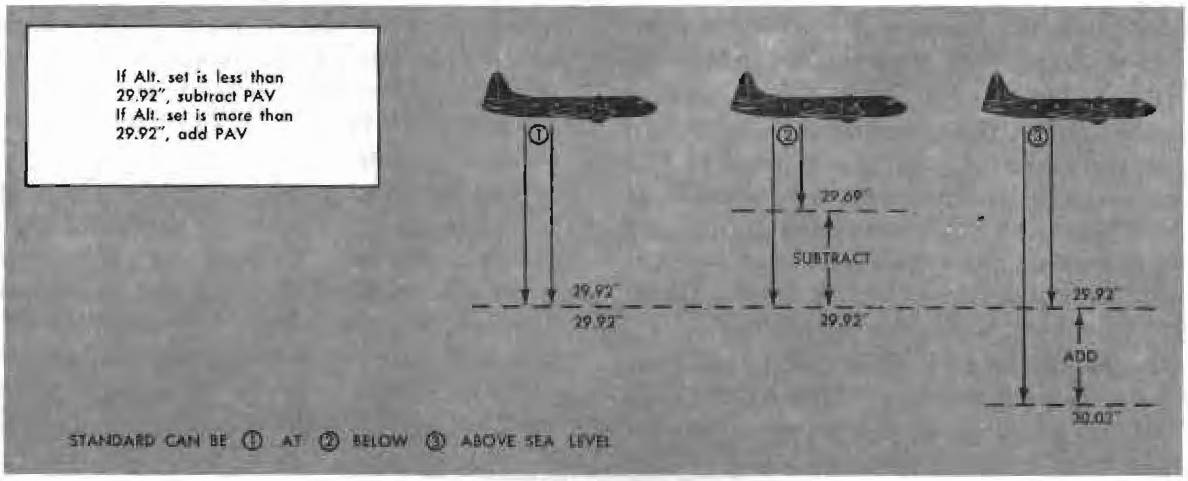


Figure 6–15. Finding Computed True Altitude with Pressure-Altitude Variation

If altimeter setting is 29.69 in., the pressure altitude variation is 230 ft. Since 29.92 in. is greater than 29.69 in., the standard datum plane is below the plane above which the true altitude is measured. Therefore, considering PAV only, pressure altitude is greater than true altitude, and pressure altitude variation must be subtracted from pressure altitude corrected for temperature to find computed true altitude.

If altimeter setting is greater than 29.92 in., pressure altitude variation must be added to pressure altitude corrected for temperature. If the altimeter setting is less than 29.92 in., pressure altitude variation must be subtracted from pressure altitude corrected for temperature. These rules are harder to remember than they are to figure. Just remember that pressure decreases upward and think of the relative positions of the datum planes.

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

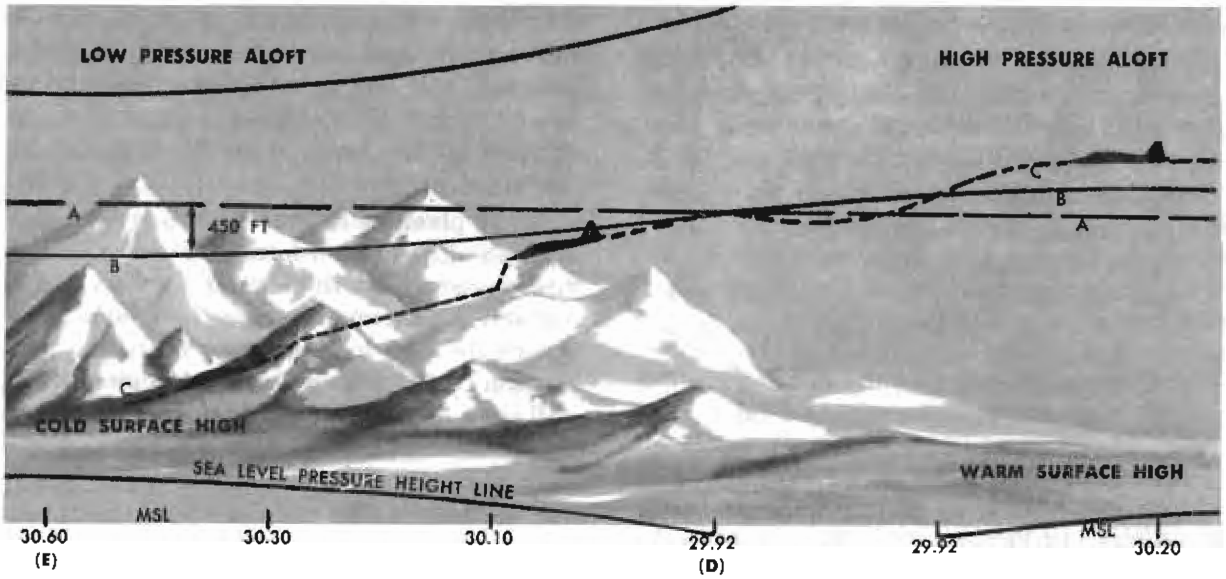
The example in Figure 6-16 shows a flight proceeding from one area of high *surface* pressure to another area of high *surface* pressure. Although altimeter settings are received at regular intervals along the route and flight corrections made accordingly, the net result in this particular example is a gradual loss of actual altitude above mean sea level. Notice that if a pilot made an actual flight under the conditions as illustrated, disaster could result if the aircraft were flying at minimum altitude clearance in mountainous terrain. The basic information needed to make such a flight safely is readily available in the weather station.

Suppose a flight is proposed at 11,000 feet MSL. The forecaster can determine whether the height indicated by the pressure altimeter at the proposed filing height is above or below the actual height. The forecaster determines this by doing two things. First, he makes a quick check of the current altimeter setting in the area of highest terrain against the standard setting (29.92). This is one factor which helps to determine how much the aircraft will be above or below the proposed altitude while flying at an indicated altitude of 11,000 feet. Next, he will examine the constant pressure charts to determine where the 11,000-foot pressure altitude level will be during the flight compared to where it would be if the conditions of a standard atmosphere prevailed. The difference is the so-called *altimeter correction* or *D value.*

The D value equals the difference between the true altitude of the pressure surface in question, minus the altitude of this surface in the standard atmosphere.

$$D = \text{True Altitude} - \text{Standard Altitude}$$

In the example (Figure 6-16), the forecaster finds that the 11,000-foot pressure altitude



A. PROPOSED FLIGHT PATH OF 11,000 FEET MSL.

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

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

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

Figure 6–16. Flight Path into an Area Having a Cold Surface High with Low Pressure Aloft

level over the highest terrain is 450 feet lower than it would be in the standard atmosphere. If a constant altimeter setting of 29.92 in. were used, the true altitude of the aircraft would be 450 feet less than the pressure altitude over this highest terrain. However, the altimeter setting will be changed en route to the current values. By adding algebraically these two effects (altimeter setting vs 29.92 in. and D value), it is possible to determine if the aircraft can clear the terrain.

In our problem, a flight is planned into an area of a cold high at the surface coupled with a cold low aloft. This is the most dangerous situation for terrain clearance, but is common in the atmosphere. If this flight were conducted with an altimeter setting of 29.92 in., the flight path would closely approximate curve "B" in the illustration. However, the pilot has used various altimeter settings along the route as required by FAA regulations. Notice that flying toward the cold surface high requires use of consecutively higher altimeter settings. Each time the pilot raises his setting, his altimeter reads a corresponding increase in indicated altitude, although the true altitude is actually decreasing. In order to maintain a constant indicated altitude of 11,000 feet, the pilot must descend each time a higher setting is used (as the altimeter setting is raised the indicated altitude increases).

Now let us see how low our aircraft would be when it reaches the left of the sketch. The altimeter setting near the end of the flight is 30.60 in. Since one inch of pressure is equivalent to approximately 1,000 feet of altitude near sea level in the standard atmosphere, we multiply the .68-inch difference (30.60 minus 29.92) by 1,000. This gives a difference of 680 feet between the basic pressure altitude at this point and the indicated altitude with 30.60 in. in the Kollsman window.

Current setting	30.60 inches
Standard setting	29.92 inches
	.68 inches
	equivalent to 680 feet

In this example, the 680 feet PAV is positive since the altimeter setting is higher than standard. With 30.60 in. in the Kollsman window the altimeter will indicate an altitude 680 feet higher than the pressure altitude. Since the pressure altitude is 450 feet lower than the true altitude (D value equals minus 450 feet) we can combine the two effects:

Pressure surface lower than standard (D value)	— 450 feet
Altimeter setting higher than standard	— 680 feet
	—1130 feet

Although the altimeter would be indicating 11,000 feet, the total error would be —1130 feet. Therefore, the actual altitude would be approximately 9870 feet above sea level. Notice in the illustration that although the pilot is reading 11,000 feet on the altimeter, he could easily fly into a 10,000-foot mountain. Changing the altimeter setting en route in this example could cause a crash if the clearance problem were not solved before takeoff.

While flying toward the warm high at the right side of the figure, changing altimeter settings en route will give an indicated altitude which is closer to the true altitude than the pressure altitude. For instance, changing the setting from 29.92 in. to 30.20 in. will produce a higher indicated reading. In order to maintain 11,000 feet MSL indicated altitude, the pilot must descend accordingly. This descent will enable flight closer to the actual 11,000-foot MSL level.

The advantage of changing altimeter settings for landing is obvious. In order for all aircraft in a region to use the same basis for altitude separation, it is necessary for all aircraft to use the same setting. Therefore, the FAA requires use of altimeter settings at or below 23,500 feet MSL. A pilot must not forget the hazard of flying toward lower pressure at flight altitude or toward colder mean temperatures. When you fly into both lower pressure aloft with higher pressure at the surface and colder temperatures, the most hazardous condition exists; or "HALP", a high altimeter setting coupled with low pressure at flight altitude, is the dangerous combination.

Remember, an altimeter measures only the weight of the air above it. Any altimeter setting a pilot receives from a ground station is a measure of the weight of the air above the station. If a pilot is flying above that station, its report will include the weight of the air both above and below him. The altimeter, however, will only give a measure of the air above the aircraft and will not correct or compensate for the layer of air between the aircraft and the station.

When planning a flight over mountains, it is advisable to obtain a forecast *D value* and altimeter setting at the highest terrain. The D value will immediately give the true altitude departure from basic pressure altitude. Comparison of 29.92 in. with the altimeter setting will show any additional error that will be incurred. This should be done during flight planning.

BASIC WIND THEORY

If it were not for wind, air navigation would be simple, for a reliable position could be obtained from only true heading, airspeed, and elapsed time. It is moving air, sometimes blowing at speeds of less than 5 knots and at other times "jet streaming" up to 200 knots, that complicates navigation. A general understanding of wind theory is, therefore, imperative for the aviator.

Pressure Gradient Force

We have learned that horizontal atmospheric pressure variations do exist. The pressure equalizing force that tries to drive air from high to low pressure is called the *pressure gradient force.* A pressure gradient is the rate of change of pressure in the horizontal plane measured perpendicular to the isobars (or contours) toward lower pressure. The closer the isobars (or contours) are together, the stronger will be the pressure gradient force and the resulting wind speed,(Figure 6-17).

Coriolis Effect (Force)

When first inspecting a constant pressure chart, it is apparent that the wind is not blowing from high to low pressure regions, but rather more along the contours. This phenomenon is the result of the *coriolis force.*

If the earth did not rotate, air would flow directly from high pressure to low pressure. This actually occurs for small-scale air movements, but when air moves over a considerable distance, the air is deflected because of the spinning of the earth on its axis. In fact, any freely-moving body is deflected as it moves over the earth. This deflection is with respect to the earth. Freely-moving bodies over the earth as viewed from a point in the solar system, away from the earth, would appear to follow a straight line course; but to one stationed on earth, the path of the freely-moving body would appear to be a curve. Hence, the "force" is only an apparent force, but the deflection is a real deflection as far as we on earth are concerned.

This deflective force is called the coriolis force, named after G. G. Coriolis, a French mathematical physicist. The deflection is to the right of the direction of motion in the Northern Hemisphere, and to the left of the direction of motion in the Southern Hemisphere. By convention, all wind directions are designated as the direction *from* which the wind is blowing. A wind blowing from Kansas City toward St. Louis would be a *west* wind; *from* the west. A west wind without coriolis force becomes a north wind with coriolis force. A north wind without coriolis force becomes an east wind with coriolis force, and so on, in the Northern Hemisphere. Before attempting an explanation of the effects of coriolis force, let us gain a few fundamental concepts about the earth.

Since the earth makes one complete rotation each 24 hours, it can be seen in Figure 6-18 that the linear speed of a point on the earth's surface decreases with increasing latitude. If a freely moving body were forced northward from point A to point B, it would enter an area of slower rotating surface of the earth (rotating toward the east) and would show a deflection to the right when viewed from the earth. From space, the body's path would be a straight resultant of the original eastward velocity of the body and the northward thrust.

Imagine twirling a rock tied to a string. As the string is shortened, the rock twirls faster. In the sketch, as the distance of the body from the earth's axis is shortened, an increase in easterly speed is observed. The ice skater uses this principle when he folds his arms during a spin.

Now consider an unguided body moving from point C toward point D. It will enter an area of faster rotating surface of the earth and will show a deflection to the right again to the observer on the spinning earth. Remember, if the length of the string is increased, the rock

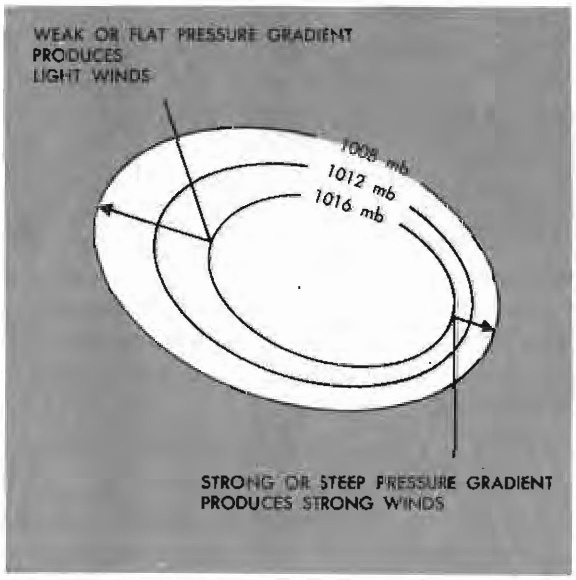


Figure 6–17. Pressure Gradient Force

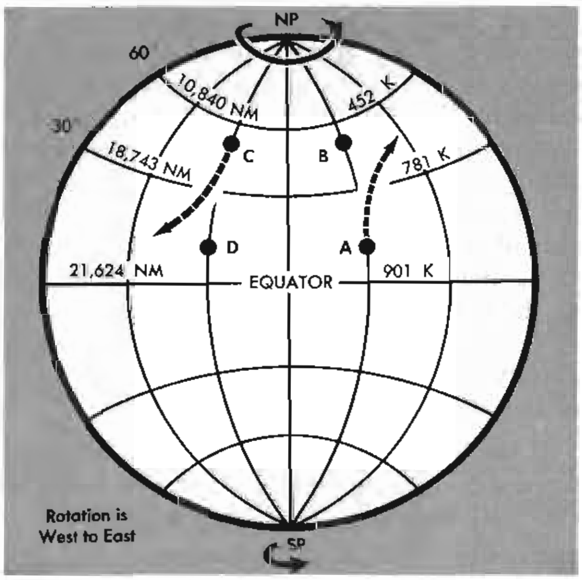


Figure 6–18. Horizontal Deflecting Force

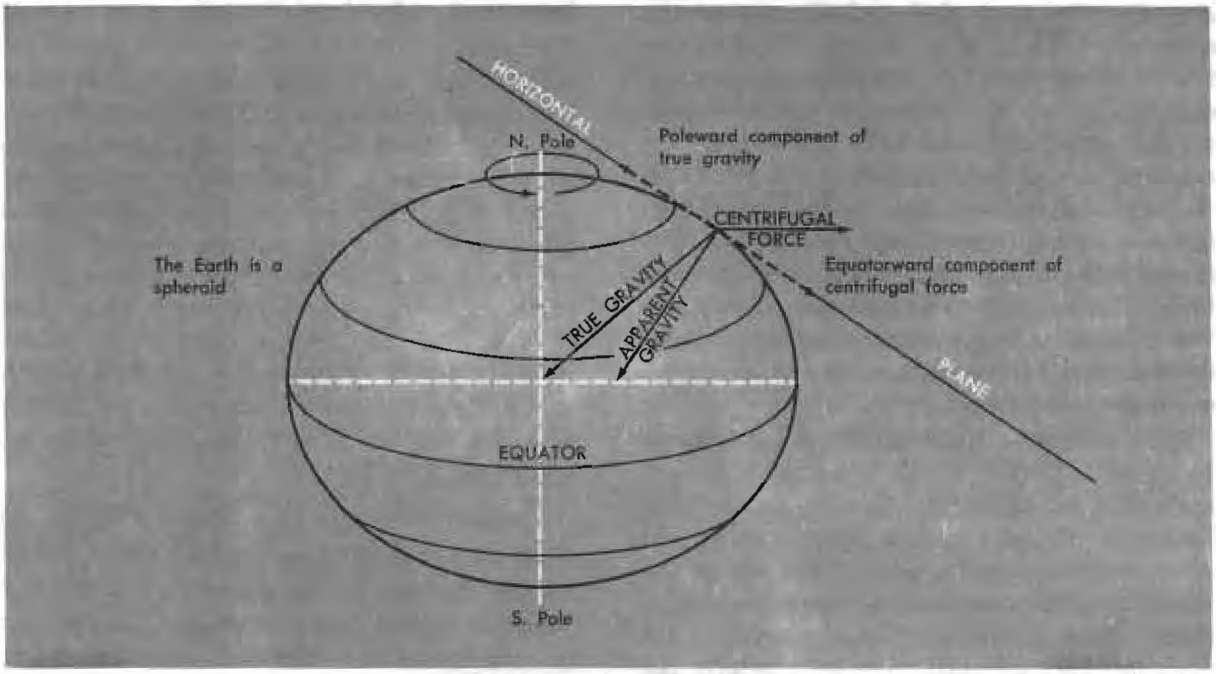


Figure 6–19. The Coriolis Effect on Eastward and Westward Movement

will twirl more slowly. In the example of the freely-moving body the distance to the earth's axis has been increased and the body must show a decrease in its original easterly rotating speed. The earth at point D has outrun the body.

It is more difficult to visualize the coriolis effect on eastward and westward freely-moving bodies. Apparent gravity, the force we feel, is composed of true gravity, directed toward the earth's center, and the centrifugal force produced by the earth's rotation. Apparent gravity is perpendicular to the earth's surface while true gravity is not. Notice that true gravity has a poleward component. Under the influence of true gravity alone, a body free to move along the earth's surface would come to rest at the pole. For a body at rest with respect to the earth this poleward component of true gravity is balanced by the equatorward component of centrifugal force.

If the body moves eastward, the centrifugal force will increase because the total speed of the body has increased (centrifugal force is proportional to the speed squared), since the body now has a speed around the earth's axis which is the sum of the speed of the earth's surface around the axis plus the speed of the body relative to the earth. The equatorward component of centrifugal force will also increase. Now the equatorward component of centrifugal force is greater than the poleward component of true gravity (which does not change), and the body is again deflected to the right of its direction of motion.

Lastly, consider coriolis effect on westward moving unsteered bodies. Since the body is moving relative to the earth in a direction opposite that of the earth's rotation, its total speed around the earth's axis is less than that of a body motionless on the earth's surface. Since the speed is less, the total centrifugal force exerted on the body is less, and the equatorward component of the centrifugal force is less. Now the poleward component of true gravity is greater than the equatorward component of centrifugal force, and the body is deflected to the right.

We have shown that regardless of the direction of motion, an unsteered body, such as a parcel of air, will always be deflected to the right of the direction of motion in the Northern Hemisphere. The same conclusion must be reached if one imagines himself riding on

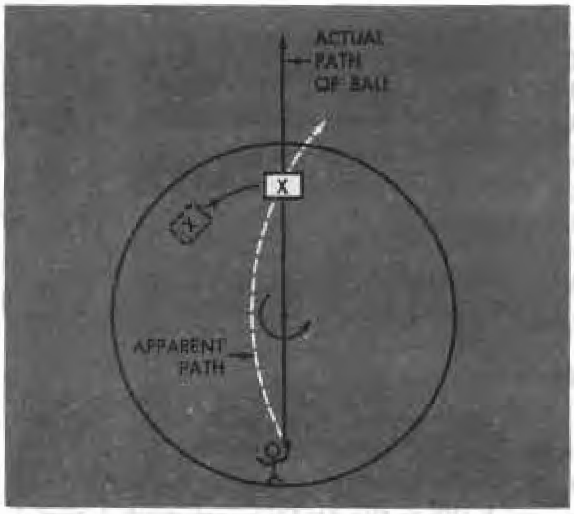


Figure 6–20. Merry-go-Round Example of Coriolis Effect

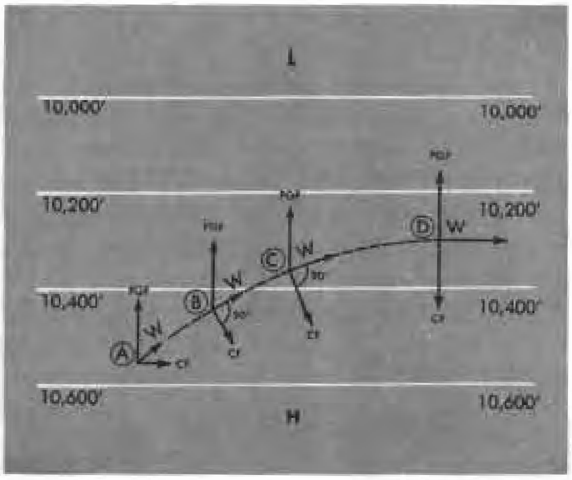


Figure 6–21. Geostrophic Wind

the merry-go-round, shown in Figure 6-20. If a ball is thrown at target "X" without allowing for the rotation of the merry-go-round, the ball will always miss. "X" will move with time to the position shown by the dashed box, and a deflection of the ball to the right will be apparent to the observer on the merry-go-round.

Since the earth is essentially symmetrical with respect to its equator, it could be shown that while coriolis deflects to the right in the Northern Hemisphere, it deflects to the left of the direction of motion in the Southern Hemisphere. The magnitude of this deflecting force is a function of several things, two of which are latitude and speed. The coriolis force is zero at the equator and increases toward the poles to become a maximum for any given speed at the poles. It might be noted that the coriolis force is zero for a motionless body at any point on the earth's surface.

GEOSTROPHIC WIND

Considering only pressure gradient and coriolis forces, it is possible to discuss a theoretical wind of great importance to meteorology and navigation. This wind is called the *geostrophic wind* to signify one resulting from the turning of the earth and pressure gradient forces.

Referring to Figure 6-21, assume a parcel of air at "A" is subjected to a pressure gradient force (PGF) directed toward the north (as indicated). While the parcel was at rest, the coriolis force (CF) was zero. Now as the parcel initially moves across the contours toward lower pressure, coriolis force increases from zero. The resultant of these two forces is the wind (W). At point "B" the parcel is still moving toward lower pressure and accelerates. Accordingly, the coriolis force increases, and the air is deflected more toward the right. This process of accelerating wind speed increasing the deflecting force continues at point "C". At point "D" the pressure gradient force and coriolis force are equal in magnitude and opposite in direction. Without additional accelerating forces, the parcel will move indefinitely in a direction parallel to the contours with the speed it has attained at point "D".

This first wind is blowing parallel to the contours with lower pressure (height) to the left and higher pressure (height) to the right of the direction of motion. Buys-Ballot's Law states that in the Northern Hemisphere with one's back to the wind lower pressure is to the left. The reverse is true in the Southern Hemisphere, since coriolis deflects to the left there.

This theoretical geostrophic wind demands straight parallel contours and the absence of friction. Happily, above the surface friction layer (generally two to three thousand feet), it is an excellent approximation to the actual wind in the majority of cases. However, at

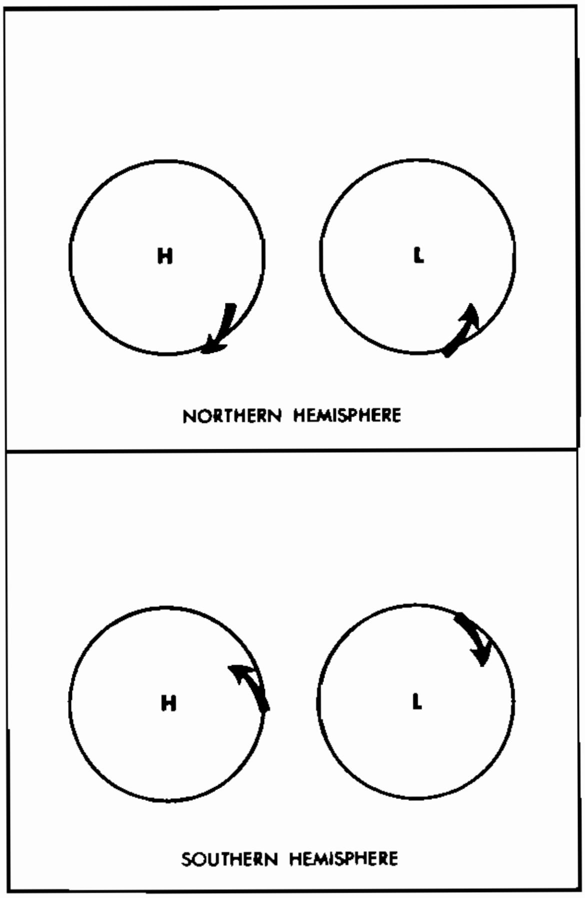


Figure 6–22. Circulations

low latitudes (20°N to 20°S) the geostrophic wind is a poor approximation because coriolis force decreases with latitude to zero at the equator, regardless of the wind speed. The crosswind component of the geostrophic wind can be computed in flight by the navigator using the radio and pressure altimeters. The geostrophic wind thus forms the basis for pressure differential navigation. It is also used for *single heading*, *minimal*, *pressure pattern*, and *optimum* flight planning. (References: AFM 51-40, Vol. II, and MATS Manual 55-7.)

Circulations

Consider the accompanying diagrams concerning *circulations*. Remember that coriolis deflects the wind to the right in the Northern Hemisphere, and to the left in the Southern Hemisphere. The pressure gradient force always acts toward lower pressure. Since the direction of deflection caused by the coriolis force is opposite in the Northern and Southern Hemispheres, we notice clockwise flow about highs in the Northern Hemsphere and about lows in the Southern Hemisphere. We define circulation about lows as always being cyclonic, and the contours are said to have cyclonic curvature. Circulation about highs is said to be anticyclonic, and the contours have anticyclonic curvature.

Gradient Wind

It was pointed out earlier that the geostrophic wind could only be the same as the true wind when the contours are straight and parallel. When the contours are curved, centrifugal force must be considered in addition to the pressure gradient and coriolis forces. The gradient wind is, therefore, simply the geostrophic wind corrected for contour curvature. Sometimes meteorologists speak of the wind at the *gradient wind level*, meaning not the gradient wind here defined but the wind observed at the top of the friction layer — see *Friction*, later in this chapter. In the related illustration (Figure 6-23), the wind at point "A" is geostrophic since the contours are straight and parallel.

The air will strive to continue movement to the west. We treat this tendency of the air to move in its original direction as the effect of centrifugal force. But now the wind is blowing toward higher pressure which cannot continue for any considerable distance as it is countered by the pressure-gradient force. The wind speed will decrease and coriolis and centrifugal forces also will decrease because of their dependence on speed. Pressure-gradient force is now stronger than the combination of the constantly decreasing coriolis and centrifugal forces, and the wind is deflected back toward lower pressure. This battle of forces continues until an equilibrium is reached where the pressure-gradient force is equal and opposite to the sum of coriolis force and centrifugal force. This is represented at Point "B" in the accompanying sketch, Figure 6–23.

Now we again have flow parallel to the contours. But in this equalizing process, what has happened to the wind speed? It has decreased. Therefore, we say that the gradient wind speed about a low is less than the geostrophic speed

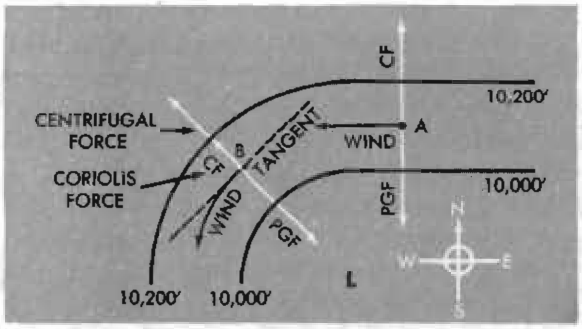


Figure 6–23. Gradient Wind About a Low

(for the same contour spacing and latitude).

Now consider the gradient wind about a high, Figure 6-24. At point "A" we have a geostrophic wind where the contours are

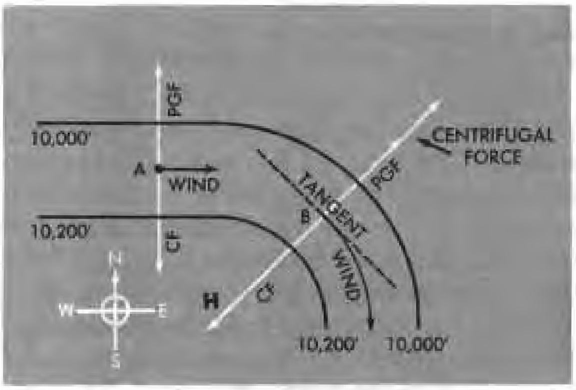


Figure 6–24. Gradient Wind About a High

straight and parallel. At point "B" we have the result of the balance of forces where coriolis force equals the sum of centrifugal and pressure-gradient forces, and the flow is again parallel to the contours. During the equalizing process the wind blew across the contours toward lower pressure and thereby increased in speed. This in turn increases the coriolis effect curving the wind toward the right in the diagram. We find that the gradient wind speed about a high is greater than the geostrophic speed (for same contour spacing and latitude).

The centrifugal force for unit mass is equal to the speed of the body squared, divided by the radius of curvature. The isobars or contours of most highs or anticyclones have a rather large radius of curvature and wide spacing, (slow wind speeds) so that the centrifugal force correction to the geostrophic wind is often unimportant. Moreover, notice that while coriolis force increases with the wind speed, centrifugal force increases with the speed squared. When the speed is large and the radius of curvature small, as in pronounced ridges, coriolis force cannot increase rapidly enough to compensate for centrifugal force, and the wind will not return to flow parallel to the contours.

Friction

Friction with the earth's surface slows the wind speed in the lowest layers of the atmosphere, thereby reducing the coriolis effect. The wind then blows across the isobars toward lower pressure. We have a spiraling inward toward lows and a spiraling outward from highs.

The angle that the wind direction makes with the isobars is usually about 10° over water and about 30° over land. The rougher the terrain, the greater the angle will be. It sometimes reaches 90° over rough land surfaces such as mountainous areas and results in situations where the wind direction shows little relationship to the isobars.

The wind flow may be considered frictionless above 1,000 to 2,000 feet over smooth water and above 2,000 to 3,000 feet over smooth land. Over mountainous areas the frictional layer may extend 6,000 feet or more above the range. Above these altitudes the gradient and geostrophic winds are good approximations to the actual wind.

Gustiness

The wind is usually not steady near the surface, especially over irregular terrain, where it moves as a succession of gusts and lulls of variable speed and direction. Eddy currents such as these are caused by friction between air and terrain, and are called *gusts* or *turbulence.* This type of turbulence is proportional to the wind speed and the roughness of the terrain.

Gustiness is also produced by locally unequal heating of the earth's surface. Gusts occur when the cooler air of adjacent areas rush-

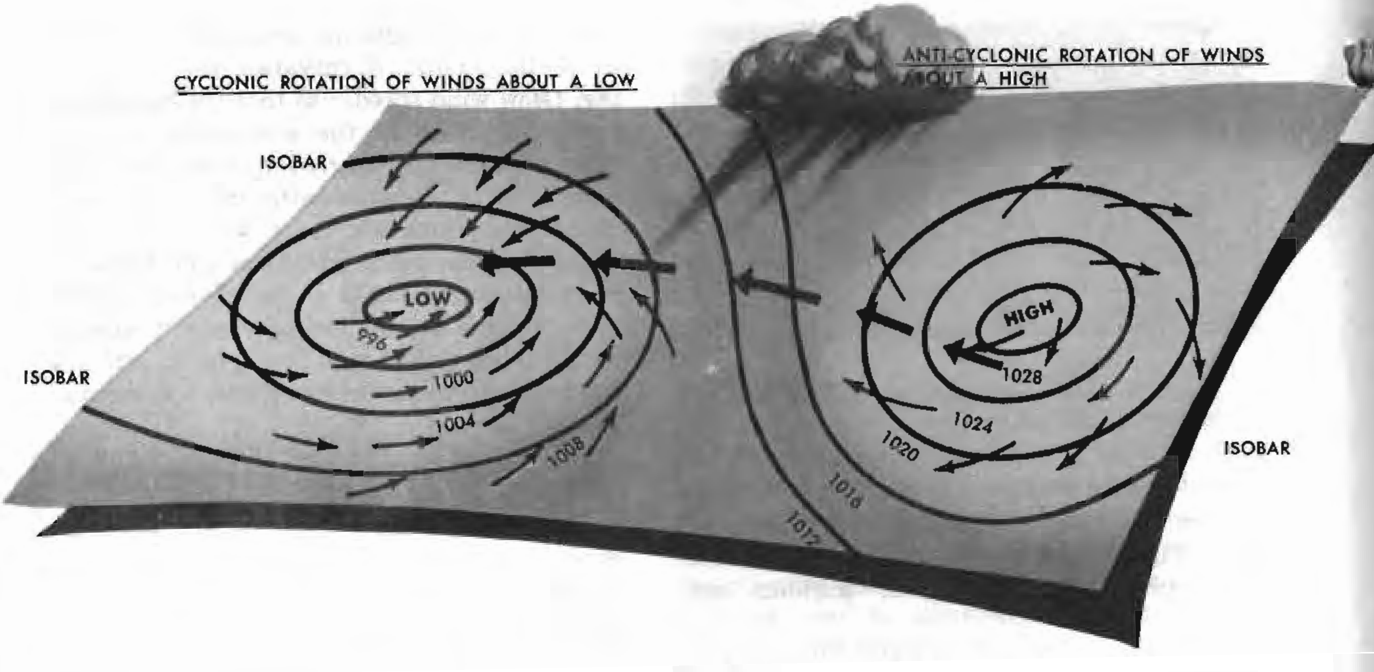


Figure 6–25. The Relation of Surface Wind to Pressure Patterns

es in to replace the rising warm air from heated areas (convection). Trees, lakes, hills, and buildings cause variations in surface friction and thereby produce gusts. Gustiness produced by surface friction is usually intensified on a sunny afternoon by this convection from heating at the earth's surface. Gustiness is also quite pronounced on occasions when the wind and pressure systems are such that cool air is advected over warmer terrain causing convection in much the same manner as solar heating.

GENERAL CIRCULATION

Any discussion of the subject of meteorology would be incomplete without mentioning the general circulation of the earth's atmosphere. Yet, the subject of general circulation is one of the most complex of all subjects falling under the heading of meteorology, because it involves such a vast area and so many variables. Several theories are found in the literature that attempt an explanation of the mean wind flow of the earth, but none of those so far offered have been universally accepted.

We do have some important facts that can be mentioned here, even though a discussion of all the theories that have been put forth is beyond the scope of this manual. We know that on a yearly average the earth receives considerably more energy from the sun in the tropical regions than in the polar regions. Yet, the tropical regions do not continue to get hotter and hotter, nor the polar regions colder. Therefore, something carries heat from the tropics toward the poles. We know, also, that the Northern Hemisphere surface winds follow a relatively constant pattern, with easterly winds generally blowing between the equator and about 30° latitude, westerly winds blowing between 30° and 60° latitude, and easterlies between 60° and 90° latitude. The winds aloft follow a somewhat similar pattern with the mid-latitude westerlies extending much farther north.

We observe that season after season, year after year, there tend to be definite belts of high and low pressure along given latitudes. We know that wind tends to blow from high pressure toward low pressure and is given direction by the coriolis effect. The winds, therefore, must be the result of the interaction of all the forces acting on the atmosphere. A complete list of these forces would have to include in addition to the pressure-gradient and

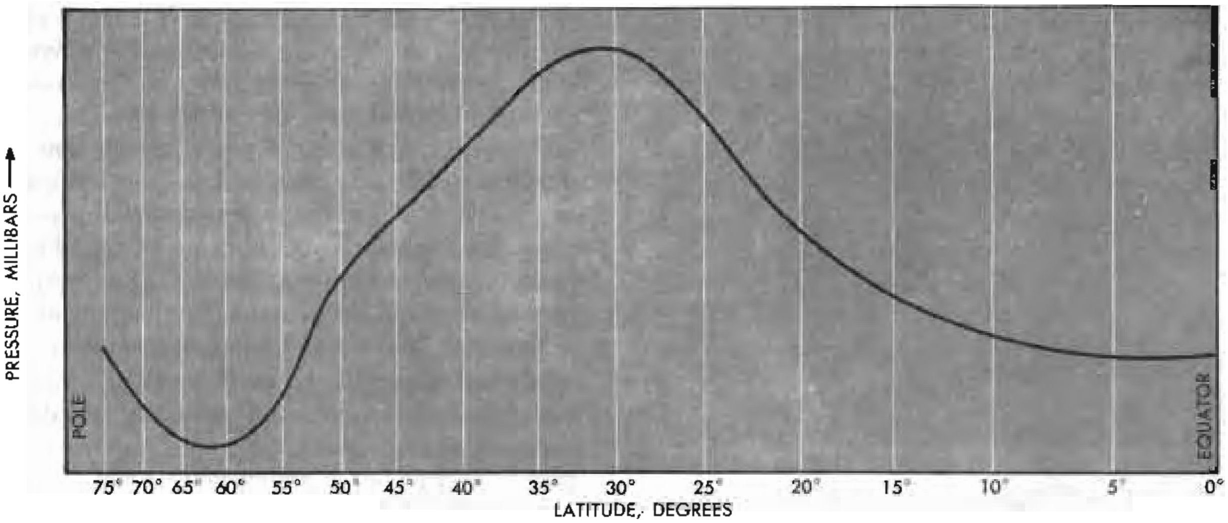


Figure 6–26. Mean Meridional Pressure Profile — Schematic, for Northern Hemisphere

coriolis forces, centrifugal force and the various effects that result from friction with the earth's surface, radiation, condensation, and evaporation. So any general circulation pattern that is discussed will not necessarily occur every day in every place, but will simply represent a statistical average over a year's or several years' period of time, and must be in agreement with the general pressure patterns.

The mean large-scale pressure and wind patterns have now been established by observations. Published maps of these patterns are available in many weather offices. Figure 6-26 shows a mean meridional pressure profile at

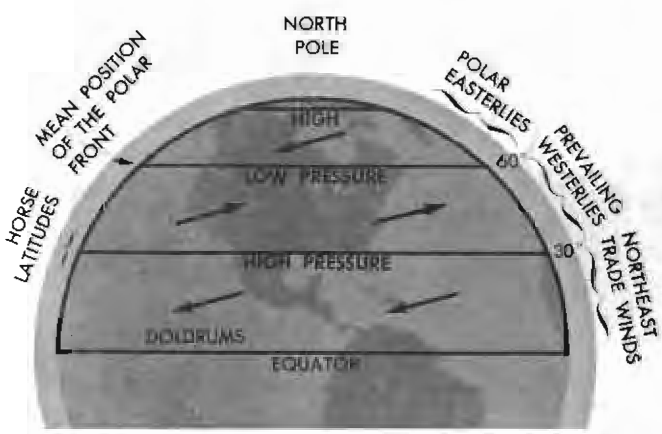


Figure 6–27. Diagram Showing General Surface Pressure Patterns and Resulting Wind Regimes

sea level in winter for the Northern Hemisphere and is based on observations, not theories. Notice that low pressure occurs at 60°N even though large high pressure areas occasionally occur at this latitude. Remember these data represent *the mean* value around all the hemisphere.

Translating the information of Figure 6-26 onto a view of the Northern Hemisphere and drawing arrows to represent the wind flow pattern at the surface, shows how the general circulation for the surface would appear for the Northern Hemisphere.

In general, the pressure pattern aloft at an altitude of about 15,000 feet shows the highest pressure at latitude 30° and lowest pressures in the polar regions, with the largest differential occurring in winter. Figure 6-28 displays this pattern in a hemispherical view with the resulting wind arrows drawn. Notice that a wide band of westerly winds occurs, and that there is no band of polar easterlies. Keep in mind that these are *mean conditions*. Certainly, easterly winds are occasionally found on the polar side of moving low pressure areas aloft in the arctic, particularly when the polar low aloft is displaced some distance from the pole, which often happens.

Thus, we have shown how the winds and pressure patterns must be in agreement, and the observed pressures and observed winds do agree. Why the patterns exist as they do has

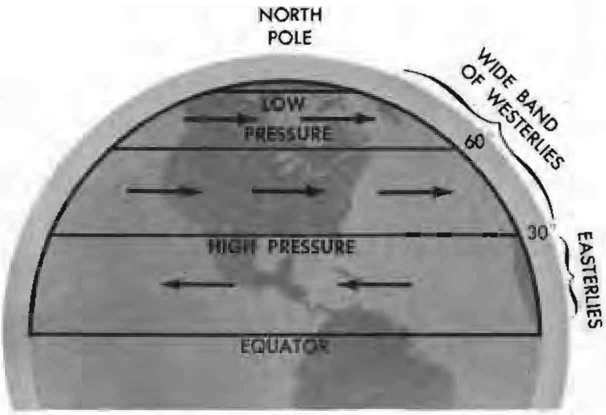


Figure 6–28. Diagram Showing General Pressure Pattern at 15,000 Feet and Resulting Wind Flow

not been mentioned, since it is in this area that complete agreement is not found. However, the literature does contain several very complete and comprehensive attempts at explaining the "whys" of the general circulation pattern.

Additionally, the mean wind flow gives a good clue to the normal direction of movement of migratory pressure systems. Moving highs and lows, even individual thunderstorms, tend to be "steered" in the direction of the major air flow in which they are imbedded.

Certainly, as more and more information is gathered in this satellite age, new frontiers and explanations of weather systems and the general circulation will be found. Meteorology stands to gain considerably in this space age, since high atmospheric data have been sparse in the past and the solutions to problems of radiative balance of the earth and the atmosphere must be solved before many weather problems can be solved.

Local Wind Effects

Local pressure and wind systems created by mountains, valleys, and water masses are superimposed on the general pressure and wind systems, and change the characteristics of the weather of the area. Some local wind effects of interest to aircrews are discussed in the paragraphs which follow.

LAND AND SEA BREEZES. Since land masses absorb and radiate heat more rapidly than water masses, the land is warmer than the sea

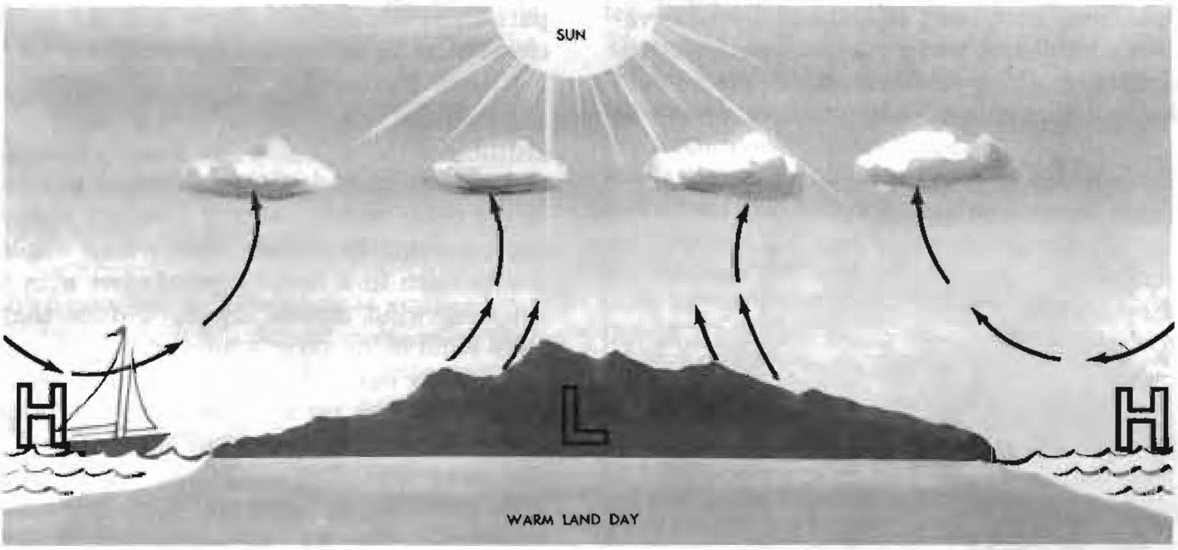


Figure 6–29. Sea Breeze, Onshore

Sea Breeze. Pressure over the warm land is lower than over the cooler water, as a result of daytime heating. This causes cool moist air to flow from over the water to over the land. (Onshore) during the day and colder at night. This difference in temperature is most noticeable during the summer months. In coastal areas, this variation of temperature produces a corresponding variation in pressure; during the day, the pressure over the warm land is lower than that over the colder water.

The local pressure and temperature distribution in coastal areas is such that the warm air over the land rises to a higher altitude and then moves horizontally out to sea. To replace this rising warm air at the surface, the colder air over the water moves onto the land (sea breeze). At night, the circulation is reversed so that the air movement at the surface is from land to sea (land breeze). The sea breezes are usually stronger than the land breezes, but they seldom penetrate far inland. Both land and sea breezes are shallow in depth.

The land and water temperature contrast so necessary for land and sea breezes is important for the large scale Asiatic Monsoon (seasonal wind). The related illustrations show the January and July mean surface pressure distributions. Notice the low over northern India in July compared to the higher pressure over the cooler Indian Ocean. This pressure distribution causes the onshore, moist, southwest wind of the summer Indian Monsoon. The orographic lifting of this air by the Himalaya Mountain Range gives Cherrapunji, Pakistan, an annual rainfall of 35½ *feet*, falling mostly in the summer months. (Contrast with 35 *inches* for many areas of our midwest).

During January the extremely cold temperatures over Siberia coupled with the warmer sea temperatures produce strong, dry, cold northwest winds along the Korean coast northward, and northerly winds over India. Observe the great cold surface high over the Asian continent in winter, Figure 6-32.

MOUNTAIN AND VALLEY WINDS. Sunlit mountain slopes and the air in contact with them are usually warmer than the free air at the same altitude (air farther from the slope) during the day, and they are colder at night. During the day, the air in contact with the slopes becomes lighter than the surrounding air and rises up the slopes. This air movement is called a *valley wind* because it seems to be flowing up out of the valley as shown in Figure 6-33.

At night, due to radiation, the air in contact with the mountain slopes becomes colder and denser than the surrounding air farther from the slopes and sinks along the slopes. This air movement is called the *mountain breeze* because it seems to flow down from the moun-

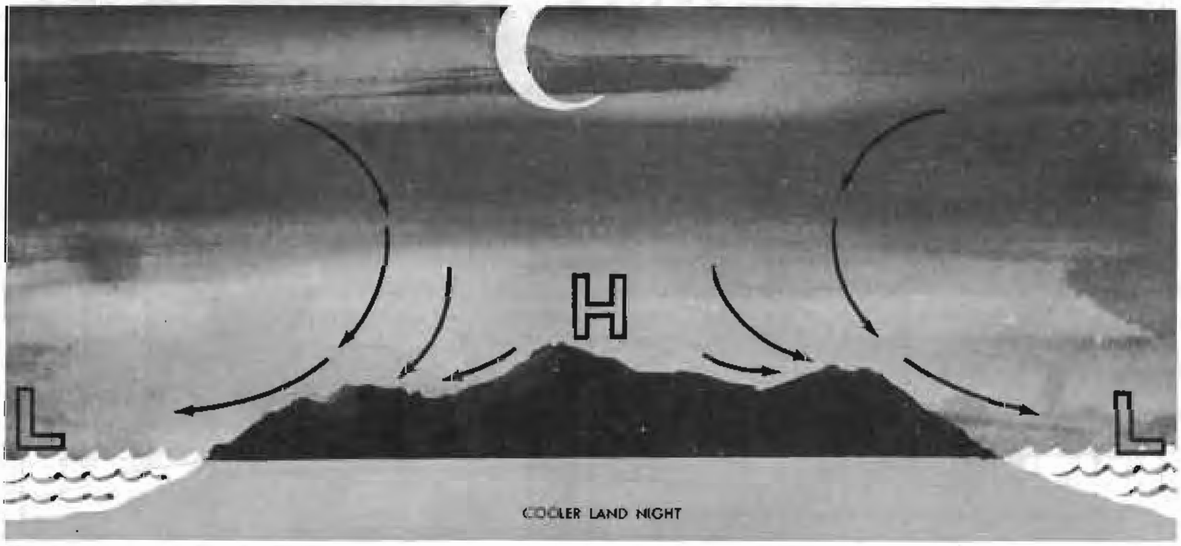


Figure 6–30. Land Breeze, Offshore

Land Breeze. Pressure over the cool land is higher than over the warmer water, as a result of strong nocturnal radiation. This causes air to flow from over the land to over the water. (Offshore) leeward slopes into the valley. The downslope motion causes compression of the air and resultant heating, which causes a dry, warm wind literally blowing down from the mountains. The line of clouds that lingers along the crest of the mountains is called a foehn wall (or chinook arch).

The downslope wind which is cold is called a *bora.* To qualify for the term *bora,* a wind must be so cold in its source that even after being heated by compression during the descent it arrives in the valley at a lower temperature than the air it is replacing. The best example of the bora is perhaps encountered in winter on the Adriatic Coast. The mountain breeze, already discussed, can also be considered a special kind or weak sort of a bora, although the term is not usually applied in this case.

Many areas of the world have local names for specific variations of the winds just discussed. A complete listing of such local winds is beyond the scope of this manual, but the reader is referred to appropriate textbooks which elaborate and explain these various winds.

Figure 6–35. Sketch Showing Pattern which Produces the Warm, Dry Downslope (Foehn or Chinook) Wind. Note the Clouds of the Mountain Peaks

Air Masses

The prevailing weather over a location at a given time generally depends on either the character of the prevailing air mass or the interaction of two or more air masses. This chapter is written specifically to acquaint the aircrewman with the characteristics of air masses, their formation, and the weather associated with them. The next chapter will deal with the interaction of two or more air masses.

An air mass may be defined as a large body of air with characteristics which are approximately uniform in a horizontal plane. An air mass extends over a large area, usually a thousand miles or more across. The basic properties of an air mass are described in terms of temperature and water vapor content. Except where the above-mentioned interactions are taking place, the weather is usually similar throughout an area covered by the same air mass. However, some modifications do occur because of the local effects of mountains, valleys, and large water masses.

FACTORS WHICH DETERMINE AIR MASS CHARACTERISTICS

The properties of an air mass are determined largely by the surface over which it forms. A body of air that has been nearly stationary or traveling for a long period of time over a region of the earth's surface which has relatively uniform moisture and temperature characteristics, develops correspondingly uniform characteristics.

The regions where characteristic types of air masses are formed are called *source regions.* The source region is the essential determining factor of the initial properties of the air mass. However, air masses that form over a given source region but at different seasons may develop different temperature and moisture characteristics because the characteristics of the source region may vary during the year. The depth to which an air mass takes on the properties of its source region depends upon the length of time it remains there and the stability of the air. Thus, we recognize shallow and deep air masses.

Other factors which determine the eventual characteristics of an air mass are: (1) the characteristics of the surface over which the air mass travels after leaving the source region, and (2) the amount of time that the air mass has been away from its source region.

SOURCE REGIONS

In order to fulfill the requirements for a good source region, an area must be of uniform surface (either all land or all water), of uniform temperature, and preferably a large area of high pressure where the air has a tendency to stagnate. Many regions of the earth do not fulfill these requirements. For example, most mid-latitude regions are either too variable with respect to winds and temperature, or they have an irregular distribution of land and sea surface. On the other hand, large snow or ice covered polar regions, tropical oceans, and large desert areas adequately fulfill source region requirements.

Classification of Air Masses

An air mass is classified first according to its source region, which may have been either polar or tropical, and these two types are subdivided next as to whether they are either continental or maritime.

Air masses which develop in stagnant high-pressure systems of the polar regions are characterized by the low temperatures which they acquire there. They are called cold air masses. Those which develop in the persistent high-pressure systems near 30° latitude are characterized by the high temperatures which they acquire there. They are called warm air masses.

Air masses which originate over large bodies of water usually have a relatively large amount of water vapor in their lower layers. They are called moist or maritime air masses. On the other hand, those which originate over land areas are usually relatively dry, and are called dry or continental air masses.

Accordingly, we may say that there are four basic types of air masses: *cold dry*, *cold moist*, *warm dry*, and *warm moist*. A pilot will hear the weather forecaster refer to them as *continental polar, maritime polar, continental tropical*, and *maritime tropical*. They mean the *same* thing.

In the Northern Hemisphere, Alaska, Canada, and Siberia are the principal winter source regions for cold, dry air masses; while the polar portions of the Atlantic and Pacific Oceans are the main source regions for cold, moist air masses.

The tropical regions of the Atlantic and Pacific Oceans are the main source regions for warm moist air masses, and arid regions of Africa, Asia and Australia are the principal source regions for warm, dry air masses. On occasion, the southwestern portion of the United States and the northern portion of

Figure 7–1. Trajectories of Air Masses Into North America

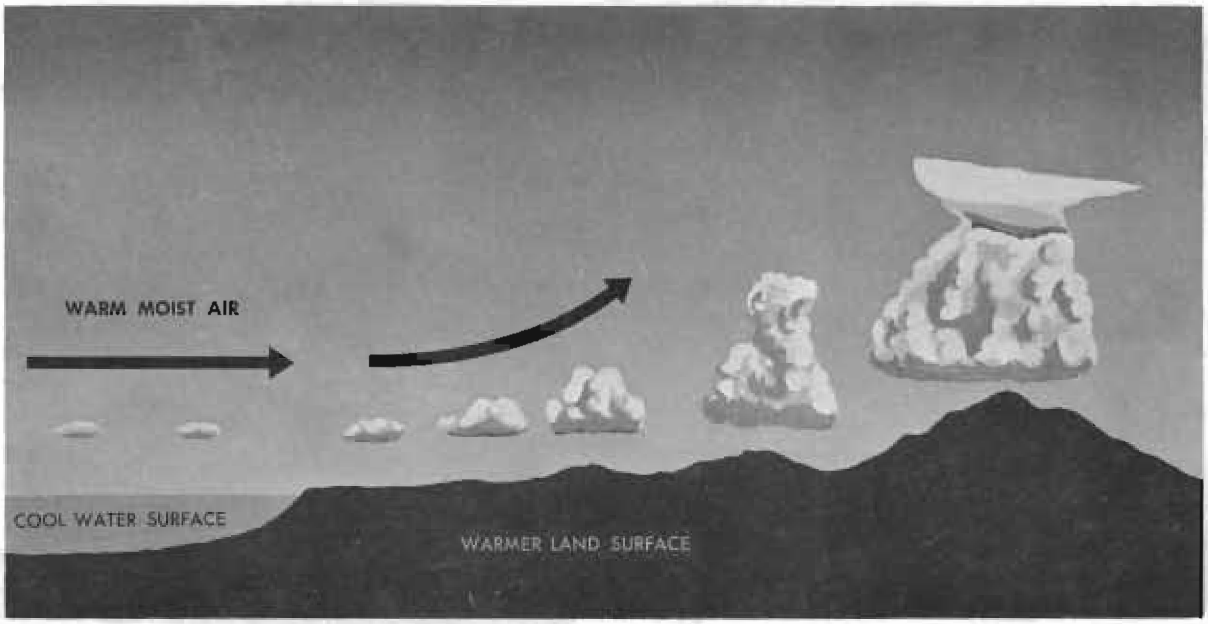


Figure 7–2. Warm, Moist Air Moving Over Warmer Surface

Mexico become a source region for warm, dry air masses in North America. The accompanying illustration shows the trajectories of air masses from their source regions into the North American continent.

In addition to the temperature and moisture content of an air mass, its stability is an important factor in its classification. This is particularly important when it is moving away from its source region, as we shall see in the next section. If an air mass is colder than the surface over which it is located, it becomes *unstable* (i.e., with convective up and down currents) in its lower levels because of the heating which it receives from the warmer surface; that is, it is heated from below. Conversely, when an air mass is warmer than the surface over which it is located, it becomes *stable* because of the cooling effect of the colder surface; that is, it is cooled from below. For further discussion on stability, refer to Chapter 4.

Modifications of an Air Mass After Leaving a Source Region

Just as an air mass takes on characteristics dependent on the underlying surface of the region where it forms, so it also tends to have its properties altered by contact with the underlying surface as it moves out of the source region. The degree of modification of an air mass is dependent on the speed at which it travels over the underlying surface, the nature of the surface, and the temperature contrast between the surface and the air mass.

When a cold, dry air mass moves slowly over a warm body of water, both the temperature and humidity of the air mass are increased; also, it becomes less stable. These changes occur in the lowest layers first, of course; when the lowest layers become sufficiently unstable, convective currents tend to spread modifications to progressively higher levels.

From a consideration of the modifying influences of the surface we can anticipate the flying conditions which will generally prevail over a wide area within an air mass. However, in individual localities, there may be special factors which further modify the flying conditions for that locality.

The accompanying sketches illustrate the effects which take place in an air mass due to surface thermal influences. In these cases, the heat content or temperature of the air mass has been changed by the contrasting temperature of the surface over which it is moving.

Figure 7–3. Warm, Moist Air Moving Over Colder Surface

An air mass over land is often additionally heated during the daytime when the surface is being heated by solar radiation. Similarly, the air temperature near the ground decreases at night when the surface is undergoing radiational cooling (if the sky is clear).

Figure 7–4. Heating of Land and Air by Insolation

Figure 7–5. Cooling of Land and Air by Radiation

An air mass can also be modified by mechanical influences on its motion, which produce lifting, sinking, and mixing actions; these alter the vertical distribution of temperature and water vapor within the air mass. Turbulent mixing occurs when an air mass moves over rough terrain. Sinking motions on the lee sides of mountains increase the temperature and decrease the relative humidity of the air mass. On the other hand, the air mass temperature and relative humidity are altered by passage over a mountain range if clouds and precipitation occur on the windward side. This is shown in the accompanying illustration.

Figure 7–6. Orographic Lifting Resulting in Cloud Formation, Drying of the Air and Warming of the Air on Descent Downslope

Because of evaporation, the water vapor content of an air mass increases as the air mass moves over a water surface or moist ground, providing the temperature of the water or ground is higher than the dew point of the air mass. Rain falling through an air mass usually raises its humidity.

TYPES AND ACTIVITIES OF AIR MASSES

Thus far, we have been concerned mainly in getting acquainted with how air masses come into being, the different types, and what modifications they undergo as they move over the earth's surface. Now, we are ready to discuss the weather within the different types of air masses. But first, let's briefly review some of the atmospheric activities covered in this and previous chapters. This review will help make the discussion of air mass weather easier to understand.

We recall that:

- The advection of cold air over a warmer surface gradually increases the temperature of the air next to the ground and makes it unstable.
- The advection of warm air over a colder surface decreases the temperature of the air next to the ground and makes it stable.
- Ascending air currents (convection) are produced by the heat the air receives while in contact with a hot surface (conduction), by the air being forced over mountains (orographic lifting), or by warm air being forced over colder air (frontal lifting).
- The relative humidity of air increases as the air expands and cools in ascending currents. When the lifting and resulting expansion and cooling are extensive enough to produce saturation, clouds and turbulence are the usual result.
- The relative humidity of air decreases as the air contracts and warms in descending air currents. For this reason, clouds generally dissipate in descending air currents.
- Turbulence, vertical air currents, cumuliform clouds, showery precipitation, and good surface visibility (except in showers or dust) are normally associated with unstable air.
- Smooth flying weather, stratiform clouds, and fair to poor surface visibilities are normally associated with stable air.
- Air masses acquire water vapor by evaporation from underlying water masses or from precipitation falling through the air mass, (providing the water temperature exceeds the dew point).

• The water vapor content of an air mass is reduced by the formation of clouds and precipitation on the windward side of mountains over which air is flowing (orographic lifting). As the air descends the leeward side, it is heated and consequently dried.

• Mountains, valleys, and water masses modify the temperature and humidity of the air mass over a given locality.

• There is a daily cyclical variation of temperature in the surface layers of an air mass. Minimum temperatures normally occur near daybreak. The temperature then steadily rises and reaches a maximum value between 1400 and 1600 local time. A steady decrease in temperature (nocturnal cooling) then takes place during the night and early morning.

Cold, Moist Air Masses

The cold, moist air masses which invade the United States arrive from two different source regions — the North Pacific Ocean and the northwestern portion of the North Atlantic Ocean. Cold, moist air masses originating over the North Atlantic appear rather frequently over the northeastern coast of the United States. Those originating over the Pacific Ocean dominate the weather along the Pacific Coast of the United States and Canada.

WINTER. The winter weather and flying conditions associated with the cold, moist air invading the Pacific Coast vary greatly because of the different trajectories over the ocean which these air masses can have. Those which come from the northwest, that is, from the Aleutian Islands and across the Gulf of Alaska, are unstable in the lower layers. Originally very cold, they have gained heat and moisture during the comparatively short trip across the warmer waters of the gulf. The accompanying illustration shows the typical result. As the air is lifted over the coastal range of mountains, cumuliform clouds are formed. They extend to great heights, and their bases will obscure the tops of the mountains.

The cold, moist air masses which approach

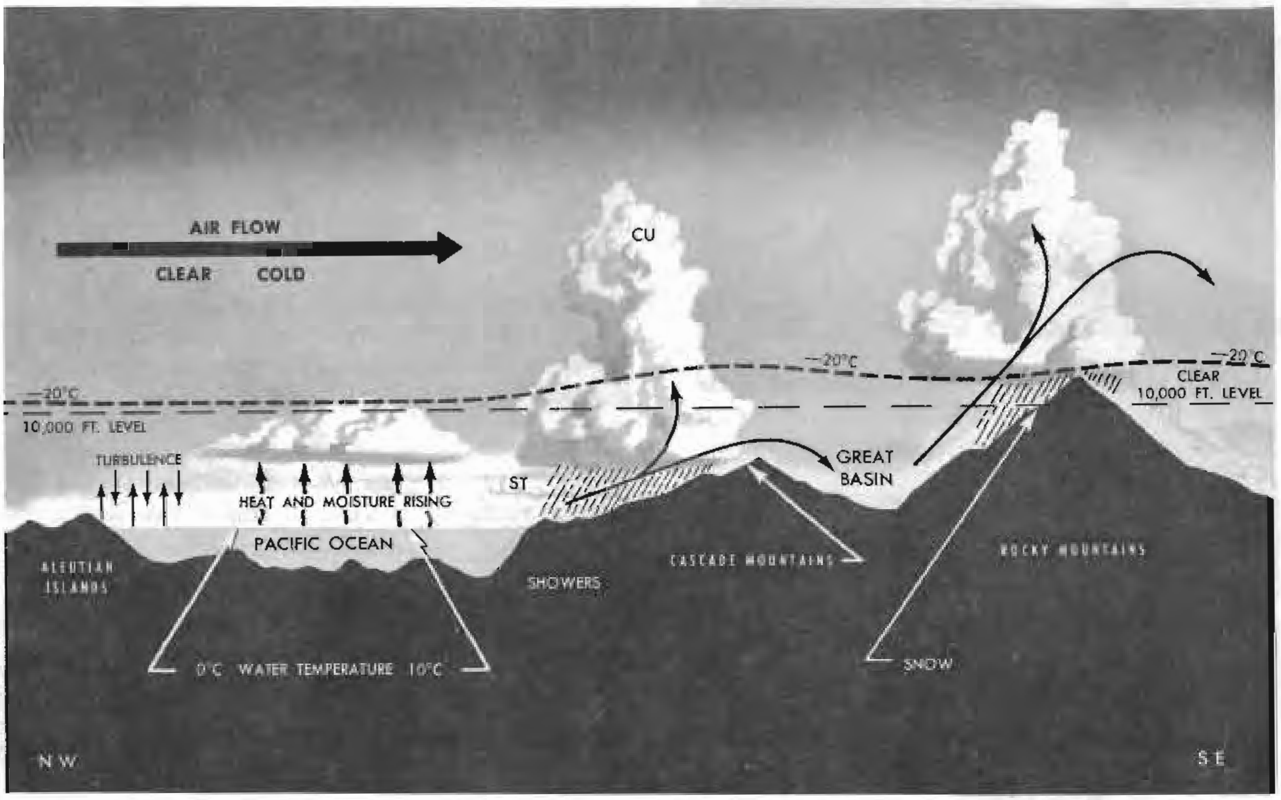


Figure 7–7. Winter Movement of Cold, Moist Air Mass from Gulf of Alaska into Northwest United States

the Pacific Coast more directly from the west probably originated in Siberia, although in some cases they have come from Alaska. Whichever it was, they have had a long overwater trajectory and have picked up considerable heat and moisture. There has been time to distribute it throughout the lower layers, so that the air is no longer strongly unstable when it reaches the coast. Stratus and stratocumulus clouds are common throughout the Pacific Coast.

Whichever direction they come from, the cold, moist air masses cause extensive precipitation as they move eastward up the western slopes of the mountains. East of the mountains, skies are generally clear, and the air is warm and dry as it comes down the eastern slopes (foehn-like effect).

It is this air mass that causes the Pacific Northwest, including British Columbia, to have more precipitation than any other region in North America (especially extreme western Washington in the Olympic Mountains).

Across the country, in the northeastern section of the United States, cold, moist air masses move in from a northeasterly direction. New Englanders use the term "Nor'easter" to describe the weather associated with a strong flow of this air. These air masses are usually colder and more stable than those approaching the west coast from a northwesterly direction; instability is confined to a shallow layer near the surface. Low stratiform clouds form as the air masses move inland.

Summer. Although the water temperatures are generally lower than the land temperature during the summer, the cold, moist air masses are cooled by the cold water belts found along some coast lines. Consequently, they approach these coasts in a more stable condition than they otherwise would have. Fog and low stratus are common along these coasts, some of which are the coasts of California, Peru, Chile, Morroco, southwest Africa, east Canada, etc. Inland, however, lifting along the western slopes of mountains tends to produce unstable conditions, and cumuliform clouds develop.

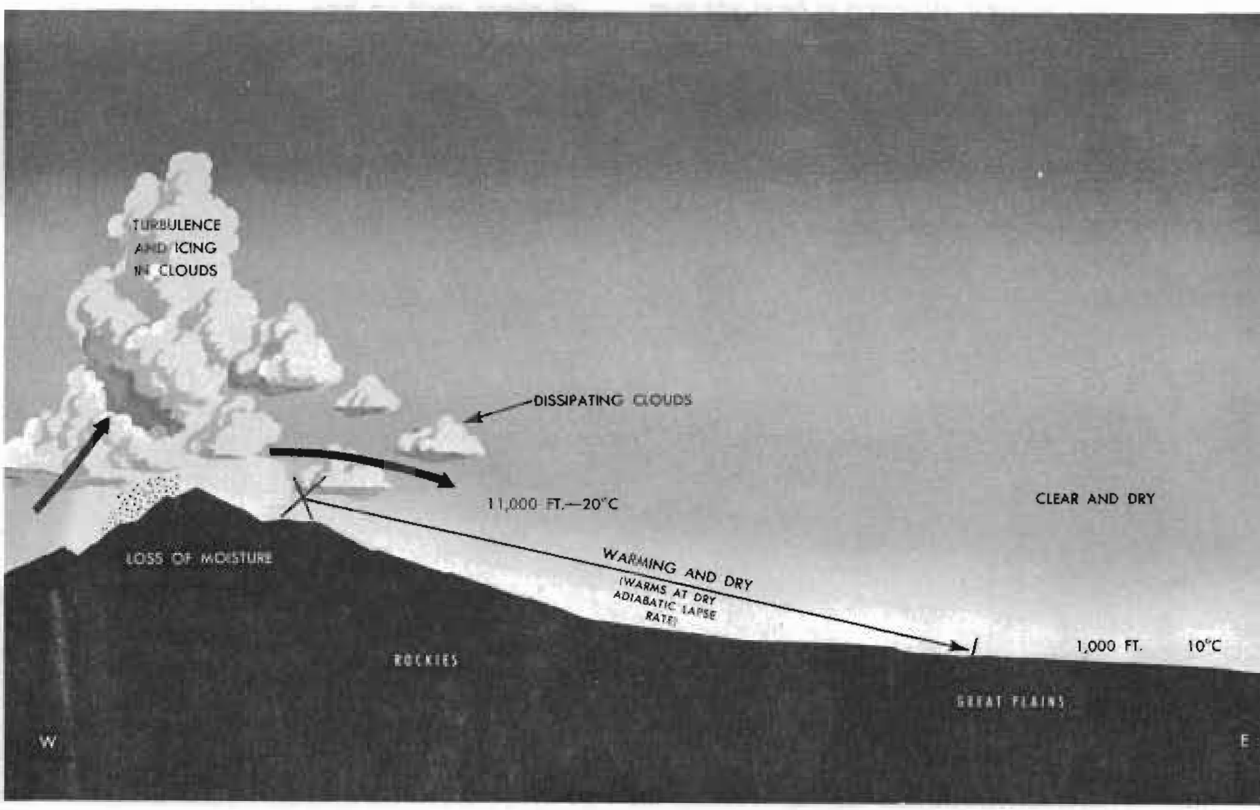


Figure 7–8. Cold, Moist Air Crossing the Rockies

These are intensified by daytime surface heating. After this air mass has crossed the mountains, it becomes so heated that the relative humidity falls to low values, and clear skies and dry weather prevail.

The summer weather and flying conditions associated with cold, moist air masses over the northeastern United States are similar to those of winter: stratiform clouds or fog.

Cold, Dry Air Masses

The cold, dry air masses which invade the United States originate over northern Canada and Alaska; during the winter they also form over the frozen Arctic Ocean.

WINTER. In winter these cold, dry air masses are stable in their source regions. As they move southward toward the United States, they are heated by the underlying surface. During the daylight hours the air generally becomes unstable, especially if the sky is clear and there is no snow on the ground. At night the air tends to become stable again.

At times, when these cold, dry air masses move over the warmer waters of the Great Lakes, they acquire a great deal of water vapor and become more unstable. When this happens, cumuliform clouds develop and produce snow flurries to the lee of the lakes, as shown in the accompanying illustration. In some cases very heavy snowfalls occur close to the lake shores. As the modified air mass moves southeastward, cumuliform clouds build up over the Appalachian Mountains. The skies are usually clear east of the mountains.

SUMMER. Cold, dry air masses have different characteristics in the summer than they do in the winter. Since the thawed land source regions are now warm from solar heating, the air is less stable in the lower layers. The air is cool and dry when it reaches the United States. Cumuliform clouds form during the day when the heating from the sun causes it to become slightly unstable; but the air becomes stable again at night. When these air masses move over the Great Lakes in summer, they are cooled from below, which in-

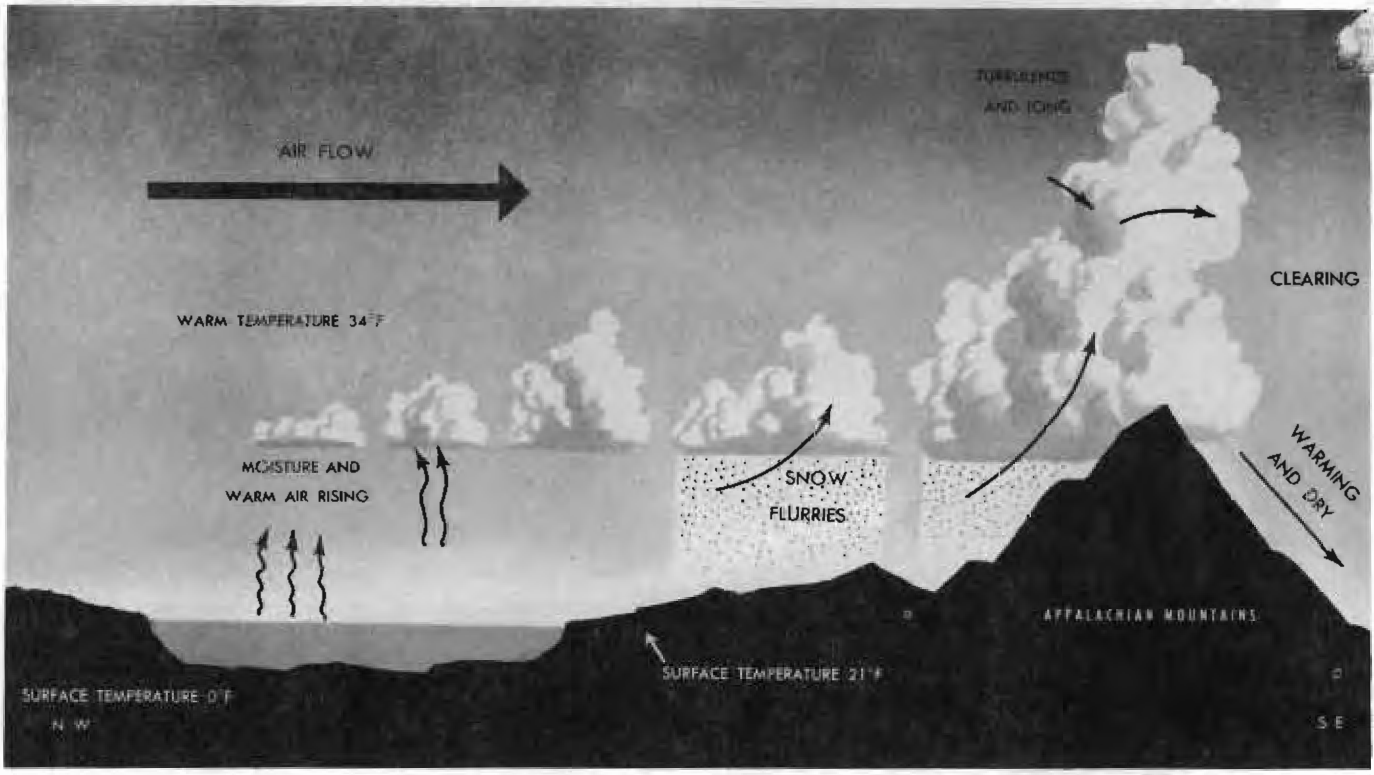


Figure 7–9. Cold, Dry Air Crossing the Great Lakes in Winter

creases their stability, causing fog. Stratiform clouds are then common over the lakes and to the lee of the lakes. Over the central and southern United States this air is rapidly heated and moistened.

Warm, Moist Air Masses

The source regions of warm, moist air invading middle latitudes, are the vast oceanic areas around 15° to 30° latitude. Semipermanent high-pressure systems are located over these regions in all seasons of the year. The one having the greatest influence on the United States is known as the Bermuda High. Depending upon the strength and location of the Bermuda High, warm, moist air masses originating over the Atlantic Ocean move in over the United States along the southeast coast or through the Gulf of Mexico along the Gulf Coast. Warm, moist air which originates over the Pacific Ocean is observed only occasionally along the coast of California.

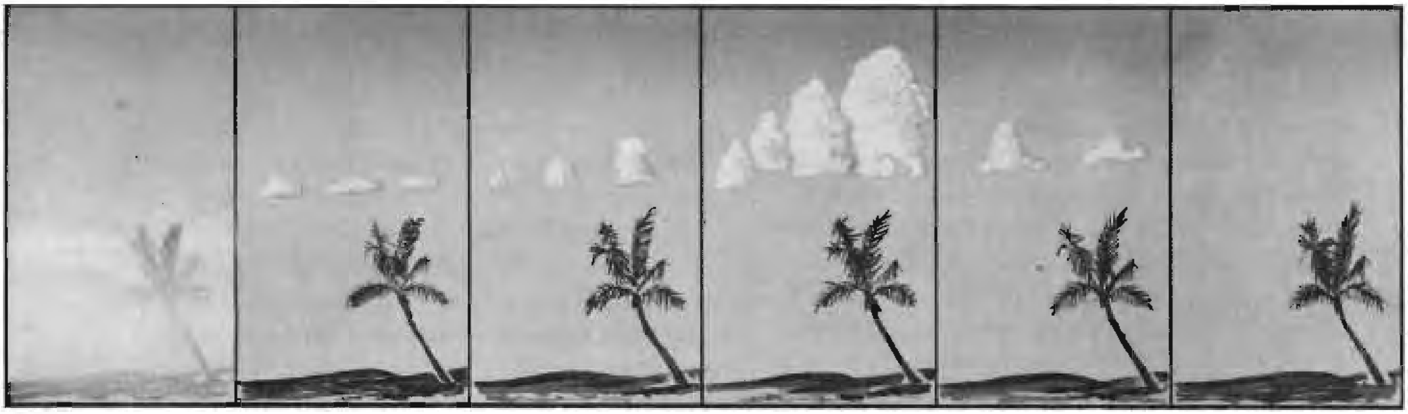
WINTER. Again, because the land is colder than the water, the warm, moist air masses are cooled from below, and as they move inland over the southeastern and Gulf Coast states during the winter, they tend to become stable in the lowest layers. As a whole, however, the warm, moist air mass is very nearly unstable, and in the coastal areas displays a characteristic diurnal variation as illustrated in Figure 7-10. Fog or low stratiform clouds form at night; they tend to dissipate during the late morning; shortly thereafter, patch cumulus appear, increasing in amount during the afternoon, and disappearing after sundown.

The extent to which the cloudiness and fog spread inland depends on the differences between the surface and air temperatures and strength of onshore winds. When the surface temperatures are cold, the fog and stratiform clouds extend inland for considerable distances if the winds are favorable. On occasion, when the land temperatures are extremely cold, extensive surface temperature inversions develop. Under such conditions, daytime heating usually does not eliminate the surface temperature inversion, and the fog and stratus may persist for several days.

When the warm, moist air moves far enough over the land surface that it is lifted up the slope of extensive mountain ranges such as the Appalachians, it becomes unstable. The weather and flying conditions are then characterized by heavy cumuliform and stratiform clouds which obscure the mountain tops.

SUMMER. Warm, moist air covers the eastern half of the United States during the greater portion of the summer. Since in the summer the land is normally warmer than the water, particularly during the day, this air mass is heated from below by the surface and becomes unstable as it moves inland.

Along the coastal regions, stratiform clouds are common during the early morning hours. These clouds usually change during the late morning to scattered cumuliform clouds, which continue to grow in size and number during the afternoon. Frequently, these cumuliform clouds will develop into extensive thunder-



0600 LST 1000 L 1200 L 1500 L 1900 L 2100 L

Figure 7–10. Diurnal Variation in Warm, Moist Air Along the Gulf Coast (Winter) storms by late afternoon. These thunderstorm clouds, however, are generally widely scattered. In mountainous areas, cumuliform clouds and thunderstorms are usually more numerous and intense on the windward side of mountain ranges than on the lee side.

Warm, Dry Air Masses

Warm, dry air masses are observed over the western United States primarily in summer, and mainly in their source region which is the Mexico-Great Basin area, from where they sometimes spread farther east. This air mass is characterized by high temperatures, low humidities, and sparse cloudiness which, when present, is of the cumuliform type and mainly over the mountains. The bases of these cumuliform clouds are exceptionally high for this type of cloud. For example, they may be found as high as 10,000 or 11,000 feet, instead of the more usual heights of 4,000 to 5,000 feet.

The formation of these "low clouds" at such high levels is a graphic indication of the great vertical extent of the effects of surface heating. Flying is often rough at middle and low levels, especially during daylight hours. Occasional dust storms present another significant hazard to flying, since the dust or sand may extend to rather high altitudes and reduce visibilities for extended periods.

Summary

This chapter has described the primary characteristics of the various air masses which influence the weather and flying conditions in the United States. As has been suggested, the aircrew member will encounter many variations of these generalized conditions. To determine the specific conditions for flight within any air mass on a specific day, consult the current weather reports and charts, and discuss the matter with the duty forecaster at the time of the weather briefing. Guidance for this is presented in Chapter 18.

CHAPTER 8

Fronts

Where two or more air masses with different properties and characteristics come into contact, the boundary between them is not a sharp wall. Instead, there is generally a zone of transition (sometimes referred to as a frontal zone or zone of discontinuity) which is often many miles thick. For simplicity, however, it is customarily called a frontal surface or, merely a front. The geographic location of a front is indicated on a weather map by a line. Since air masses have depth as well as horizontal extent, it follows that the zone of transition extends upward from the surface. We learned in the preceding chapter that the characteristics of air masses originate at the surface, and work gradually upward.

It is apparent that the contrast between different air masses will be most significant in the lower layers. At some level above the surface, which varies depending on whether the air masses are shallow or deep, the differences between them become small, and the concept of a frontal surface becomes meaningless. As their associated weather is usually confined to the lower troposphere, fronts are seldom recognizable above 15,000 or 20,000 feet, although the temperature change may extend to the tropopause in some cases.

When flying through a front (or when a front moves past a weather observing station), the change from the properties and characteristics of one air mass to those of the other is sometimes quite abrupt, (narrow transition zone); at other times, it is very gradual (diffuse transition zone). Fronts tend to be sharpest where there is a tendency for the cold and warm air to blow toward one another (converge).

The polar regions are dominated by cold air masses. The tropics are dominated by warm air masses. Middle latitudes are regions where cold and warm air masses continually interact with each other — the cold air moving southward and the warm air moving northward, in alternating tongues or waves. The zone which separates these air masses is called the polar front.

The polar front is not stationary. At places a strong flow of cold air pushes southward and replaces the tropical air. At other places it retreats ahead of the advancing warm air. In general we find that while the polar front is advancing southward in one region it is retreating northward in an adjacent region, giving it a wave-like shape, as shown in the accompanying illustration. The polar front, which is observed most frequently in the temperate zone (mid-latitudes), may occasionally move well into the tropics in winter when the cold air masses are dominant. In summer, when the warm air masses are dominant, the polar front may move as far north as 60° N.

Although the polar front is the main zone of discontinuity in each hemisphere, fronts may form between any air masses if the air masses are sufficiently dissimilar.

Transition zones (fronts) which are confined mainly to the tropics and polar regions, or rarely observed in the temperate zone, are discussed in Chapters 13 and 14.

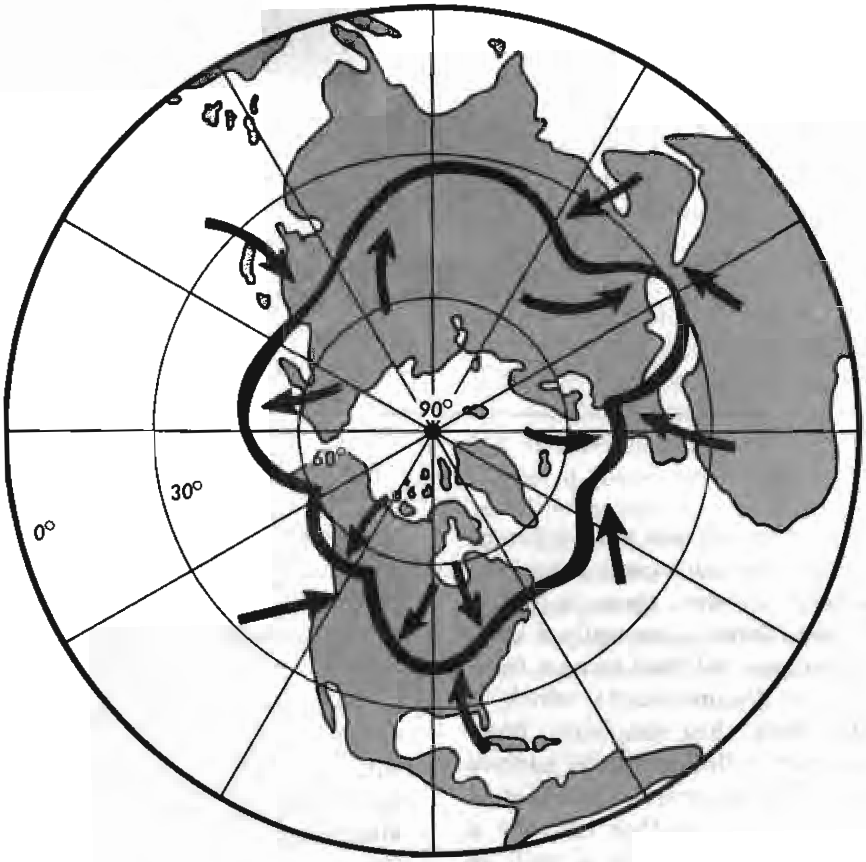


Figure 8–1. Hemispheric View of the Polar Front

A hypothetical example, showing the semipermanent zone of discontinuity between regions of polar air masses and regions of tropical air masses.

THE FRONTAL SLOPE

Since the air masses separated by a front have different temperature and water vapor characteristics, they also have different densities. When air masses with significantly different densities are adjoining, there is a tendency for the denser air to slide or wedge under the less dense air.

Conversely, we find that the less dense warmer air slides up over the denser cooler air. As a result, the frontal surface is sloping. There are several factors which determine the steepness of the slope, which is measured as the angle between the earth's surface and the under side of the frontal surface.

DISCONTINUITIES ACROSS FRONTS

Discontinuities in air mass properties and characteristics, such as temperature, water vapor content, wind, cloud types, and pressure changes are used by forecasters to locate and identify fronts, and to trace their movement. Most of these are shown in various illustrations throughout this chapter. In this section, we shall discuss briefly the discontinuities in temperature and winds which are noticeable to aircrews flying through fronts. Frontal clouds and weather hazards are discussed later in the chapter.

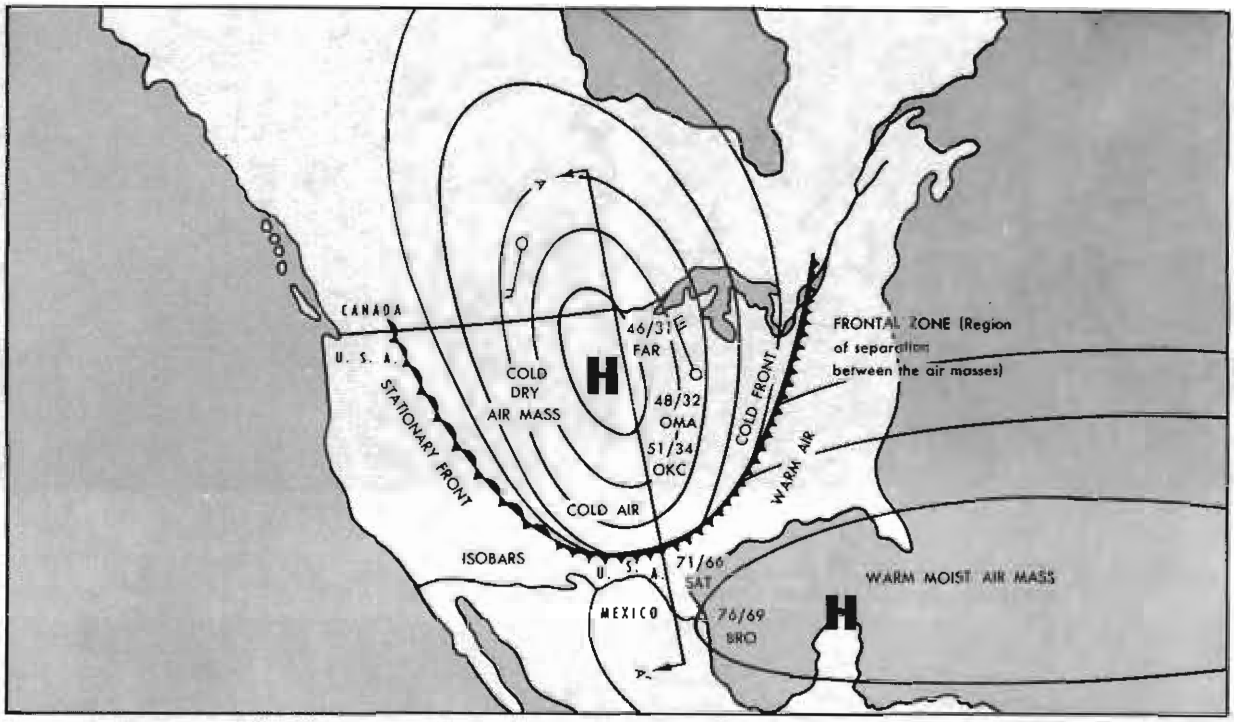


Figure 8–2. The Polar Front in North America

Temperature

One of the most easily recognized discontinuities across a front is temperature. At the earth's surface, the passage of a front is usually characterized by a noticeable change in temperature. The rate and amount of the change

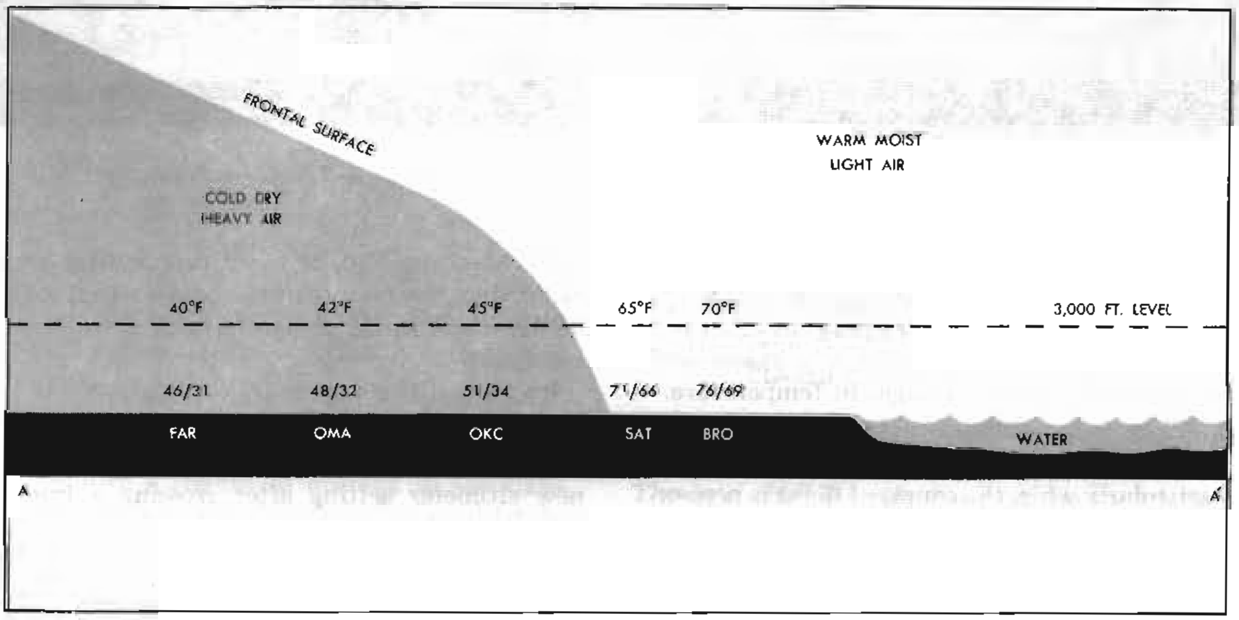


Figure 8–3. Vertical Cross Section Across the Polar Front, Showing Typical Temperature/Dew Point Distribution

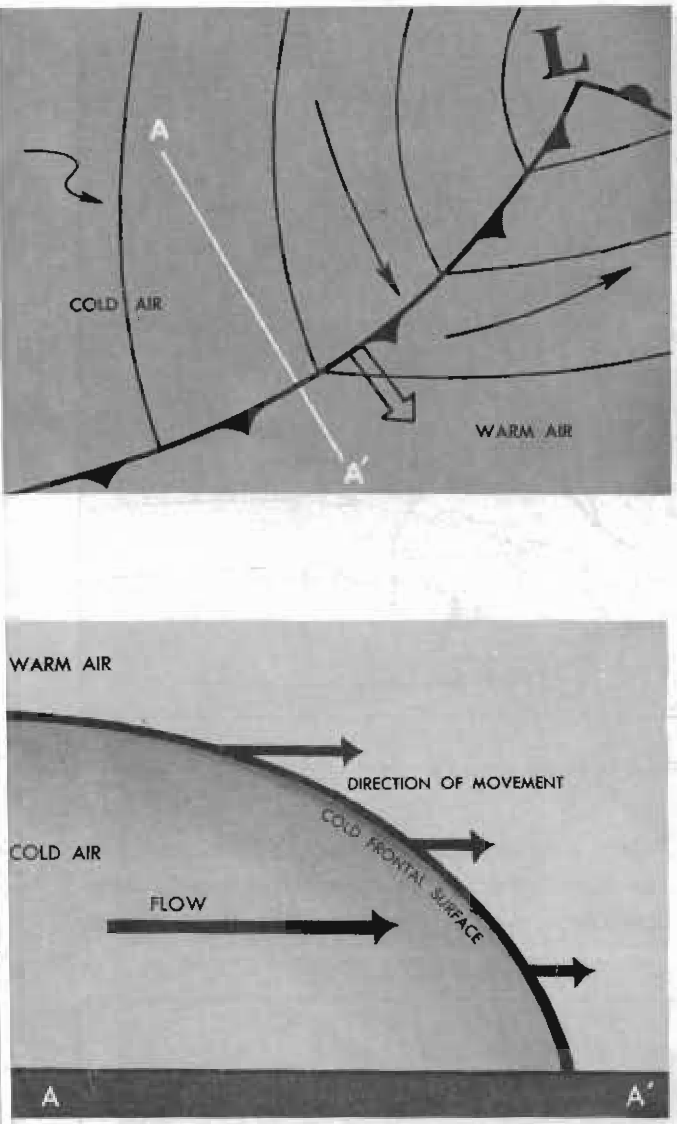


Figure 8–4. Cold Front

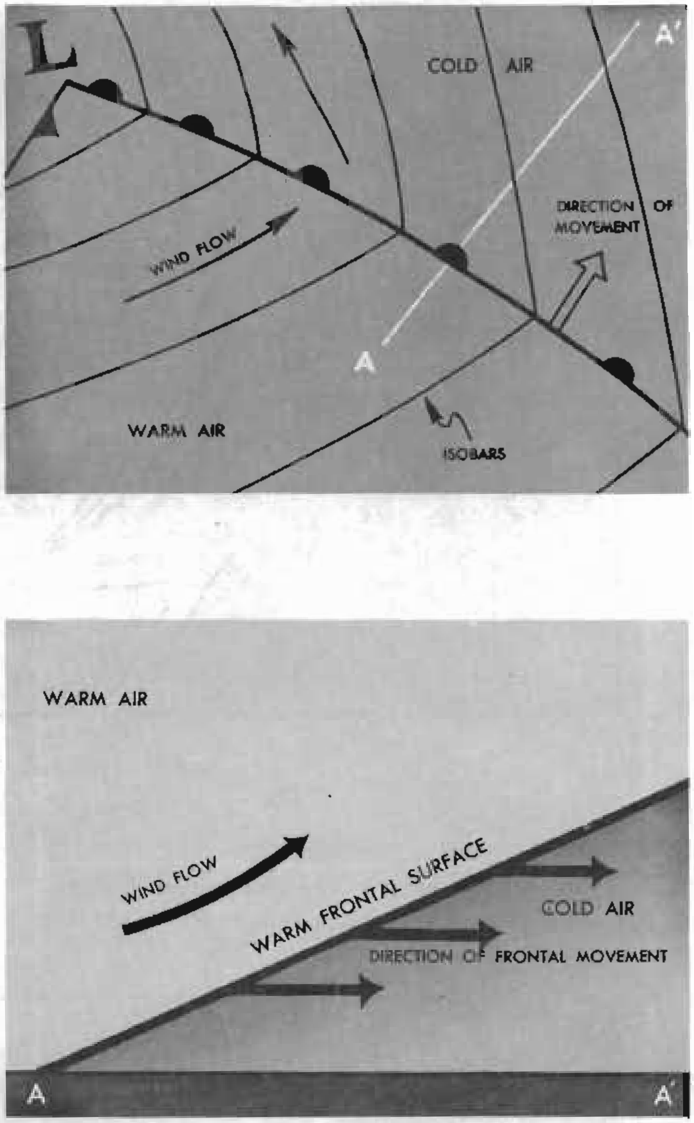


Figure 8–5. Warm Front

is a partial indication of the intensity of the front. Strong or sharp fronts are accompanied by abrupt and sizeable temperature changes, while weak or diffuse fronts are characterized by gradual or minor changes in temperature.

When flying through a front, you will observe a significant change in temperature, particularly when the course of flight is perpendicular to the front. The change in temperature occurs within a short period of time and/or in a short distance (on the order of 1 to 20 miles), and is usually more pronounced at lower altitudes than at higher altitudes. The point to remember is that this change, even when gradual, is more pronounced and rapid than the temperature change which may be observed during a flight wholly within one air mass.

The change in temperature is an indication of a change in air mass density across the front. Therefore, it is advisable to obtain a new altimeter setting after crossing a front. If you are flying at a flight level for which a standard setting of 29.92" is used, this advice does not apply. However, at such an altitude, the temperature change is not likely to be noticeable anyway. (For more on altimetry see Chapter 6.)