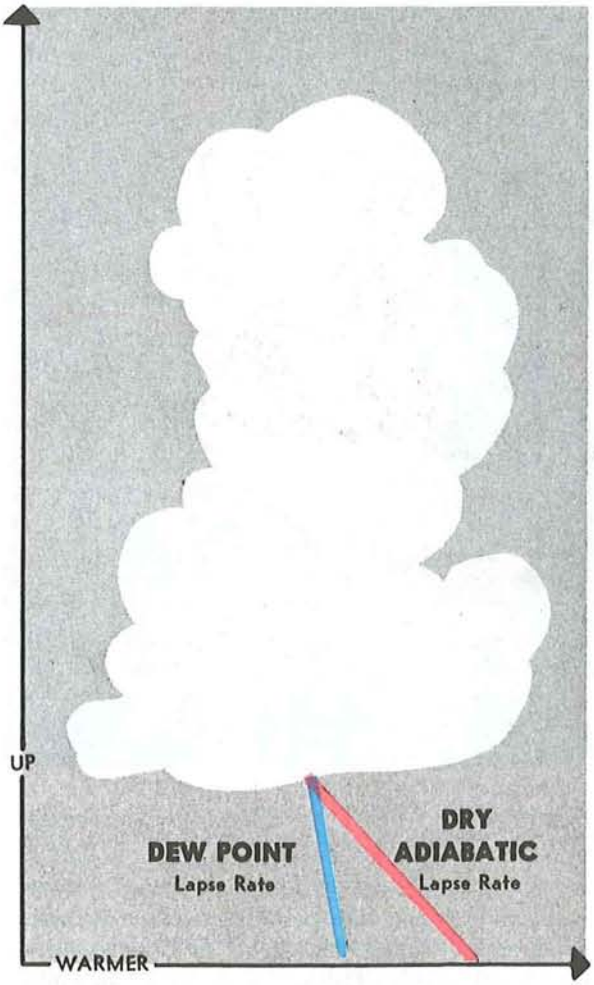

FIGURE 44. Cloud base determination. Temperature and dew point in upward moving air converge at a rate of about 4° F or 2.2° C per 1,000 feet.

Can we fly in unstable air? Stable air? Certainly we can and ordinarily do since air is seldom neutrally stable. The usual convection in unstable air gives a "bumpy" ride; only at times is it violent enough to be hazardous. In stable air, flying is usually smooth but sometimes can be plagued by low ceiling and visibility. It behooves us in preflight planning to take into account stability or instability and any associated hazards. Certain observations you can make on your own:

1. Thunderstorms are sure signs of violently unstable air. Give these storms a wide berth.
2. Showers and clouds towering upward with great ambition indicate strong updrafts and rough (turbulent) air. Stay clear of these clouds.
3. Fair weather cumulus clouds often indicate bumpy turbulence beneath and in the clouds. The cloud tops indicate the approximate upper limit of convection; flight above is usually smooth.
4. Dust devils are a sign of dry, unstable air, usually to considerable height. Your ride may be fairly rough unless you can get above the instability.
5. Stratiform clouds indicate stable air. Flight generally will be smooth, but low ceiling and visibility might require IFR.
6. Restricted visibility at or near the surface over large areas usually indicates stable air. Expect a smooth ride, but poor visibility may require IFR.
7. Thunderstorms may be embedded in stratiform clouds posing an unseen threat to instrument flight.
8. Even in clear weather, you have some clues to stability, viz.:
   a. When temperature decreases uniformly and rapidly as you climb (approaching 3° C per 1,000 feet), you have an indication of unstable air.
   b. If temperature remains unchanged or decreases only slightly with altitude, the air tends to be stable.
   c. If the temperature increases with altitude through a layer—an inversion—the layer is stable and convection is suppressed. Air may be unstable beneath the inversion.
   d. When air near the surface is warm and moist, suspect instability. Surface heating, cooling aloft, converging or upslope winds, or an invading mass of colder air may lead to instability and cumuliform clouds.